

Imperial College London
Department of Physics

The Role of Atmospheric Stability in the Forcing-Feedbacks-Adjustments Framework

Pietro Salvi

Submitted in part fulfilment of the requirements
for the degree of Doctor of Philosophy in Physics
of Imperial College London, February 2023

Statement of originality

I, Pietro Salvi, confirm that this work is my own. Other works used here have been appropriately referenced to the best of my ability.

Copyright declaration

The copyright of this thesis rests with the author. Unless otherwise indicated, its contents are licensed under a Creative Commons Attribution-Non Commercial 4.0 International Licence (CC BY-NC). Under this licence, you may copy and redistribute the material in any medium or format. You may also create and distribute modified versions of the work. This is on the condition that: you credit the author and do not use it, or any derivative works, for a commercial purpose. When reusing or sharing this work, ensure you make the licence terms clear to others by naming the licence and linking to the licence text. Where a work has been adapted, you should indicate that the work has been changed and describe those changes. Please seek permission from the copyright holder for uses of this work that are not included in this licence or permitted under UK Copyright Law.

Abstract

One of the most recognised measures of climate change, global warming, is still subject to significant uncertainty. Even with the best tools available in the form of climate models, wide-ranging estimates are given of the increase in surface temperature for a given emissions scenario. This thesis investigates various mechanisms by which temperature change occurs as driven by energy imbalances introduced into the Earth system.

The focus of this work lies on atmospheric stability, a measure of vertical temperature gradients that is well related to convection and thus in turn clouds. Greater stability suppresses convection and encourages low cloud formation which provides a cooling effect on the planet as these low clouds reflect incoming solar radiation. The structure of energy imbalances induced by forcing agents is found to be significant for stability in many ways, from the vertical structure of atmospheric heating, to the horizontal distribution of radiative flux that affects the pattern of sea surface temperature (SST) change, to the time-evolution of the forcing agents themselves. The result of analysing the impact of these structures through stability is that stability is found to explain much in terms of differences between forcing agents, as well as how temperatures evolve in the historical period.

This thesis incorporates analysis of publicly available data from Climate Model Inter-comparison Project 6 (CMIP6) historical experiments in order to compare and contrast feedbacks from time-evolving aerosol, greenhouse gas (GHG), and natural forcings. Work here shows that historical aerosols cause more amplifying feedbacks (higher effective climate sensitivity) than GHGs. Data from idealised computational experiments run for this project demonstrate that this is due to the extra-tropical skew of aerosol forcing. Furthermore, the time-evolving feedbacks of the historical period are found to be explained by the time-evolving feedbacks for the components of historical forcing.

Acknowledgements

My thanks and appreciation go out first to my supervisors Paulo Ceppi and Jonathan Gregory. They have managed to be supportive whilst keeping me challenged by and focused on the research and work that has led to the writing of this thesis.

Special thanks also go to Helen Brindley and Jean-Louis Dufresne, who acted as helpful and engaging examiners for this thesis.

Vegans, socialists, devout Christians, Europeans, Americans, proof-readers, secretaries, physicists, biochemists, computer scientists, materials scientists, philosophy, tea, coffee, the LGBT+ community, Dungeons & Dragons, family, accountants, retail workers, technical forum posts, the Grantham Institute of Imperial College London, and Python: some combination of these is what has led to this thesis, but without a control experiment it is hard to know which to thank in particular. As such, I dedicate this to all the above.

Sometimes a scream is better than a thesis

Ralph Waldo Emerson

Glossary

AA: Anthropogenic Aerosol

AGCM: Atmosphere General Circulation Model

AMIP: Atmospheric Model Intercomparison Project

AOGCM: Atmosphere-Ocean General Circulation Model

BC: Black Carbon

CMIP: Climate Model Intercomparison Project

CRE: Cloud Radiative Effect

DAMIP: Detection and Attribution Model Intercomparison Project

ECS: Equilibrium climate sensitivity

EIS: Estimated Inversion Strength

ERF: Effective Radiative Forcing

GHG: GreenHouse Gas

LW: LongWave

NH: Northern Hemisphere

RFMIP: Radiative Forcing Model Intercomparison Project

SH: Southern Hemisphere

SST: Sea Surface Temperature

SW: ShortWave

TOA: Top Of Atmosphere

CONTENTS

Statement of originality	2
Abstract	3
Acknowledgements	4
Glossary	5
Contents	6
List of Tables	8
List of Figures	9
1 Introduction	15
1.1 Introduction	15
1.2 Objectives	16
2 A review of the forcing-feedbacks-adjustments framework	18
2.1 Introduction	18
2.2 The global energy balance	19
2.3 The forcing–adjustment–feedback framework	22
2.4 Calculating forcing, adjustments, and feedbacks	25
2.5 The forcing agent dependence of radiative adjustments	27
2.6 The agent dependence of radiative feedbacks	29
2.7 The role of stability in radiative adjustments and feedbacks	31
2.8 Discussion and conclusions	33

3	The role of stability in rapid cloud radiative adjustments¹	35
3.1	Introduction	35
3.2	Data and methods	37
3.3	Vertically localised atmospheric heating experiments	42
3.4	Interpreting cloud-radiative adjustments to common forcing agents	48
3.5	Summary and conclusions	53
4	The dependence of radiative feedbacks on forcing agent²	55
4.1	Introduction	56
4.2	Data and methods	57
4.3	Investigating radiative feedback differences in terms of stability differences	61
4.4	Explaining stability differences in hist-aer in terms of extratropical forcing	67
4.5	Summary and conclusions	70
5	Time-evolution of feedbacks in the historical period	72
5.1	Introduction	72
5.2	Data and methods	75
5.3	Dependence of climate feedback on forcing agent	78
5.4	Time-evolution of radiative feedbacks on a forcing agent basis	82
5.5	Reconstructing the historical case from individual historical forcing experiments	86
5.6	Explaining a stronger pattern effect from volcanic forcing	88
5.7	Summary and conclusions	92
6	Conclusion	97
6.1	Summary of Thesis Achievements	97
6.2	Future Work	99
A	Why $\frac{X_{hist}}{T_{hist}} - \frac{X_{GHG}}{T_{GHG}}$ anti-correlates well with $\frac{X_{aer}}{T_{aer}} - \frac{X_{GHG}}{T_{GHG}}$	102
	Bibliography	104

¹This chapter is largely a recreation of published work (doi.org/10.1029/2021gl093616)

²This chapter is largely a recreation of published work (doi.org/10.1029/2022gl097766)

LIST OF TABLES

3.1	Experiment names and details for the experiments using common forcing agents in this chapter. All experiments use the same control SSTs and sea ice.	38
3.2	Experiment names and details for the simplified experiments used in this chapter. All experiments use the same control SSTs and sea ice. The heating rates were prescribed as extra terms in the heating rate equations within CAM4’s radiation scheme.	39
4.1	Experiment names and details for the CMIP6 experiments used in this chapter. The “piClim” prefix denotes fixed-SST experiments, while the rest are coupled atmosphere-ocean experiments. piClim-control is an AGCM experiment with SSTs and sea ice climatologies taken from piControl. The fourth column includes three prescribed-SST experiments: piClim-histall, piClim-histGHG and piClim-histaer.	58
5.1	Names and details of the CMIP6 experiments specific to this chapter. Data from experiments described in Table 4.1 are also used, with the same models as before for the piClim-histnat and hist-nat experiments. Unavailable data is marked by an x.	75

LIST OF FIGURES

2.1	An overview of the fluxes that make up Earth’s energy budget. Taken from Wild et al. (2015) (CC BY)	21
2.2	A visual comparison, through temperature profiles, of two version of forcing. (a) instantaneous radiative forcing (IRF) immediately following the introduction of a forcing agent, (b) effective radiative forcing (ERF), where surface temperatures are fixed whilst atmospheric adjustments are allowed to happen.	24
2.3	An example of a Gregory plot (for a 4xCO ₂ with the coupled atmosphere-ocean climate model HadGEM3-GC31-LL, single ensemble member r1i1p1f3). A linear regression is applied to annually-averaged values of surface temperature change and net downward flux.	26
2.4	Cloud radiative adjustments in shortwave (SW), longwave (LW), and in the net across a variety of CMIP5 models and forcing agents. Taken from Smith et al. (2018) (CC BY 4.0)	29
3.1	Heating profiles of all of the vertically-localized heating experiments in this chapter. Dashed lines show the heating rates defined in Eq. 3.1 of the main text, while the solid lines show the profiles interpolated from the model grid. 40	
3.2	(a–b) Globally- (weighted by clear-sky surface albedo) and time-averaged histograms of the cloud kernels obtained from Zelinka et al. (2012) (c) Globally- and time-averaged histogram of the changes in satellite-observed cloud fraction between the 2×CO ₂ and control cases.	41

3.3	Profiles for (a) applied heating rates Q , as well as (b) temperature, (c) relative humidity, and (d) cloud fraction (CLD) change profiles for vertically localised heating experiments with $\phi = 875$ hPa (<i>blue</i>), $\phi = 625$ hPa (<i>orange</i>), and $\phi = 375$ hPa (<i>green</i>). Profiles shown in solid lines are the average of the positive and negative of the negative vertically localised heating tendencies set at the same heights and magnitudes. The profiles for the heating experiments are shown separately as dashed lines. Heating profiles are interpolated from model input, rather than those defined in Eq. 3.1. All values shown are global means. The grey horizontal lines demarcate bounds between high, mid, and low levels according to the ISCCP simulator. Changes to temperature, relative humidity, and cloud are obtained as the difference between the equilibrium fixed SST state and a control state.	43
3.4	(a) Changes to bulk tropospheric stability (BTS, see Eq. 3.3), normalised by the vertically integrated tropospheric (200–1000 hPa) heating anomalies, versus the vertical centre of mass of tropospheric heating (Eq. 3.2) for each experiment performed in this study. (b) Normalised cloud-radiative adjustments against normalised BTS changes. (c) Normalised cloud-radiative adjustments versus heating centre of mass. Least-squares linear fits are shown by the dashed lines. Only experiments with significant tropospheric forcing are shown, which excludes the 0 hPa and 125 hPa forcing experiments. (d) Cloud-radiative adjustments from the six vertically localised forcings centered between 875 and 250 hPa, in 125-hPa increments, broken down into contributions from low (e), mid-level (f), and high (g) cloud changes. The lowest altitude forcing at 1000 hPa is excluded because it is truncated; furthermore, the two highest altitude forcings are excluded because they are mainly in the stratosphere, and have little impact on clouds. Crosses represent the averages between the responses to heating anomalies and the negative of the responses to cooling anomalies at the same altitudes and of the same magnitudes, with the thin horizontal lines around the crosses representing the results from the heating experiments.	45
3.5	Maps of adjustments to low cloud (a–b), high cloud (c–d), and total cloud (e–f) fractions, and kernel-derived longwave (g–h) and shortwave (i–j) cloud radiative adjustments for the experiments involving localised atmospheric heating at 875 hPa (<i>left</i>) and 375 hPa (<i>right</i>).	46
3.6	Global-mean vertical heating anomaly profiles (Q) and rapid adjustments of temperature (T), relative humidity (RH) and cloud fraction. Shown are the results from the original cases (<i>blue</i>) and linear combinations of the results from the vertically localised forcing experiments (<i>orange</i>) including an appropriate surface term (Eq. 3.5). Fitted heating rates are best fits to the original heating rates (Eq. 3.4) in all cases (except for $2\times\text{CO}_2$ where we set the stratospheric heating anomalies to zero before fitting). The orange lines for other variables are created from the fits of heating rates (Eq. 3.5).	50
3.7	The same as Fig. 3.6, but excluding contributions from surface forcings. Note that only the 3%Sol case changes significantly.	51

3.8	A comparison of the cloud-radiative adjustments predicted from linearly combining vertically localised forcing experiments (<i>empty bars</i>) versus the radiative adjustments calculated from the relevant experiments themselves (<i>solid bars</i>).	52
4.1	Top-of-atmosphere radiative forcing patterns over years 1995–2015 for hist-GHG (a) and hist-aer (b). The zonal-mean profiles of forcing are shown in (c), where values for hist-GHG (<i>orange</i>) and hist-aer (<i>blue</i>) are the average effective forcings of the relevant AGCM experiments (piClim-histghg and piClim-histaer, respectively) minus control. For nh_extrop (<i>grey</i>) the dashed line shows the prescribed instantaneous forcing whilst the solid line shows the effective forcing calculated from the HadAM3 (atmosphere-only) simulation. Note that the <i>x</i> -axis is scaled by geographical area.	61
4.2	(a) allsky radiative feedback parameter (α , <i>bars</i>) alongside CRE (α_{CRE} , <i>minuses</i>) and clearsky radiative feedback parameters (α_{CS} , <i>pluses</i>) for each model and the multi-model mean, as the difference from hist-GHG values. (b) Difference of dS/dT from hist-GHG values. (c,e,g) Net (c) clearsky, (e) CRE, and (g) allsky radiative feedback parameters against net dS/dT . (d,f,h) as row above, except with values relative to hist-GHG. Confidence intervals in (c-h) denote $\pm$ one standard deviation based off combined regression and ensemble member uncertainties. The Pearson correlation ρ is shown in (c-h), across models and both excluding (<i>black</i>) and including (<i>grey</i>) the data points from the HadSM3 experiments. The tropical-only forcing experiment is included in (c,e,g), as part of the inter-model trend, but excluded (both in plotting and ρ) in (d,f,h) for clarity and to focus on the HadSM3 experiments that are more similar to hist-aer and hist-GHG. Also shown are <i>p</i> -values for the statistical significance of the correlations.	63
4.3	A similar plot as Fig. 4.2, comparing the all-sky radiative feedback parameter (α) to CRE (α_{CRE}) and clear-sky radiative feedback parameters (α_{CS}) for each model and the multi-model mean. Values are shown as absolutes (<i>top</i> row) and as the difference from hist-GHG values (<i>bottom</i> row). All panels contain a 1:1 line (<i>solid grey</i>).	64
4.4	A geometric explanation of why the historical all-forcing feedbacks do not lie between the historical AA and historical GHG feedbacks. If the all historical is roughly a linear combination of the historical AA (<i>blue</i>) and historical GHG (<i>orange</i>) results, and the historical AA induces a negative temperature change with negative forcing unlike the positive forcing and temperature change from GHGs, then the all historical feedbacks parameter (the gradient of the <i>green</i> line) will be more positive than the hist-GHG feedback parameter.	66

4.5	MMM zonal-mean profiles of local air temperature regressed onto global-mean surface air temperature. Values shown are absolute (a–b), and relative to hist-GHG (e–f). Also shown are the results from the HadSM3 Northern hemisphere (NH) extratropical forcing (c) and tropical-only forcing (d), and these relative to the HadSM3 2xco2 experiment (g–h). Note that the x -axis is scaled by geographical area.	68
4.6	MMM zonal-mean values regressed onto global-mean surface air temperature for the historical cases (<i>top</i> row) and differenced relative to hist-GHG (<i>bottom</i> row) in terms of (a,f) temperature, (b,g) estimated inversion strength, (c,h) net upwards radiative response, (d,i) upwards radiative response in clearsky and (e,j) upwards radiative response from CRE. Note that the x -axis is scaled by geographical area.	69
5.1	Annual-means of effective TOA radiative forcing over time (<i>top</i>) and global average surface temperature change (<i>bottom</i>) in hist-nat over the 1850–2015 historical period. The seven largest peaks of negative forcing of the period (corresponding to the seven most climatologically significant volcanic eruptions) are marked by red dots, whilst years of measured negative forcing (i.e. volcanic forcing, which makes up “hist-nat-volc”) are shaded in grey. Instantaneous radiative forcing (IRF) from changes in downwelling shortwave at the TOA is shown in blue.	76
5.2	MMM radiative feedbacks α against dS/dT , with values calculated from regression over the full historical period (1850–2015) for hist-GHG, hist-aer, and hist-nat. Also shown are values for amip-piForcing over available years (1870–2015), and regression over hist-nat using only volcanic years of hist-nat (“hist-nat-volc”, see Fig. 5.1 for selection). Gradients m of least-squares fits are shown in each panel. Error bars denote estimated regression errors only.	79
5.3	Forcing over time in the historical case and for simulations of component historical forcing agents. Values were calculated from RFMIP piClim-histxxx experiments with piClim-control as the control. Vertical grey lines demarcate the three different periods of different forcing flavours.	81

5.4	<i>Top and middle rows:</i> Time-evolving feedback parameter α (<i>top</i> row) and dS/dT (<i>middle</i> row) for various historical experiments as calculated using 30-year sliding windows. Coloured shaded regions denote regression errors. Dashed time-series in the top row show best linear fits of dS/dT to α using the non-overlapping windows from 1855–2015. Pearson correlation coefficients of these fits to α (ρ), as well as regression significance (p), are shown. In brackets, we show ρ and p for fits using overlapping windows. For hist-nat, both regressions ignore periods free from significant eruptions. Peaks of volcanic forcing are shown by red crosses for the hist-nat and historical cases. <i>Bottom row:</i> year-by-year scatter of α and dS/dT from 30-year windows. Circles with black edges show non-overlapping windows, whilst dots show windows that overlap with these. Linear fit coefficients for the gradient σ and intercept τ are shown, across independent windows (black, solid line fit) and all available windows (grey, dotted line fit). The window free of significant volcanic eruptions is ignored for hist-nat as above, and is shown by a white circle with a black edge in panel (m).	83
5.5	MMM responses R against surface temperature change T, for (a) hist-aer, (b) hist-GHG, (c) hist-nat, and (d) the historical experiments. Each dot marks a yearly average, with linear best-fits (of gradient α) in solid grey.	83
5.6	Model results from the historical (<i>black</i>) compared to the sum of model results from hist-nat, hist-GHG, and hist-aer cases (<i>grey</i>), along with the reconstruction using fixed α for AA and GHG forcings but time-varying α for natural forcing (<i>red</i>, detailed in the text). (a-c) 3-year running means of (a) the radiative response, (b) 50S-50N EIS changes over ocean, (c) surface air temperature change. (d-f) 30-year regressions of same variables as to their left (d-e) over surface temperature and (f) over time. Note that for T and dT/dt, the reconstruction and sum are identical as the reconstruction uses the sum of model output temperatures.	87
5.7	Zonal mean radiative forcing by latitude for various volcanic eruptions. Values are time-averages using year preceding, the year of, and the three years following the eruption.	89
5.8	Radiative feedback parameter α against dS/dT, both calculated using regressions centred on various volcanic eruptions, using 15-year windows (<i>left</i>) and 30-year windows (<i>right</i>). Also shown are values calculated across the full 1850–2015 time series for hist-GHG. All values are for the MMM.	90

5.9	(a) Zonal patterns of forcing, normalised by global average forcing as in Fig. 4.1 but including forcing from the two main volcanic events in the 1850–2015 period. (b-c) Patterns of global average TOA effective forcing (b) and surface temperature change (c) in climatological tropical SST space. <i>y</i> -axis values are normalised for each experiment by the mean value over tropical SSTs i.e. such that all values would be unity for uniform tropical forcing/temperature change. Vertical grey dashed lines delimit the 30th percentile bounds as used in the SST# measure, whilst horizontal dashed lines denote average values on each side of the delimitation.	91
5.10	Map of climatological SSTs within the tropics, calculated from piControl experiments.	92

Chapter 1

INTRODUCTION

1.1 Introduction

Long-term future global warming is driven by forcing agents, primarily anthropogenic greenhouse gases, that break the balance of incoming and outgoing radiation at the top-of-atmosphere (TOA), causing an increase in net downwards radiative flux in the earth system (Cook et al., 2016; Forster et al., 2021). Unfortunately, there is still significant uncertainty in the estimates of global warming under even highly idealised scenarios (Andrews, Gregory, Webb, & Taylor, 2012). This fact complicates climate change mitigation efforts due to the large range of potential consequences (Watkiss et al., 2018). The uncertainty arises both from rapid radiative adjustments, which modify forcing whilst unrelated to surface warming, as well as radiative feedbacks, which modify the earth's energy imbalance as driven by surface temperature change.

The radiative effects of cloud changes have been highlighted as important for understanding the uncertainty in both rapid adjustments (Colman & McAvaney, 2011; Zelinka et al., 2013; Smith et al., 2020) and radiative feedbacks (Andrews, Gregory, Webb, & Taylor, 2012; Webb et al., 2017). With uncertainties for future climate change not only in the intermodel spread in response to CO₂ forcing, but also on a forcing agent basis for both radiative feedbacks (Marvel et al., 2016; Shindell, 2014; Smith & Forster, 2021) and radiative adjustments (Smith et al., 2018), it appears therefore of interest to gain a currently lacking general understanding of the forcing agent dependence on cloud adjustments and

feedbacks. Recent work points to the SST pattern effect (Stevens et al., 2016) being important in determining radiative feedbacks. The effect of different warming patterns has been described in terms of different stability changes which are then tied back to cloud formation and thus radiative effects, particularly in feedbacks (Ceppi & Gregory, 2019; Kamae et al., 2019; Mackie et al., 2021). A process understanding, of cloud radiative changes as related to stability changes, would allow for better evaluation of future warming, not just by ascertaining how different forcing agents impact warming, but also by providing a way of evaluating model inaccuracies so as to reduce the intermodel spread.

1.2 Objectives

This project aims to improve our understanding of what aspects of radiative forcing lead to differences between forcing agents, and how this results in historical surface temperature change. From this, the main questions addressed by this PhD research are:

1. **How and why do different rapid radiative adjustments occur for different forcing agents?** What mechanisms unrelated to surface warming explain differences between forcing agents?
2. **Are feedbacks different between forcing agents, and if so what forcing features explain different feedbacks?** Do long-term surface warming-driven feedbacks differ as a result of differences in the pattern of forcing?
3. **Can the evolution of historical surface temperatures be understood by understanding the contributions of individual forcing agents?** Does the time-evolution of historical forcing agents explain the time-evolution of feedbacks in the historical case?

We seek to answer these questions through chapters 2–5. Following this introduction, chapter 2 serves as a literature review that details the background physical theory that is required to understand the methods and physics used in this thesis. It then goes on to review relevant work that has been done using established physical models, and provides

an overview of open questions in the field. Chapters 3–5 deal directly with the main questions presented above, in the same order. Chapter 3 assesses the impact of vertical structure on rapid cloud-radiative adjustments using the Community Atmosphere Model 4 (CAM4). Chapter 4 makes use of historical experiments from phase 6 of the Climate Model Intercomparison Project (CMIP6) to explain differences in radiative feedbacks as related to stability changes for the two most prevalent historical forcing agents (aerosols and greenhouse gases) over the full (1850-2014) historical period. Chapter 5 also uses data from CMIP6 historical experiments, this time to explain the historical time-evolution of radiative feedbacks using the most prevalent historical forcing agents. Two of these chapters (3 and 4) have been adapted from published peer-reviewed journal articles (Salvi et al., 2021, 2022).

Chapter 2

A REVIEW OF THE FORCING-FEEDBACKS-ADJUSTMENTS FRAMEWORK

To begin, this chapter provides an overview of key concepts that form the forcing-feedbacks-adjustments framework for understanding future surface warming. Following this, we review past works that reveal unresolved problems relating to different aspects of this framework, in particular relating to adjustments and feedbacks. This serves as a literature review, meaning that it does not contain original results. Instead, it forms a basis for the questions answered later on in this thesis.

2.1 Introduction

Predictions of future global warming are uncertain given the large model spread. Even when looking at basic scenarios such as a doubling of CO₂, the IPCC 5th assessment report put the likely range of equilibrium warming at 1.5–4.5 °C (Myhre et al., 2014). The model spread has persisted over the years (Charney et al., 1979), despite extra computing power and model complexity. In the recent 6th assessment report, a narrower likely warming range of 2.5–4.0°C was only provided by a Bayesian combination of multiple lines of evidence, including paleo records and observations (Forster et al., 2021). The reduced 2.5–4.0°C range still allows for a wide range of climate impacts, e.g. high risk of extinction for up to 18% of terrestrial animals at 2°C warming versus up to 39% at 4°C (Portner et al., 2022), and highlights that different factors are at play in determining

climate outcomes even after extra-model evidence is considered and even when simple climate forcing scenarios are considered. That additional lines of evidence are required to reduce the large spread in model predictions, suggests that there is an important question as to which are the key physical errors in climate models. A better understanding of the mechanisms through which models simulate warming is thus desirable, so that the accuracy of models can be better assessed.

For actual future climate change, predictions of global warming involve more complex forcing scenarios of evolving mixtures of different forcing agents. As such, it is vital to understand forcing agent dependence on the various factors that affect future global warming. Doing so allows for both a better understanding of how given agents affect the climate, useful for deciding on future emissions policy, whilst also informing on differences in models by revealing differences in how they handle distinct forcing agents.

The questions we seek to answer in this chapter include:

- How do we describe future temperature change and its relationship to the global energy budget?
- What is our current understanding of radiative adjustments and feedbacks?
- How can we explain adjustment and feedback differences among forcing agents?

Sections 2.2–2.7 of this chapter address these questions in the same order, with the final section 2.8 discussing the open questions of work analysed in its preceding sections.

2.2 The global energy balance

In order to fully grasp global warming, the energetics of the Earth system must be first understood. These are also necessary to understand how there can be patterns and structure to the energetics of Earth that ultimately depend upon the balance of energy flux of the system. From the first law of thermodynamics, the heat added to

a closed system is equivalent to the change in internal energy of the system minus the work done by the system (Mandl, 2013). The Earth is an approximately closed system with negligible mass exchange (solar wind and emissions to space) but significant energy exchange (incoming and outgoing radiation). There is also little work done by the Earth on its surroundings. Hence, the balance of energy coming and going into the Earth should translate principally into a change in its internal energy. This balance of energy is what is called Earth’s “energy budget” (Hartmann, 2016).

There are various effects that impact this budget, which is shown visually in Fig. 2.1. Different fractions of incoming solar shortwave (SW) radiation are absorbed by the atmosphere, reflected back into space from the atmosphere and clouds or by the surface, or absorbed by the Earth’s surface which causes surface warming (and some minor changes in chemistry e.g. from photosynthesis). The Earth emits longwave (LW) radiation due to its temperature. Some of this is absorbed and re-emitted back downwards by the atmosphere as various GHGs such as water vapour and CO₂ provide good absorption of LW, to result in reabsorption by the surface. A portion of the LW that is absorbed by the atmosphere is emitted to space after being absorbed and re-emitted upwards by the atmosphere. There is also an exchange of heat from the surface to the atmosphere in the form of sensible heat, with convective and conductive heat transfer vertically, and latent heat, in the form of melting snow and ice, evaporation, and precipitation (Hartmann, 2016).

Cloud interactions with radiation are complex. Formed of liquid water droplets, ice crystals or a mixture of the two, clouds are efficient absorbers of LW radiation. Dependent on their optical thickness and altitude they can substantially reduce the amount of surface LW emission escaping to space (Peixoto & Oort, 1992; G. A. Schmidt et al., 2010). Clouds also often have higher albedo than the surface below them, meaning that they contribute to increased reflection of incoming solar radiation. As such, they provide an upwards SW flux (Hartmann, 2016). This results in different clouds having different radiative effects, depending on how they interact with SW vs. LW. For example, low clouds can

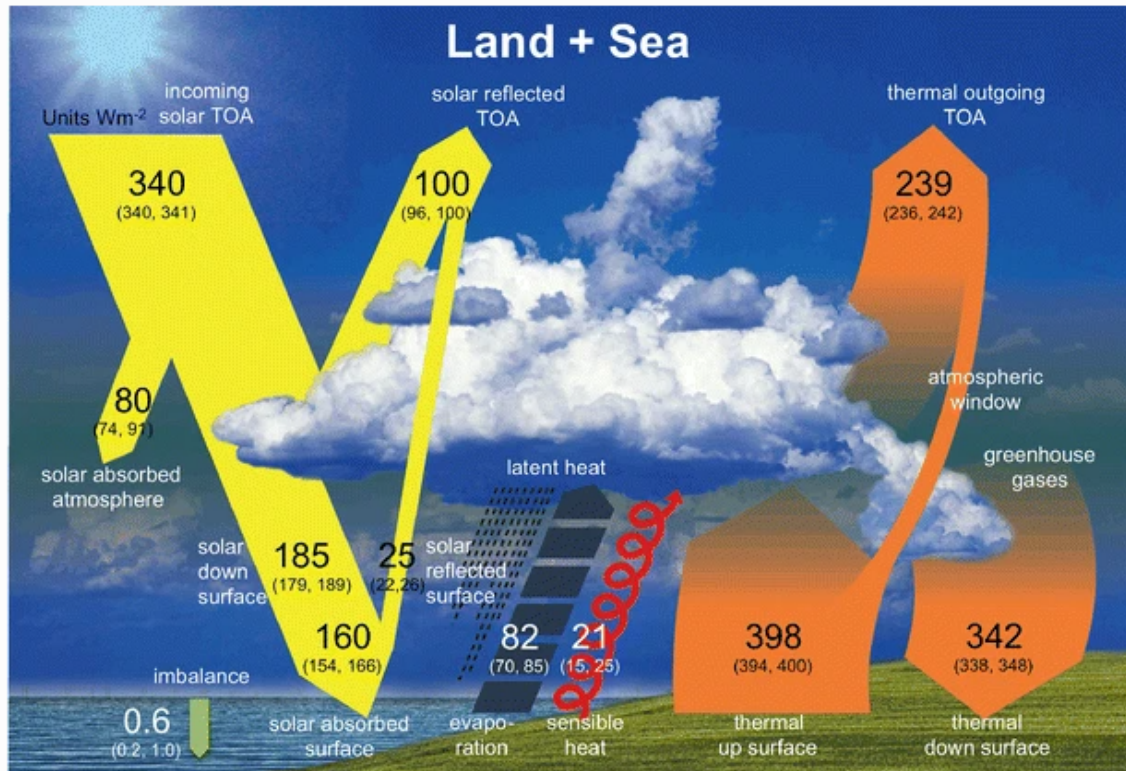


Figure 2.1: An overview of the fluxes that make up Earth's energy budget. Taken from Wild et al. (2015) (CC BY)

provide significant upward flux, for a net cooling effect. This is because they are at similar temperatures to the surface and thus have little effect on LW, whilst they reflect incoming SW. High clouds, on the other hand, can provide little net or even a net downward flux for a warming effect. This is because they have much lower temperatures than the surface and thus effectively absorb LW without emitting much to the top of atmosphere (TOA) (P. C. Taylor, 2012).

Aside from their height, various other aspects affect the impact of clouds on the radiative budget, including cloud fraction, the clear-sky state, microphysics, and optical depth. Cloud fraction, in isolation, affects CRE in a straightforward way, with greater cloud amount related to a stronger cloud effect. Similarly, the clear-sky state is relevant: high clouds will trap less LW if there is less underlying LW emission from the surface. Cloud microphysics are complex, affected by aerosol-cloud interactions which can see albedo changes from changing water droplet size and/or formation of ice in mixed-phase clouds (Boucher et al., 2013). Cloud optical depth is especially important for high clouds as

the amount of SW reflectivity and LW trapping change at different rates with optical depth (Roeckner et al., 1987; Schlesinger, 1988). In effect, this means that reductions in optical depth of high clouds lead to a net downward radiative change as there is a greater reduction in SW reflection than LW trapping (Zelinka et al., 2012b).

In order to measure the impacts of clouds on the radiative budget, we discuss the cloud-radiative effect (CRE). This is measured as the difference between clearsky and allsky radiative fluxes at the TOA:

$$\text{CRE} = N - N_{\text{clearsky}} \quad (2.1)$$

for net downward radiative flux N (W m^{-2}) and without clouds present N_{clearsky} (W m^{-2}) i.e. the CRE is a measure of how much the radiative budget is changed by the presence of clouds. The interactions of clouds with LW and SW mentioned in the previous paragraphs do not necessarily translate into the same effects on the CRE. An example of this is that clouds have high albedo, which causes high SW reflection. However, if clouds lie above ground that has a high albedo then the CRE from these is minimal since the ground would effectively reflect SW radiation anyway, so there is little change in SW reflection with or without clouds in this case. Cloud overlap also complicates CRE (Collins, 2001), for example with higher clouds obscuring lower clouds for little effect on overall SW reflection.

2.3 The forcing–adjustment–feedback framework

A formulation of global average surface temperature change T (K), as a response to some imposed anomalous radiative flux across the top of the atmosphere which we denote “forcing” F (W m^{-2}), is given by:

$$N = F - \alpha T \quad (2.2)$$

for climate feedback parameter α ($\text{W m}^{-2} \text{K}^{-1}$). This equation uses a negative sign placed in front of a generally positive α . This convention is chosen so that the αT term can be understood as the radiative response of the Earth system to a change in incoming energy F . Earth temperatures increase in response to positive F , resulting in increased outgoing energy flux that acts to cancel out this radiative imbalance (Gregory et al., 2004). A positive F is required for surface warming, whilst a negative F results in cooling.

With the application of forcing, the resultant change in temperatures is referred to as the “response”. For comparison between models, the “Equilibrium Climate Sensitivity” (ECS) is defined as the response at global-mean surface temperature equilibrium following a doubling in the atmospheric concentration of CO_2 from levels prior to the industrial revolution at around 280 ppm (Etheridge & Steele, 1996; Forster et al., 2021; Knutti et al., 2017; Myhre et al., 2014). We refer to “forcing agents” throughout this thesis as any change to the system that results in forcing. These changes can be anthropogenic, such as an increase in GHGs from emissions, or natural, such as a change in solar irradiance.

Because changes to the Earth’s climate are not generally instantaneous, but rather occur over different timescales and at different points in the process of climate change, it is helpful to break these down by timescale. When trying to establish how the final state of a simulation arose, it is vital to understand how the simulated system evolved over time to reach its end point.

Before any changes are allowed to occur within the system, the presence of an introduced forcing agent induces an immediate change in radiative fluxes. This is referred to as instantaneous radiative forcing (Hansen et al., 1997). While the instantaneous forcing is straightforward to assess, there are additional changes that occur to the climate upon the introduction of a forcing agent, without being tied to changes in global surface temperature. These so-called “adjustments” can have a significant effect on total downward flux.

Because they are not responses to global mean surface warming, these adjustments are

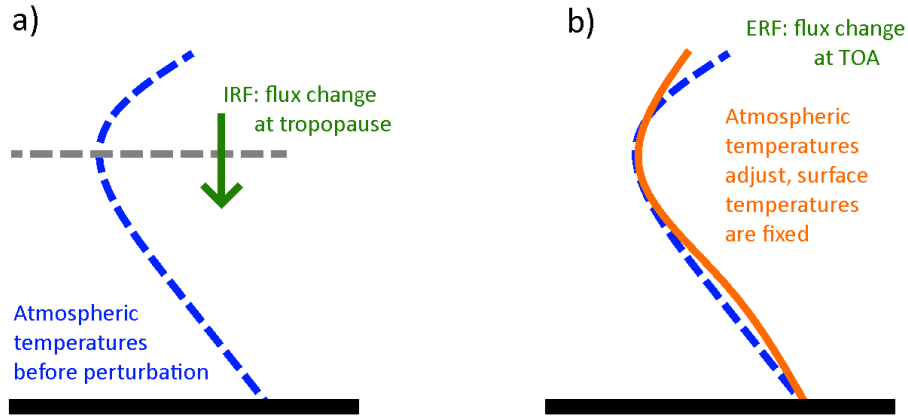


Figure 2.2: A visual comparison, through temperature profiles, of two version of forcing. (a) instantaneous radiative forcing (IRF) immediately following the introduction of a forcing agent, (b) effective radiative forcing (ERF), where surface temperatures are fixed whilst atmospheric adjustments are allowed to happen.

generally treated as part of the forcing, with the sum of instantaneous forcing and adjustment called “effective radiative forcing” (ERF; Myhre et al., 2014). The adjustments can happen on different timescales, with cloud adjustments often believed to occur within a few days or weeks (Kamae et al., 2015) while adjustments such as changes in atmospheric circulation can take longer than a month (Wu et al., 2011). To distinguish these changes from relatively slow responses to SST warming, the term “rapid adjustment” is used, although there is as of yet no formally agreed definition for the cutoff timescale in this distinction (Sherwood et al., 2015).

Feedbacks, by contrast, are responses to the forcing, mediated by surface warming, that in turn affect the energy budget (Forster et al., 2021; Myhre et al., 2014). These therefore happen over longer timescales as the system approaches equilibrium. An example of this would be: as forcing occurs the Earth warms, which results in atmospheric warming that causes an increase in saturation water vapour as determined by the Clausius-Clapeyron relationship. Given that relative humidity is approximately constant, this leads to increased actual humidity and further heating as the water vapour acts as a GHG. One way of defining a feedback is as one that “changes the sensitivity of the response” (Hartmann, 2016), with positive and negative feedbacks increasing and decreasing the magnitude of

the response respectively.

From the concept of ECS, given that α may vary among forcing agents, we refer to the concept of “effective climate sensitivity” (Murphy, 1995) which is the temperature change $T = F_{2\times}/\alpha$ that would be reached at equilibrium with the ERF from a doubling of CO_2 ($F_{2\times}$) for a given constant feedback parameter α . Although it is formally defined for CO_2 only, when applied to other forcing agents this metric serves as an indication of the efficiency of this forcing agent at changing surface temperatures, relative to the amount of forcing exerted. We see from this that forcing agents with a lower α than we see for a doubling of CO_2 will cause relatively amplified temperature change, and conversely for those with higher α .

2.4 Calculating forcing, adjustments, and feedbacks

In order to establish the impact of forcing agents on the Earth’s climate, to determine how these affect the forcing patterns that dictate the overall warming, computational models are used to simulate the effects without requiring global real-world experiments. The measures used to calculate radiative forcings from a given agent are inherently idealised, by computational necessity. However, idealised simulations can still provide answers to important physical questions.

Throughout this thesis, global climate models are used to calculate forcings. For calculating instantaneous forcings (Fig. 2.2a), offline radiative transfer codes can be used e.g. the Parallel Offline Radiative Transfer (PORT) tool for CAM4 (Conley et al., 2013). For ERF (Fig. 2.2b), Atmosphere General Circulation Models (AGCMs) are used to allow for atmospheric adjustments to occur. Though ideally all surface temperatures should be held fixed, in practice this is difficult to do (G. A. Schmidt et al., 2006) and hence SSTs and sea ice are held constant whilst land temperatures are allowed to adjust. Hence, these are referred to as “fixed SST” experiments. The land temperature adjustments can be accounted for using radiative kernels, but this has little impact on

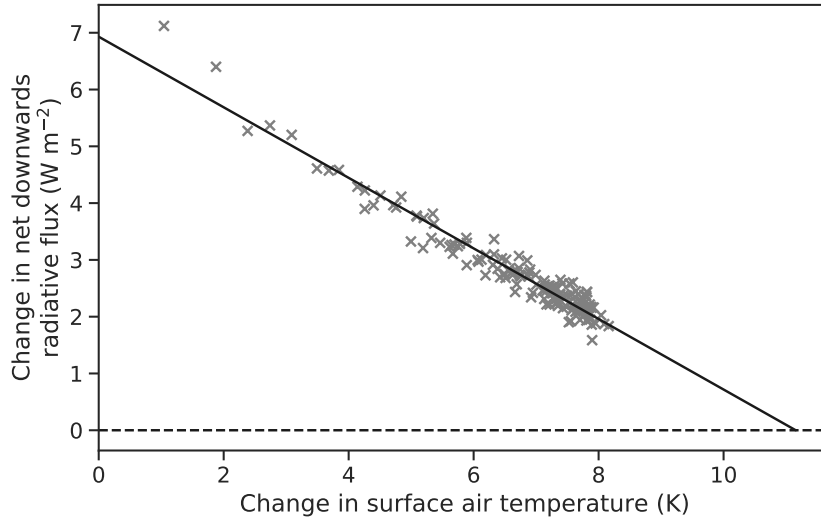


Figure 2.3: An example of a Gregory plot (for a $4\times\text{CO}_2$ with the coupled atmosphere-ocean climate model HadGEM3-GC31-LL, single ensemble member r1i1p1f3). A linear regression is applied to annually-averaged values of surface temperature change and net downward flux.

estimates of effective forcing (Richardson et al., 2019). With a scenario set up, with some forcing agent introduced to a reference state, a simulation is run by a model generally until equilibrium has been reached. For further accuracy, an average over time of the climate around equilibrium is taken in order to remove the noise of natural climate variation. Though an ensemble average of model runs is still ideal, variability from SSTs is not present since these are prescribed such that significant internal variability is removed. The forcing is then the total downward flux at equilibrium, compared to the reference climate state without the forcing agent (Hansen et al., 2002; Rotstayn & Penner, 2001).

In order to calculate radiative feedbacks, we can look at equation 2.2. From this equation, it is easy to see that if F is held constant, and assuming a roughly constant α , then there should be (at least approximately) a linear relationship between N and T . With a scenario where a suddenly introduced forcing agent is held fixed, a model can be run with the simulated climate allowed to freely respond to the forcing. Taking monthly-averaged model outputs allows for a series of global temperature change values alongside the contemporary downward flux. These values must be relative to a control simulation of the same setup except without the application of the forcing agent. A linear regression of this series with N against T then gives the negative of α as the gradient, as shown

in figure 2.3. This is referred to as the “Gregory” method, which gives an estimate of equilibrium surface temperature change, as the x -intercept, without the need for running experiments to equilibrium (Gregory et al., 2004). This relies on approximately constant gradients/feedbacks, however, which is not necessarily valid (Andrews et al., 2015; Dong et al., 2020; Proistosescu & Huybers, 2017; M. A. A. Rugenstein & Armour, 2021). For evolving feedbacks, the same principle of linear regressions can be used by performing regressions over smaller time windows, instead of the full time series (Gregory et al., 2020).

In order to apply this method, most previous studies have used a fully coupled atmosphere-ocean model with yearly-averaged output, with slowly warming oceans and annually averaged snapshots of radiative fluxes (Knutti & Rugenstein, 2015; Stouffer, 2004). If experiments are performed with relatively small forcings, i.e. ones that cause a response that is similar or smaller in magnitude than natural variability, then an ensemble average is essential to separate the response from such natural variability and/or noise.

2.5 The forcing agent dependence of radiative adjustments

The first consideration with regards to understanding model and forcing agent differences in warming predictions is the forcing itself. The intermodel spread of instantaneous forcing generally does not explain the spread in ERF (Richardson et al., 2019), which really means that adjustments are large enough to dominate model disagreements in ERF. One of the main sources of model discrepancy in measuring the ECS of Earth is cloud effects (Andrews, Gregory, Forster, & Webb, 2012; Myhre et al., 2014). With a rapid tropospheric adjustment that occurs for a doubling of CO₂, significant changes to clouds occur in most models, which leads to a non-negligible cloud adjustment (Andrews & Forster, 2008; Gregory & Webb, 2008). Smith et al. (2020) have further highlighted a forcing agent dependence of radiative adjustments, with much of this also driven by

cloud radiative adjustments.

Many past works have focused on adjustments to CO_2 , finding consistently positive (amplifying) adjustments. CO_2 forcing heats the lower troposphere, which results in local drying and stabilisation of the boundary layer, causing a lowering in height of the cloud layer and reduction of low cloud (Kamae & Watanabe, 2013; Kamae et al., 2015; Zelinka et al., 2013). This results in a positive SW cloud adjustment as less incoming solar radiation is reflected back, which means that the overall cloud adjustment is positive as LW effects are small relative to SW effects (Zelinka et al., 2013).

Many other studies have investigated adjustments to black carbon (BC) forcing. Studies looking at multiples of pre-industrial amounts highlight a negative adjustment driven by increases in low cloud and reductions in high and mid-level cloud (Stjern et al., 2017). Several others, however, highlight the impact of injection height of BC aerosol. Cloud reductions occur near injection level, from warming and cloud burnoff. Conversely, cloud increases can occur above the injection height as this promotes convection upward, whilst cloud cover can increase below the injection height as convection below is suppressed. These effects result in increasingly negative adjustments with injection height (Ban-Weiss et al., 2012; Koch & Del Genio, 2010; Samset & Myhre, 2015; Sand et al., 2020). This understanding comes from analysis of aerosol-radiative and semi-direct effects without in-depth analysis of aerosol-cloud interactions i.e. “indirect effects” which occur via altered cloud microphysics such as changes in cloud condensation nuclei (Hansen et al., 1997; Charlson et al., 1992). As noted by Bellouin et al. (2020), these aerosol-cloud interactions can be substantial for rapid adjustments and depend strongly on model configurations. However, although our understanding of aerosol-cloud indirect effects is limited, the highlighted works demonstrate the impact of changing energy input within the atmosphere even in the absence of indirect aerosol effects.

The limited further work that has looked into adjustments in response to a variety of forcing agents shows that the forcing agent dependence on adjustments is not limited to CO_2 and BC. Smith et al. (2018) have highlighted that adjustments from different

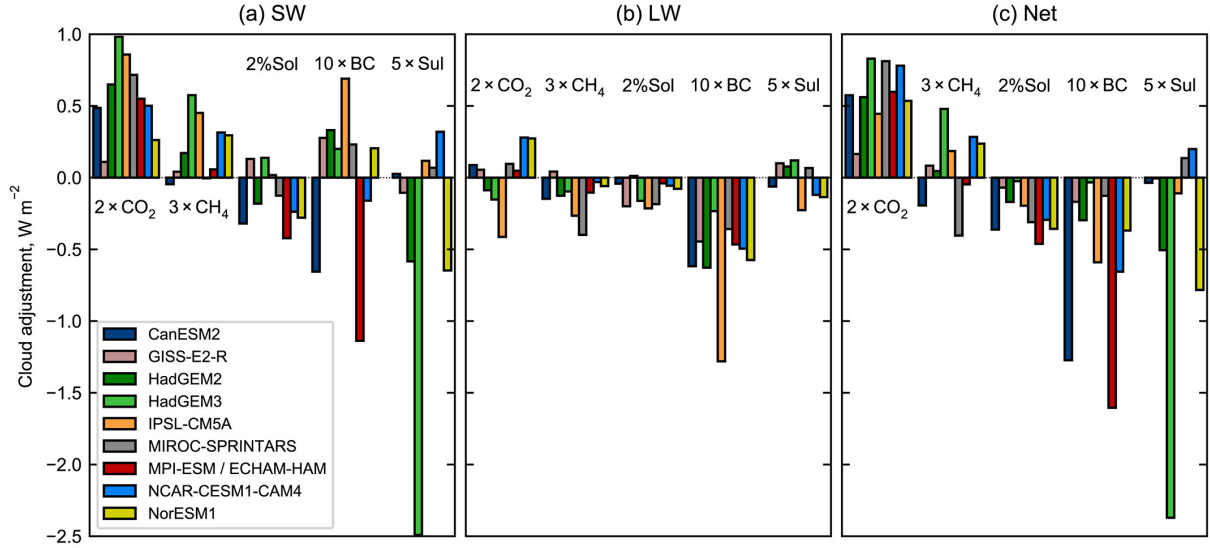


Figure 2.4: Cloud radiative adjustments in shortwave (SW), longwave (LW), and in the net across a variety of CMIP5 models and forcing agents. Taken from Smith et al. (2018) (CC BY 4.0)

agents not only have different magnitudes but are also different in their underlying physical mechanisms. Here we focus on the cloud-radiative adjustments (Fig. 2.4) and their dependence on forcing agent. Models generally show that CO_2 adjustments enhance SW warming through a reduction in low clouds. BC, on the other hand, shows consistent LW cloud adjustments between models, but not SW ones. Cloud adjustments from solar forcing lead to diminished warming, whilst they are similar to but weaker than CO_2 adjustments for methane (provided methane’s SW absorption is accounted for in the models), and sulphate adjustments vary significantly owing to aerosol-cloud interactions. Volcanic forcing has also been analysed, with rapid adjustments that counteract such forcing through cloud effects (Gregory et al., 2016; A. Schmidt et al., 2018). Largely lacking, though, has been a general explanation for how different forcing agents result in their respective adjustments.

2.6 The agent dependence of radiative feedbacks

The second key component of future warming is the feedbacks. These have been shown to vary significantly across models (Andrews, Gregory, Webb, & Taylor, 2012; Caldwell

et al., 2016; Colman & Hanson, 2017; Vial et al., 2013; Zelinka et al., 2020), driving most of the inter-model spread in ECS. Of interest to predictions of the future is whether there is also a dependence of feedbacks on forcing agent. Some work has highlighted differences in feedbacks between historically significant forcing agents (Marvel et al., 2016; Shindell, 2014; Smith & Forster, 2021), though other work has suggested that such differences might be negligible (Richardson et al., 2019). If there are differences, then evolving ratios of different forcing agents will result in evolving feedbacks that result from a forcing signal instead of just from natural variability. Analysis of feedbacks over time in model historical experiments suggests that this is indeed the case, with important effects from GHGs, anthropogenic aerosols (AAs), and natural (mostly volcanic) forcings changing over time (Gregory et al., 2020).

Even in cases where forcing agents do not vary in time, feedbacks show time-evolution (Andrews et al., 2015; Dong et al., 2020; Proistosescu & Huybers, 2017; M. A. A. Rugenstein & Armour, 2021). Reasons given for this have included the SST “pattern effect” whereby the radiative response of the planet depends not just on global mean temperature change, but also on the pattern of SST change (Stevens et al., 2016). This effect is based off of different regions causing different radiative responses for the same amount of local warming. Per unit of warming, regions such as the western equatorial Pacific cause larger feedbacks (lower climate sensitivity), whilst regions like the eastern Pacific and Southern Ocean cause smaller feedbacks (higher climate sensitivity). As such, the same global mean mean warming can result in greater/lesser feedbacks depending on which areas are warmed more. Owing to timescales of ocean heat uptake, as well as evolving forcings (in relevant cases), the pattern of warming evolves over time such that global radiative feedbacks do as well.

Another potential reason for the variation of radiative feedbacks is temperature dependence. Under CO₂ warming, feedbacks have been found to become more amplifying as the planet warms (Bloch-Johnson et al., 2021; Meraner et al., 2013). However, this effect has not been focused on due to time and scope limitations of this thesis.

2.7 The role of stability in radiative adjustments and feedbacks

Though much work has been done on trying to understand adjustments and feedbacks in the context of CO₂ forcing, we desire a general understanding that contributes towards explaining the effects of any given forcing. Given the importance of cloud effects for both radiative adjustments (Kamae & Watanabe, 2013; Zelinka et al., 2013) and feedbacks (Forster et al., 2021), we seek a physical mechanism that may effectively explain the cloud radiative adjustments and feedbacks for different forcing agents. Tropospheric stability is found to be a useful measure for understanding low cloud amount changes (Klein & Hartmann, 1993; Wood & Bretherton, 2006). Although not the only important cloud metric, low cloud cover is an important one for understanding changes to the energy budget as discussed above. Broadly speaking, tropospheric stability is a reframing of lapse rates, which are vertical temperature gradients, into vertical *potential* temperature gradients. These describe whether air parcels are more or less encouraged from travelling vertically, and as such are descriptive of convection (Hartmann, 2016). In this way, though a simplification, stability is useful for understanding cloud cover changes as changes to stability are expected to have clear effects: a stable troposphere suppresses convection, which encourages low clouds, whilst an unstable troposphere encourages convection and thus has the opposite effect.

Radiation is affected in a few clear ways by stability changes. The first way is tied to cloud changes, whereby increased low cloud cover from increased stability leads to more SW reflection from low clouds with little effect on LW. As such, increased stability contributes to SW cooling of the planet. There is also a LW cooling effect of increased stability, unrelated to clouds. Since stability is tied to lapse rates, increasing stability means more efficient LW cooling as temperatures increase more higher up in the atmosphere than near the surface.

Atmospheric stability can be measured in a few ways. One way of doing this is through “lower tropospheric stability” (LTS), which is simply the difference in potential temperatures at 700 hPa (θ_{700}) and the surface (θ_{sfc}) (Klein & Hartmann, 1993):

$$\text{LTS} = \theta_{700} - \theta_{\text{sfc}} \quad (2.3)$$

This measure gives a sense of the planetary boundary layer inversion strength, and has been shown to be related to low-cloud formation (Wood & Hartmann, 2006). LTS has since been refined, however, to provide an improved measure of inversion strength in the planetary boundary layer and low cloud formation. This refinement is known as the estimated inversion strength (EIS), and contains a correction that accounts for convection within clouds being moist rather than dry:

$$\text{EIS} = \text{LTS} - \Gamma_m^{850}(z_{700} - \text{LCL}) \quad (2.4)$$

where Γ_m^{850} is the moist adiabat at 850 hPa, z_{700} the height of the 700hPa surface, and the LCL is the “surface-based lifting condensation level” (Wood & Bretherton, 2006).

Since stability relates to atmospheric temperature gradients, it may describe radiative adjustments as these are rapid atmospheric changes. For radiative feedbacks, the link to stability is more established by prior work. A suggested mechanism of the SST “pattern effect” resulting in different radiative feedbacks is tied to stability change, with temperatures changing by different amounts in high and low convective regions such that atmospheric stability is affected (Andrews et al., 2015; Zhou et al., 2017), itself resulting in cloud, water vapour, and lapse rate changes (Ceppi & Gregory, 2017, 2019; Kamae et al., 2019; Mackie et al., 2021).

Indices of the SST distribution have been introduced as proxies for stability, by comparing temperature change within the Indo-Pacific warm pool (deep convective regions) with average tropical temperatures (which includes less convective regions) (Fueglistaler, 2019). Such measures have provided insight into how different warming patterns affect

stability and convection. Further studies have looked into stability changes for different forcing agents (Ceppi & Gregory, 2019). By understanding how stability affects cloud formation, and thus cloud radiative feedbacks, this stability dependence provides an explanation of the forcing agent dependence of feedbacks. By understanding how different agents force the climate, a qualitative explanation can then be made for how different agents cause different forcing patterns, resulting in different surface warming patterns, which results in different stability changes, in turn explaining different radiative feedbacks. For example, such work lets us point to extratropical versus tropical contrasts to explain stability change differences, since positive tropical forcing tends to cause a greater stability increase than extratropical forcing does (Ceppi & Gregory, 2019). Such tropical forcing then is likely to suppress convection, increasing low cloud cover, and thus have a stronger (deamplifying) radiative feedback as SW reflection occurs from increased low cloud.

2.8 Discussion and conclusions

The forcing–feedbacks–adjustments framework provides a way of understanding future global warming. Key aspects of this framework remain unresolved even in the model world, with intermodel spread in forcing and feedback presenting a limiting factor in the consistency of warming predictions. Key to improving our understanding of these processes is the question of how different forcing agents, i.e. different energy inputs into the earth system, affect energy imbalances in the Earth system at timescales ranging from rapid atmospheric adjustments to long-term radiative feedbacks.

Past simply going through various forcing scenarios to establish what occurs in each case, a general conceptual understanding is desired that can be applied to a variety of scenarios by understanding features of their climate forcings. Past work suggests that atmospheric stability may be a useful physical mechanism for describing multiple aspects of the forcing–feedbacks–adjustments framework, and testing this idea forms the basis of

the present thesis.

With differences in rapid radiative adjustments among forcing agents being largely dependent on clouds (Smith et al., 2018), stability is a natural fit for explaining differences among agents as it is strongly linked to cloud formation (Ceppi & Gregory, 2017, 2019; Kamae et al., 2019; Mackie et al., 2021). For radiative feedbacks, stability has already been established as a good explanation for the SST “pattern effect”, which itself has been described as a driver of non-constant feedbacks. Questions remain regarding to what extent radiative feedbacks differ among forcing agents, but if such differences do exist, then stability may provide a useful mechanism, given its role in feedback inconstancy.

With regards to historical climate change, and future predictions, time-evolving amounts of different forcing agents may be important. If feedbacks do depend on forcing agent, and this can be explained through how these agents affect stability, then it remains to be seen whether the effects of overlapping forcings can be understood through the effects of each component.

Chapter 3

THE ROLE OF STABILITY IN RAPID CLOUD RADIATIVE ADJUSTMENTS¹

As highlighted in our review, cloud adjustments play a particular role in determining agent-dependent rapid radiative adjustments, which in turn strongly modify the effective forcings that drive future temperature change. In this chapter, we perform idealised atmospheric heating experiments to demonstrate a relationship between cloud adjustment and the vertical profile of imposed radiative heating: boundary-layer heating causes a positive cloud adjustment (a net downward radiative anomaly), while free-tropospheric heating yields a negative adjustment. This dependence is dominated by the shortwave effect of changes in low clouds. Much of the variation in cloud adjustment among common forcing agents such as CO₂, CH₄, solar forcing, and black carbon is explained by the “characteristic altitude” (i.e. the vertical center-of-mass) of their heating profiles, through its effect on tropospheric stability.

3.1 Introduction

Radiative forcing quantifies the perturbation to the Earth’s energy budget associated with a particular climatic factor, such as greenhouse gases, aerosols or solar irradiance. Forcing was originally defined as the instantaneous perturbation to the Earth’s radiative budget due to the forcing agent. Later, the concept was refined to include the effect

¹This chapter is largely a recreation of published work (doi.org/10.1029/2021gl093616)

on the tropospheric heat budget of stratospheric adjustment due to radiative changes, which are particularly important for carbon dioxide (Shine et al., 1995). More recently, “effective radiative forcing” (ERF) has become the usual metric (Myhre et al., 2014), which additionally accounts for relatively short-timescale tropospheric adjustments in temperature, moisture and clouds that are direct responses to the forcing, rather than being mediated by surface warming (Andrews & Forster, 2008; Gregory & Webb, 2008; Sherwood et al., 2015). This approach is justified by the fact that ERF is a better predictor of the surface temperature response than instantaneous radiative forcing (IRF) (Richardson et al., 2019) or stratosphere-adjusted forcing (Shine et al., 2003).

The rapid adjustments to radiative forcing have been found to make a substantial contribution to model uncertainty in ERF for a variety of forcing agents including anthropogenic greenhouse gases, aerosols and solar irradiance (Chung & Soden, 2015; Smith et al., 2018, 2020). Cloud adjustments account for a large part of these differences (Andrews, Gregory, Forster, & Webb, 2012; Colman & McAvaney, 2011; Smith et al., 2018; Zelinka et al., 2013), consistent with clouds being an important source of uncertainty in the response to forcing among climate models (Ceppi et al., 2017), especially in the case of aerosols. Aerosol–*radiation* interactions lead to cloud adjustments (known as semi-direct effects) by modifying local atmospheric conditions. Aerosol–*cloud* interactions lead to further cloud adjustments via microphysical changes (known as indirect effects; Bellouin et al., 2020). Only the cloud adjustments due to semi-direct effects will be considered in this chapter since, although indirect effects can be large for aerosols (up to 63% of SW ERF for a fivefold increase in sulphate aerosol, depending on the model) (Smith et al., 2018), the impacts of aerosol–cloud interactions on cloud adjustments are outside the scope of this work.

Several past papers have investigated the mechanisms of cloud adjustments in response to CO₂ forcing (e.g., Dinh & Fueglistaler, 2017; Kamae & Watanabe, 2012, 2013; Kamae et al., 2015, 2019; Zelinka et al., 2013) and to absorbing aerosols such as black carbon (BC) (Ban-Weiss et al., 2012; Bellouin et al., 2020; Koch & Del Genio, 2010; Samset &

Myhre, 2015; Stjern et al., 2020; Sand et al., 2020) or dust (Amiri-Farahani et al., 2017), but there is a lack of process studies involving other forcing agents. Smith et al. (2018) demonstrated a striking forcing agent dependence of cloud adjustments across Coupled Model Intercomparison Project phase 5 (CMIP5) models, with consistently positive adjustments to CO₂ and negative adjustments to solar and BC forcing (we define downward radiative anomalies at the top of atmosphere as positive). However, there is currently limited understanding of how different cloud adjustments arise in response to various instantaneous forcings.

In this chapter we address this knowledge gap by interpreting the forcing agent dependence of cloud adjustment in terms of the spatial structure of instantaneous atmospheric forcing. Specifically, we propose here that the vertical profile of atmospheric heating is a key factor for this forcing agent dependence. For absorbing aerosols, previous studies have identified a dependence of semi-direct cloud adjustments upon forcing altitude: typically positive for boundary-layer forcing, negative for free-tropospheric forcing (Amiri-Farahani et al., 2017; Ban-Weiss et al., 2012; Bellouin et al., 2020; Koch & Del Genio, 2010; Samset & Myhre, 2015; Stjern et al., 2020). Here we expand on these previous works to show that the vertical heating profile explains much of the cloud adjustment in response to a diverse range of forcing agents. This is demonstrated through comparison of idealised and common forcing experiments with a CMIP5-class climate model.

3.2 Data and methods

The simulations used for this chapter were run with the CAM4 model (Neale et al., 2010) in an atmosphere-only configuration with prescribed sea surface temperatures (SSTs) and sea ice concentrations. Prescribed SSTs and sea ice were obtained from the mean of the control climatology. A $1.9^\circ \times 2.5^\circ$ latitude/longitude grid was used with 26 vertical levels. Simulations were run for 20 years with the climatology calculated as the average of monthly-mean data output from the model for all but the first simulated year, during

Table 3.1: Experiment names and details for the experiments using common forcing agents in this chapter. All experiments use the same control SSTs and sea ice.

Experiment name	Description
control	Pre-industrial climate (year 1850)
$10\times\text{CH}_4$	10 times concentrations of methane compared to pre-industrial levels
$2\times\text{CO}_2$	Doubling of CO_2 from pre-industrial level (284.7 ppm) to 569.4 ppm
$10\times\text{BC}$	10 times concentrations of black carbon compared to pre-industrial levels
3%Sol	3% increase in the solar constant

which the atmosphere was adjusting to reach a steady state in the presence of the forcing. The vertical profiles shown in this chapter were linearly interpolated from the model’s 26 hybrid sigma–pressure levels to a 100-level pure pressure grid so as to allow for averaging along pressure levels. This was done using evenly spaced levels between 0 and 1000 hPa.

Instantaneous and effective radiative forcings were calculated for four forcing agents: a 10-fold increase in methane ($10\times\text{CH}_4$), doubled carbon dioxide ($2\times\text{CO}_2$), a 10-fold increase in black carbon ($10\times\text{BC}$), and a 3% increase to incoming solar radiation (3%Sol) (see Table 3.1). These are among the forcing agents as used in the Precipitation Driver Response Model Intercomparison Project (PDRMIP) set of experiments (Myhre et al., 2017), although not all with the same magnitudes as in the PDRMIP experiments. IRF was calculated using the Parallel Offline Radiative Transfer (PORT) tool for CAM4, by taking the difference between perturbed and control instantaneous radiative fluxes (Conley et al., 2013). For these fluxes, simulations were run for 18 months starting on January 1st with outputs every 73 time steps (roughly 1.5 days) as recommended by Conley et al. (2013). The fluxes used were the averages over the last 12 months of these simulations. To obtain the ERF, the difference in mean TOA imbalance was taken between perturbed and control CAM4 experiments with SSTs and sea ice fixed to the control state (Hansen et al., 1997). Note that the CO_2 concentration was doubled only in the radiation scheme in the $2\times\text{CO}_2$ experiment, so the model did not simulate

Table 3.2: Experiment names and details for the simplified experiments used in this chapter. All experiments use the same control SSTs and sea ice. The heating rates were prescribed as extra terms in the heating rate equations within CAM4’s radiation scheme.

Experiment name	Description
atm_4	Homogeneous heating throughout the atmosphere for a horizontally homogeneous vertically-integrated forcing of 4 W m^{-2}
sfc_4	Homogeneous 4 W m^{-2} downwards flux at the surface
vloc_ ϕ hPa	Atmospheric heating defined through Eq. 3.1

a plant physiological response to CO_2 , which has been found to cause significant cloud-radiative adjustments (Doutriaux-Boucher et al., 2009). Furthermore, CAM4 does not simulate aerosol–cloud interactions for black carbon, whose atmospheric concentrations are prescribed.

In addition to the common forcing cases (Table 3.1), experiments were also performed with idealised, horizontally homogeneous forcings (Table 3.2), prescribed as an extra heating rate in CAM4’s radiation scheme. This includes a uniform surface heating (sfc_4) of 4 W m^{-2} , and a vertically and horizontally uniform atmospheric heating (atm_4) whose vertical integral is 4 W m^{-2} . Furthermore, we also apply vertically localised heating anomalies at specific atmospheric levels ϕ (vloc_ ϕ hPa; Fig. 3.1). The applied heating rate anomalies for the vloc_ ϕ hPa forcing experiments were defined as follows:

$$\begin{aligned}
 Q(p) &= A \cos^2 \left(\frac{(p-\phi)\pi}{2a} \right) & \text{for } (\phi - a) \leq p \leq (\phi + a) \\
 Q(p) &= 0 & \text{otherwise}
 \end{aligned} \tag{3.1}$$

for pressure p and heating centred at ϕ , where $A = \pm 0.135 \text{ K day}^{-1}$, $a = 125 \text{ hPa}$, and Q is the instantaneous atmospheric heating anomaly from the control. This provides a vertically integrated heating of 2 W m^{-2} , except for the topmost and lowermost vertically localised experiments which are truncated at the pressure limits (Fig. 3.1) and thus provide around half the vertically integrated forcing. The whole atmosphere was covered by nine of these bounded $\cos^2$ heating profiles centred on multiples of 125 hPa between 0 and 1000 hPa (inclusive). The $\cos^2$ shape combined with the $2a$ heating profile widths

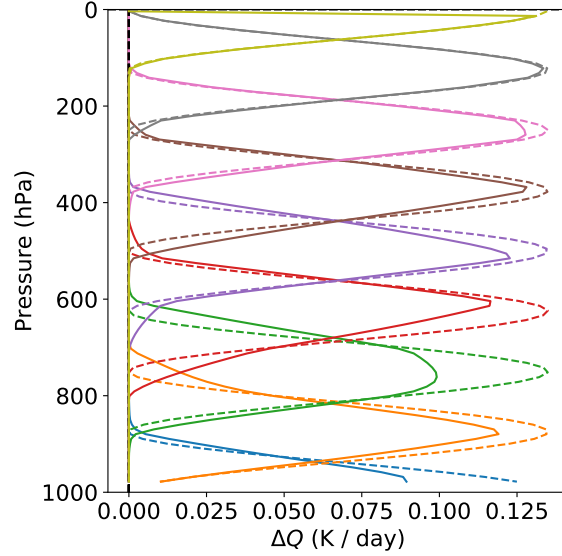


Figure 3.1: Heating profiles of all of the vertically-localized heating experiments in this chapter. Dashed lines show the heating rates defined in Eq. 3.1 of the main text, while the solid lines show the profiles interpolated from the model grid.

mean that the sum of these profiles is uniform in pressure. The width a was chosen such that the forcings are sufficiently localised in the vertical, while still being adequately resolved by CAM4’s vertical grid. To test the linearity of the responses, both positive and negative vertically localised heating tendencies were applied.

Similar approaches of attempting to understand responses to complex spatial forcing by looking at linear combinations of individual localised forcings have appeared several times in the literature (Kuang, 2010; Zhou et al., 2017; Beucler et al., 2018; Dong et al., 2019). Past studies have also investigated energy input at different vertical layers by insertion of aerosol (e.g., Samset & Myhre, 2015; Sand et al., 2020). We use our experiments with various horizontally uniform heating rates to study the effect of the vertical profile of forcing.

Rather than calculating cloud adjustments as differences in cloud-radiative effect, which is the difference between allsky and clearsky radiative fluxes at the TOA, we use cloud kernels from Zelinka et al. (2012a). These are shown in Fig. 3.2a,b, and can be understood as partial derivatives of global TOA radiation (divided into LW and SW) per change in cloud fraction, as functions of cloud top pressure (CTP) and optical depth (τ). Model

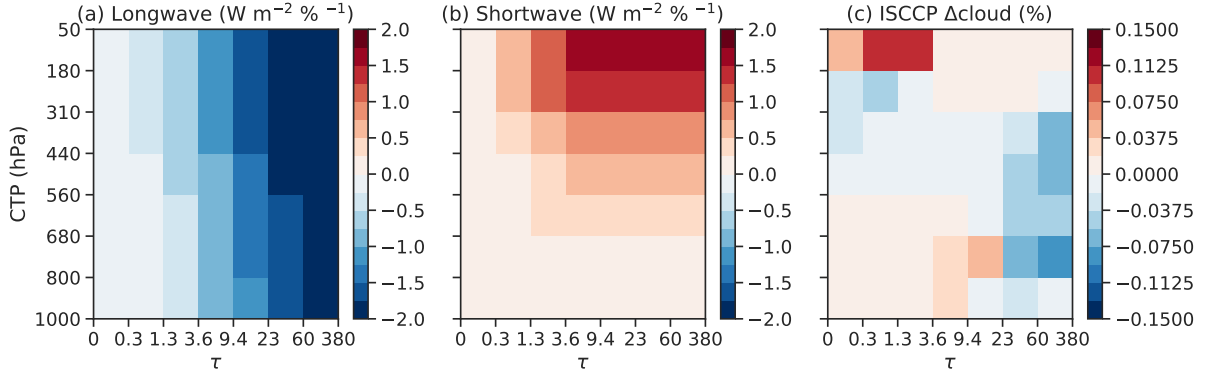


Figure 3.2: (a–b) Globally- (weighted by clear-sky surface albedo) and time-averaged histograms of the cloud kernels obtained from Zelinka et al. (2012a). (c) Globally- and time-averaged histogram of the changes in satellite-observed cloud fraction between the $2\times\text{CO}_2$ and control cases.

differences between the model we use here and those used by Zelinka et al. (2012a) to calculate the kernels could cause discrepancies, but we do not expect issues since Zelinka et al. (2012a) tested other models than those used for calculating the kernels and did not find such. Unlike CRE differences, which are affected by non-cloud adjustments (in temperature, water vapour or surface albedo), the kernels quantify the radiative impact of cloud adjustment in isolation.

The kernels allow the radiative adjustments to be broken down into contributions from high ($\text{CTP} < 440$ hPa), mid ($440 < \text{CTP} < 680$ hPa), and low ($\text{CTP} > 680$ hPa) clouds. To use the cloud kernels, the International Satellite Cloud Climatology Project (ISCCP) satellite simulator (Swales et al., 2018) was enabled in CAM4 to output the required cloud fraction histograms. The ISCCP simulator has known limitations including: low resolution in CTP and τ , diagnosis of cloud property changes only in sunlit conditions (which can affect LW changes), and potential to misdiagnose cloud changes as being caused by higher clouds by assigning CTP to the highest clouds in a column. Nevertheless, Zelinka et al. (2012a) did not find that these issues were likely to negate findings from their kernels analysis.

Using the $2\times\text{CO}_2$ case as an example, we can take the ISCCP simulated cloud changes as a function of CTP and τ (Fig. 3.2c), multiply these by the relevant radiative kernels (Fig. 3.2a,b), and take the sum of products to obtain the global mean LW and SW changes

due to cloud changes.

We introduce two measures in this chapter to help understand the relationship between the vertical profile of atmospheric forcing, tropospheric stability, and cloud-radiative adjustment. Firstly, to characterise the vertical distribution of forcing $Q(p)$ we define a “heating-weighted pressure centroid”:

$$\frac{\int_{p=200 \text{ hPa}}^{1000 \text{ hPa}} p \cdot Q(p) dp}{\int_{p=200 \text{ hPa}}^{1000 \text{ hPa}} Q(p) dp} \quad (3.2)$$

This is understood as the “centre of mass” of the forcing, defined such that larger values here denote a bottom-heavier atmospheric forcing profile. Note that the pressure centroid is positive for the vertically localised heating tendencies of either sign. In general, the pressure centroid is positive and readily interpreted for atmospheric forcings which are entirely or mostly of the same sign at all pressures, as is the case for all those we consider.

Secondly, for tropospheric stability we define the “bulk tropospheric stability” (BTS) as the difference between the average potential temperature (Θ) for the 200–800 hPa layer ($\Theta_{200-800}$), taken as representative of the free troposphere, and $\Theta_{800-1000}$, taken as representative of the boundary layer:

$$\text{BTS} = \Theta_{200-800} - \Theta_{800-1000}. \quad (3.3)$$

We use this measure of atmospheric stability throughout this chapter, instead of lower tropospheric stability measures such as LTS and EIS (equations 2.3 and 2.4), in order to capture stability changes higher up in the free troposphere.

3.3 Vertically localised atmospheric heating experiments

To gain insight into the dependence of cloud adjustments on forcing altitude, we begin with the results from the vertically localised forcing experiments. Focusing on the ver-

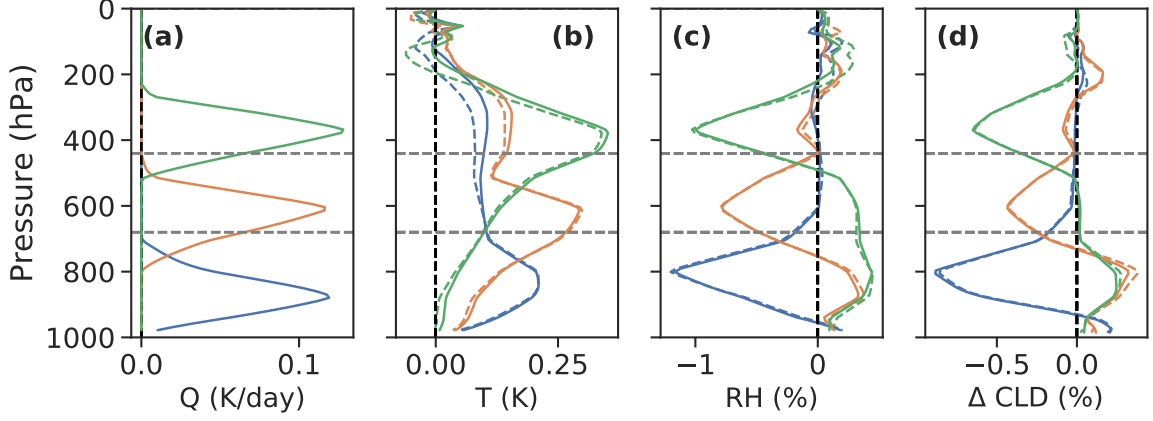


Figure 3.3: Profiles for (a) applied heating rates Q , as well as (b) temperature, (c) relative humidity, and (d) cloud fraction (CLD) change profiles for vertically localised heating experiments with $\phi = 875$ hPa (*blue*), $\phi = 625$ hPa (*orange*), and $\phi = 375$ hPa (*green*). Profiles shown in solid lines are the average of the positive and negative of the negative vertically localised heating tendencies set at the same heights and magnitudes. The profiles for the heating experiments are shown separately as dashed lines. Heating profiles are interpolated from model input, rather than those defined in Eq. 3.1. All values shown are global means. The grey horizontal lines demarcate bounds between high, mid, and low levels according to the ISCCP simulator. Changes to temperature, relative humidity, and cloud are obtained as the difference between the equilibrium fixed SST state and a control state.

tical structure, Fig. 3.3 shows the global-mean adjustments of temperature (T), relative humidity (RH) and cloud fraction for three of the vertically localised forcings (chosen as examples). A close correspondence is found between the peak of applied heating and the peaks of changes to T and RH, as well as cloud fraction. Furthermore, there is a striking similarity between the profiles of changes to RH and cloud fraction, as expected.

Increased temperature without significant specific humidity changes results in RH reductions that are associated with cloud reductions. Additionally, however, the vertically localised forcings also cause non-local changes via changes in stratification and their impacts on vertical heat and moisture fluxes. Heating at lower levels (in the boundary layer) destabilises the overlying free troposphere, leading to enhanced convection and vertical mixing and resulting in warming at higher levels, but little change in RH or cloud (Fig. 3.3, blue curves). By contrast, free-tropospheric heating causes suppressed convection at lower levels through increased tropospheric stability (Fig. 3.4a, open sym-

bols indicate greater positive ΔBTS for vertically localised heating at lower pressure, which means higher altitude), leading to increases to RH and cloud fraction at lower levels (Fig. 3.3, orange and green curves).

In the tropics, low cloud changes occur predominantly in subsidence regions with only small changes in regions of ascent such as the Indo-Pacific warm pool (Fig. 3.5a–b), whereas changes to high cloud are more evenly distributed (panels c–d). Despite locally substantial high cloud changes (Fig. 3.5c), there is no global-average change in high cloud in the low tropospheric forcing scenario (Fig. 3.4d, blue curve). This may be because slight dynamical shifts in the regions of ascent and descent cause local changes which cancel out spatially.

In summary, local cloud fraction decreases in response to localised heating at all levels and associated drying, but low-level cloud increases in response to heating at higher levels. The latter is consistent with the known dependence of low clouds on tropospheric stability (Klein & Hartmann, 1993). Generally, these results are consistent with those from Samset and Myhre (2015, their Fig. 5) and Sand et al. (2020, their Fig. 6) involving application of localised BC at different atmospheric levels.

There is a strong dependence of the cloud-radiative adjustments on the altitude of applied forcing and the associated stability changes. The net cloud adjustments in Fig. 3.4d are of substantial magnitude relative to the imposed vertically localised heating anomalies of 2 W m^{-2} – ranging from about -50% to $+20\%$ of these. This illustrates how cloud adjustments can substantially enhance or offset the instantaneous forcing, depending on the vertical distribution of anomalous heating. The cloud-radiative adjustment becomes monotonically more negative with the height of the applied heating (Fig. 3.4c–d). This trend is driven by both LW and SW changes, with the latter dominating, qualitatively consistent with previous findings for BC forcing applied at different altitudes (Samset & Myhre, 2015, their Fig. 1; Sand et al., 2020, their Fig. 2). SW cloud adjustments flip sign from positive to negative as forcing altitude increases, while LW cloud adjustments become strongly negative (Fig. 3.4d). The dependence of LW cloud adjustment on forcing

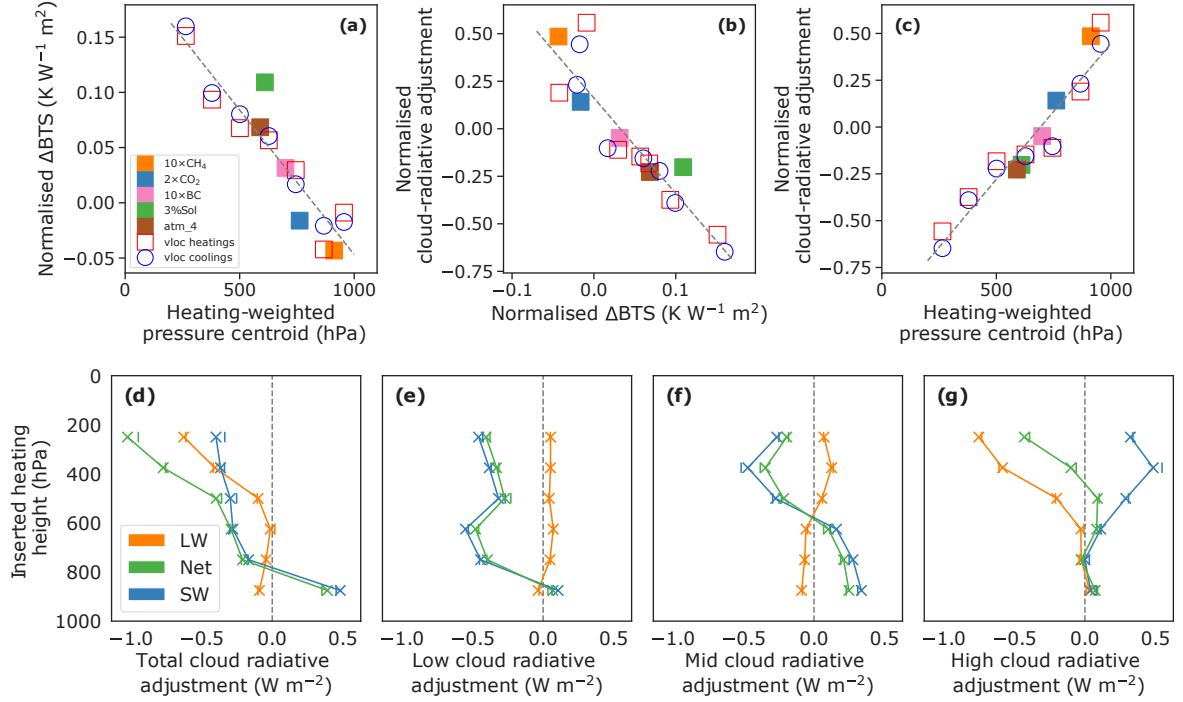


Figure 3.4: (a) Changes to bulk tropospheric stability (BTS, see Eq. 3.3), normalised by the vertically integrated tropospheric (200–1000 hPa) heating anomalies, versus the vertical centre of mass of tropospheric heating (Eq. 3.2) for each experiment performed in this study. (b) Normalised cloud-radiative adjustments against normalised BTS changes. (c) Normalised cloud-radiative adjustments versus heating centre of mass. Least-squares linear fits are shown by the dashed lines. Only experiments with significant tropospheric forcing are shown, which excludes the 0 hPa and 125 hPa forcing experiments. (d) Cloud-radiative adjustments from the six vertically localised forcings centered between 875 and 250 hPa, in 125-hPa increments, broken down into contributions from low (e), mid-level (f), and high (g) cloud changes. The lowest altitude forcing at 1000 hPa is excluded because it is truncated; furthermore, the two highest altitude forcings are excluded because they are mainly in the stratosphere, and have little impact on clouds. Crosses represent the averages between the responses to heating anomalies and the negative of the responses to cooling anomalies at the same altitudes and of the same magnitudes, with the thin horizontal lines around the crosses representing the results from the heating experiments.

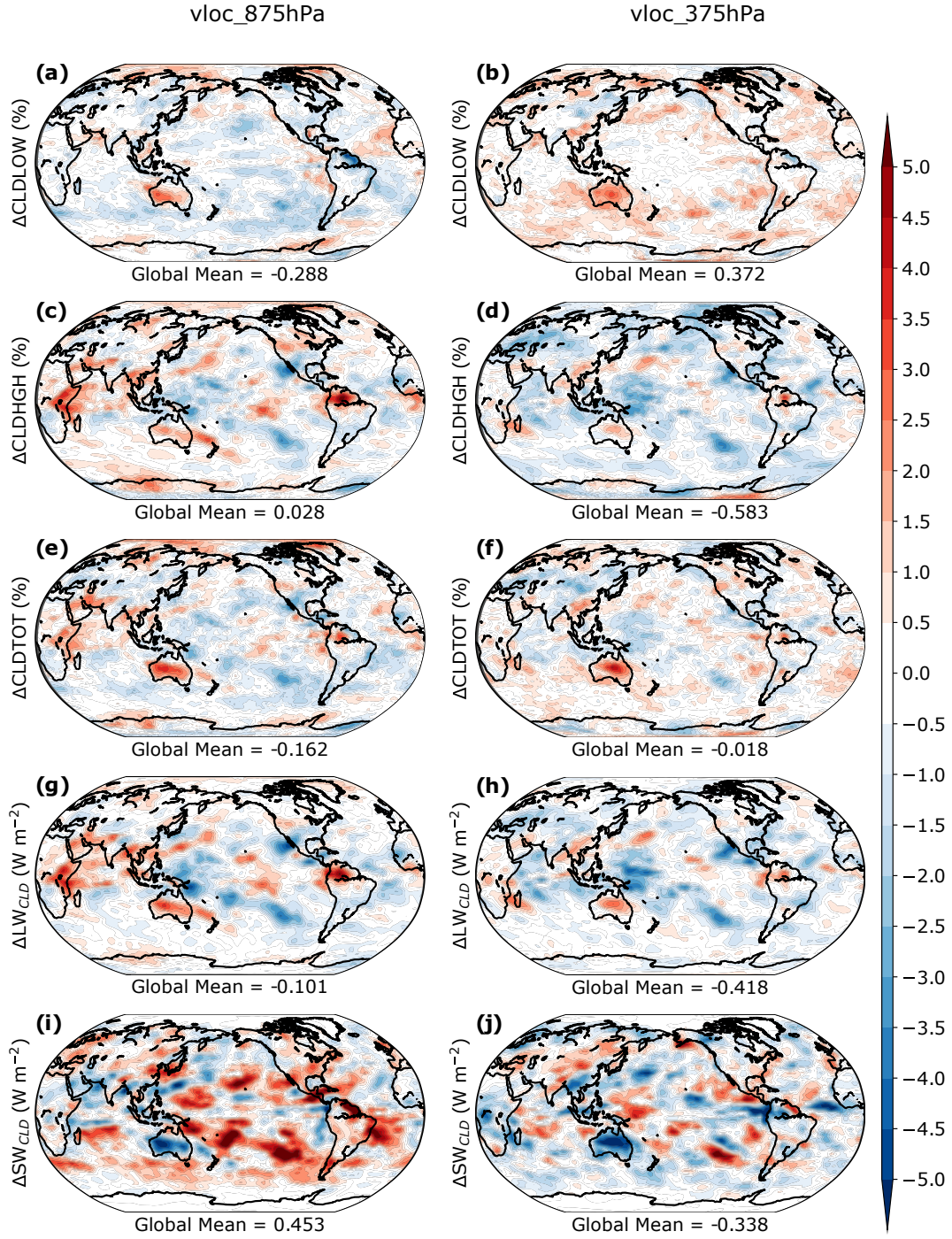


Figure 3.5: Maps of adjustments to low cloud (a–b), high cloud (c–d), and total cloud (e–f) fractions, and kernel-derived longwave (g–h) and shortwave (i–j) cloud radiative adjustments for the experiments involving localised atmospheric heating at 875 hPa (*left*) and 375 hPa (*right*).

altitude is consistent with the understanding that LW cloud-radiative effect increases with cloud altitude (Hartmann, 2016), due to increased temperature differences between higher clouds and the surface.

That mid-level cloud-radiative adjustments are the main contribution to total cloud-radiative adjustments for the lowest level of localised atmospheric heating is surprising, and seems to disagree with the cloud changes being focused below the mid cloud threshold (Fig. 3.3). This could be a sign of the limitations of the ISCCP simulator, with low-level cloud changes assigned to mid-level cloud changes as part of the issue mentioned previously where cloud changes are assigned to higher levels as only the top CTP in the column is considered (Zelinka et al., 2012a).

To interpret the dependence of SW cloud adjustment on forcing altitude (Fig. 3.4d), it is helpful to consider the breakdown of the adjustments into contributions by high-, mid- and low-level clouds (Fig. 3.4e,f,g). Low- and medium-cloud radiative adjustments tend to be negative below the level of the heating, and positive above, whilst the high-cloud radiative adjustment is negative for the highest altitude heating. All levels tend to give a negative cloud radiative adjustment below the level of heating, while mid-level cloud contributes positively above the level of heating as well. This is consistent with the findings from Fig. 3.3: localised heating causes a cloud fraction reduction locally, but a cloud fraction increase below (particularly in the boundary layer), associated with stabilisation (Fig. 3.4, open symbols) and suppressed vertical mixing. There is an additional factor contributing to the negative SW cloud adjustment below the vertically localised heating: when cloud fraction decreases at the heating level, the reduced overlap reveals and hence increases SW reflection from lower-level clouds.

Geographically, the radiative changes mirror the high cloud changes in the LW (Fig. 3.5g–h), spatially cancelling out for the boundary layer forcing case. By contrast, the SW changes (Fig. 3.5i–j) are more closely related to the changes in total cloud amount (Fig. 3.5e–f) and cannot be easily attributed to a particular region or cloud regime.

We note that in Fig. 3.3b–d, the corresponding dashed and solid lines are similar, and

in Fig. 3.4d-g the corresponding crosses and vertical lines are close together. Differences are minor even in the upper troposphere, where the differences are largest. Both thus provide evidence that positive and negative heating tendencies of the same magnitude produce responses of the same magnitude and opposite sign.

Cloud adjustments are caused also by heating at the surface, but the `sfc_4` experiment shows that they are very small, at all altitudes and in both LW and SW (not shown), with a net CRE adjustment of -0.03 W m^{-2} . This is negligible in comparison to the adjustments for the localised atmospheric heating experiments, especially per unit of forcing. This result is expected given the fixed-SST lower boundary, where surface forcing can impact the atmosphere only over land and ice regions. Note that although the rapid climate response to land warming under fixed SSTs is typically included in the ERF as part of the rapid adjustment (Forster et al., 2016; Sherwood et al., 2015), conceptually this can also be treated as a surface warming-driven radiative response (Chung & Soden, 2015; Shine et al., 2003).

3.4 Interpreting cloud-radiative adjustments to common forcing agents

This section seeks to explain adjustments to some common forcing agents (Table 3.1) in light of the findings of Section 3.3 concerning vertically localised heating. For each of the common forcings (as well as the idealised `atm_4` case) we express the vertical profile of global-mean heating as a linear combination of idealised vertically localised heating perturbations:

$$Q_{\text{fit}}(p) = \sum_{i=1}^9 (a_i \cdot Q_i(p)), \quad (3.4)$$

where Q_i and a_i are the anomalous heating rates and fitting coefficients respectively for each of the nine vertically localised forcings i . The a_i coefficients were calculated so as to minimise the least-squares difference between Q_{fit} and the global-mean heating profile

of a chosen case, Q_{case} (Fig. 3.6, left column). Best-fits are expected to be unique given that the profiles of the vertically localised forcings are mostly non-overlapping and hence mostly orthogonal.

The heating profiles for the common forcings (and atm_4) are very closely approximated by linearly combining the nine idealised vertically localised cases (Fig. 3.6, left column). The only deviations occur above the tropopause for the $2\times\text{CO}_2$ and 3%Sol cases where there is insufficient vertical resolution in the localised forcing experiments.

Using the same a_i coefficients obtained in equation 3.4, we then estimate the globally-averaged vertical profiles of change in temperature, relative humidity, and cloud fraction (each denoted by X) in response to each vertically distributed forcing, whose instantaneous surface forcing is F_{sfc} thus:

$$X_{\text{fit}}(p) = \left(\frac{F_{\text{sfc}}}{4 \text{ W m}^{-2}} \cdot X_{\text{sfc_4}} \right) + \sum_i (a_i \cdot X_i(p)). \quad (3.5)$$

The first term on the right-hand side accounts for contributions from surface forcing, using the results from a 4 W m^{-2} uniform surface heating experiment (sfc_4, see Table 3.2), appropriately weighted.

The linear combination of idealised vertically localised heating anomalies closely approximates the adjustments of temperature, RH and cloud fraction diagnosed from the model (Fig. 3.6). This suggests that the global-average vertical structure of these adjustments is primarily determined by the vertical profile of instantaneous heating anomalies.

The contributions to these profiles from the surface components of the forcings were found to be minor in general, consistent with our finding that the cloud adjustment to uniform surface-only forcing is very small. The notable exception to this was in the solar forcing case, where the majority of instantaneous forcing is from the surface component (4.88 W m^{-2} , compared to a 2.19 W m^{-2} atmospheric component), such that the surface component makes a non-negligible contribution to the cloud fraction change profile (Fig. 3.7).

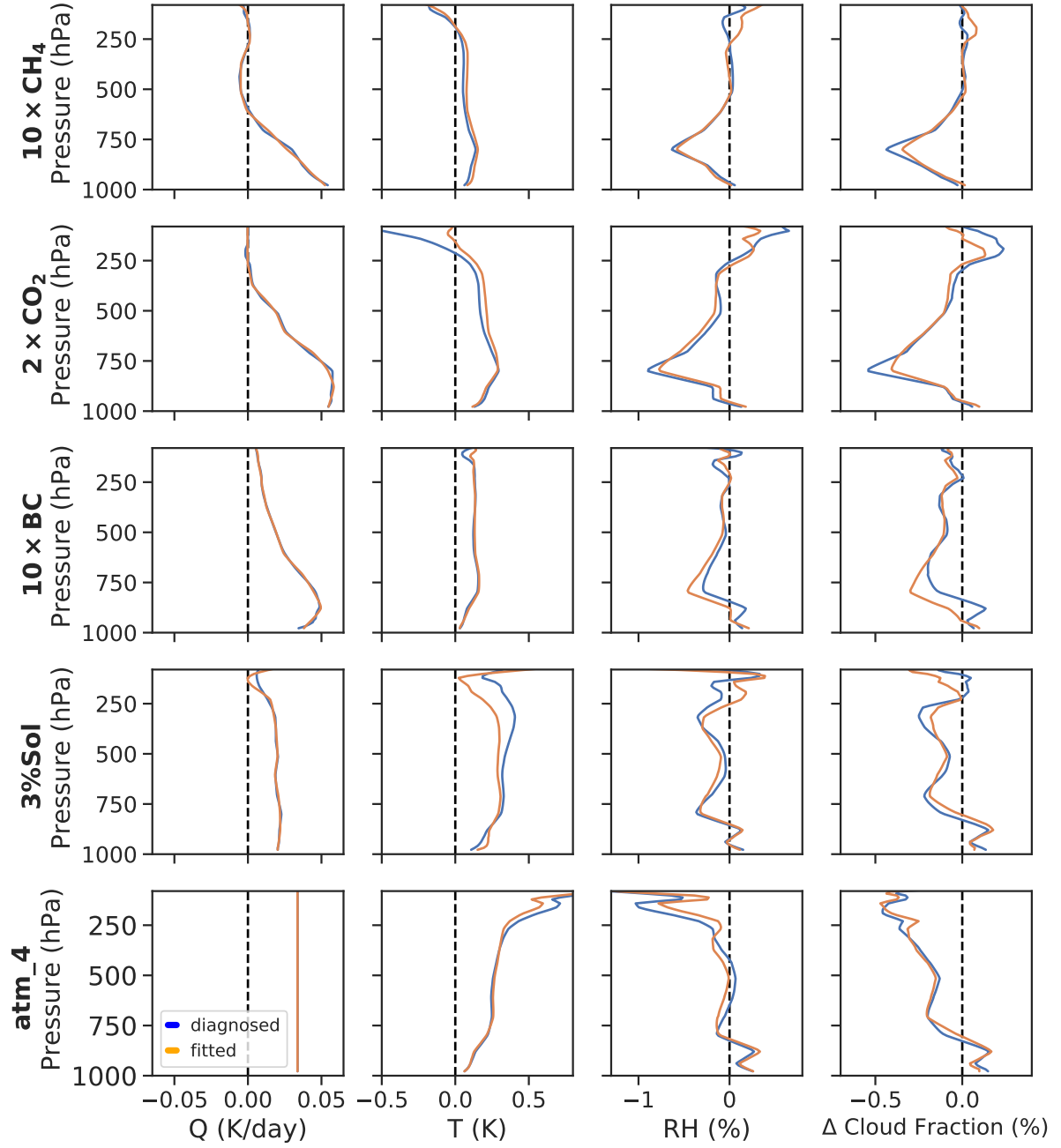


Figure 3.6: Global-mean vertical heating anomaly profiles (Q) and rapid adjustments of temperature (T), relative humidity (RH) and cloud fraction. Shown are the results from the original cases (*blue*) and linear combinations of the results from the vertically localised forcing experiments (*orange*) including an appropriate surface term (Eq. 3.5). Fitted heating rates are best fits to the original heating rates (Eq. 3.4) in all cases (except for $2\times\text{CO}_2$ where we set the stratospheric heating anomalies to zero before fitting). The orange lines for other variables are created from the fits of heating rates (Eq. 3.5).

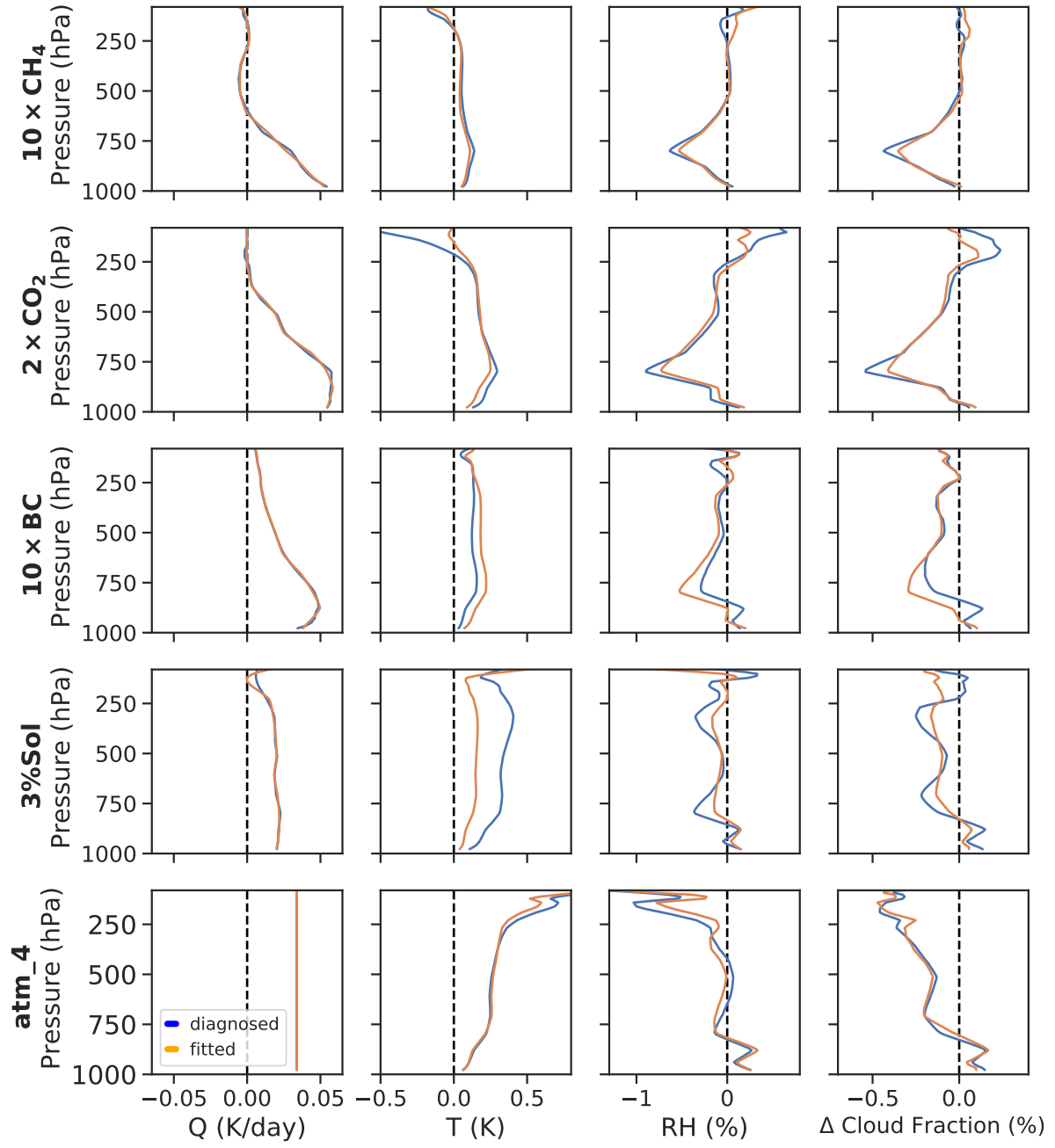


Figure 3.7: The same as Fig. 3.6, but excluding contributions from surface forcings. Note that only the 3%Sol case changes significantly.

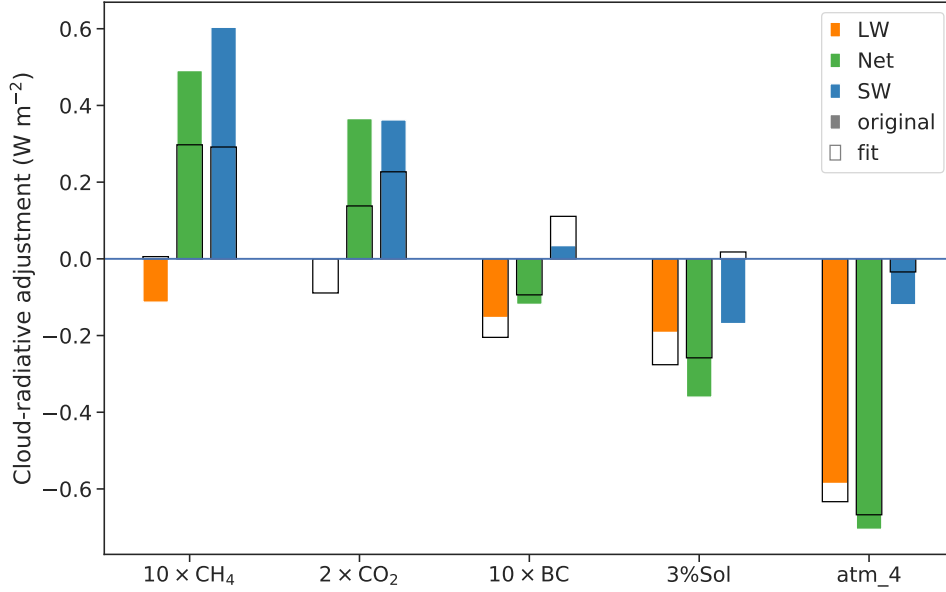


Figure 3.8: A comparison of the cloud-radiative adjustments predicted from linearly combining vertically localised forcing experiments (*empty bars*) versus the radiative adjustments calculated from the relevant experiments themselves (*solid bars*).

Considering the global average top-of-atmosphere cloud-radiative adjustments to forcings, we find that the linear combinations of vertically localised heating experiments generally predict the correct sign, and roughly the magnitude, of the vertically distributed forcings (Fig. 3.8). The cloud radiative adjustments to $10\times\text{BC}$ are surprisingly well predicted, despite the cloud responses here being the least accurately predicted (although still captured to first order, Fig. 3.6). This is possibly due to errors in the predicted cloud response fortuitously cancelling out in some way (e.g., errors in the vertical may compensate for errors in the horizontal distribution). The largest errors are for CO_2 and CH_4 , where the positive SW adjustments are considerably underestimated. Inspection of the cloud fraction profiles in Fig. 3.6 suggests this may partly result from an underestimation of the lower-tropospheric cloud fraction decrease by the simple linear combination method, and potentially from differences in estimation of cloud changes near the tropopause. That the predicted net adjustments are more accurate than the individual SW and LW adjustments can be explained by compensating errors in SW and LW. Errors in this approach may also result from the linear combination of cloud-radiative adjustments being unable to account for the non-linear effects of cloud overlap.

The results in Figs. 3.6 and 3.8 account for the finding of Fig. 3.4c (filled symbols) that the sign and magnitude of cloud-radiative adjustments are mostly explained by the vertical structure of the instantaneous atmospheric forcing, as measured by the heating-weighted pressure centroid. Positive CRE adjustments result from the “bottom-heaviest” instantaneous atmospheric heating profiles of $2\times\text{CO}_2$ and $10\times\text{CH}_4$, negative CRE adjustments from 3%Sol and atm_4, whose instantaneous heating profiles are fairly uniform with altitude, while $10\times\text{BC}$ is intermediate.

Although our interpretation is based on a single climate model, we note that the cloud-radiative adjustments in Fig. 3.8 are reasonably representative of those simulated by a range of CMIP5 models (Fig. 4 of Smith et al., 2018). In particular, models consistently simulate positive cloud-radiative adjustments to CO_2 , and negative adjustments to solar forcing, in agreement with our results. Note that the results of Smith et al. (2018) use different magnitudes of the CH_4 and solar forcings, and include the stomatal conductance effect of $2\times\text{CO}_2$. Increases to CO_2 lead to reduced evapotranspiration and thus reduced low cloud over many highly forested areas for a positive effect on the cloud-radiative adjustment (Doutriaux-Boucher et al., 2009). We found that including this effect approximately doubles the cloud adjustment to $2\times\text{CO}_2$ forcing in CAM4 (0.76 vs 0.37 W m^{-2} ; not shown).

3.5 Summary and conclusions

We have demonstrated through a series of idealised experiments with vertically localised atmospheric heating that cloud-radiative adjustments are sensitive principally to the altitude of atmospheric heating caused instantaneously by the forcing agent (Fig. 3.4c). At levels where there is an instantaneous positive heating tendency, the air becomes warmer and drier and the cloud fraction decreases. However, heating at any level above the boundary layer stabilises the troposphere below and suppresses vertical mixing, thus causing moistening and increased cloud fraction at lower levels. As the net result of

these two effects, lower-tropospheric heating results in positive cloud-radiative adjustment (dominated by the SW effect of reduced low cloud fraction), while mid- and upper-tropospheric heating causes negative adjustment (due to the combined LW and SW effects of increased low cloud and reduced free-tropospheric cloud). For negative forcings, the signs of all effects are reversed and the magnitudes are similar.

We find that the global-mean net cloud-radiative adjustments to common forcings can be partially explained by linearly combining the idealised vertically localised forcings so as to fit the vertical heating profiles caused by the various forcing agents. In particular, our results suggest that positive cloud adjustment commonly found in GCMs to the greenhouse gas forcing agents CO_2 and CH_4 is explained by their relatively bottom-heavy atmospheric heating profiles; by contrast, the negative cloud adjustment to solar forcing is caused by its relatively larger free-tropospheric heating. Our findings are consistent also with previous evidence that cloud-radiative adjustments depend on the altitude of absorbing aerosols such as black carbon (Samset & Myhre, 2015; Allen et al., 2019).

Although successful in explaining the sign and relative magnitude of adjustments for different agents, our method using global-mean vertical profiles does not give quantitatively accurate estimates. This may be for instance because of neglecting the geographical pattern and the seasonal cycle of the instantaneous heating, perhaps especially the contrast between its effects in cloudy and cloud-free air. Further investigation is needed of these aspects. With such refinement, we expect that this approach will provide a useful basis to interpret inter-model differences in cloud adjustments, to the extent that such differences result from uncertainties in the distribution of instantaneous radiative forcing.

The data used in this chapter can be found at <https://doi.org/10.6084/m9.figshare.14386799>

Chapter 4

THE DEPENDENCE OF RADIATIVE FEEDBACKS ON FORCING AGENT¹

From looking at rapid radiative adjustments, we move onto radiative feedbacks i.e. the radiative response driven by surface warming/cooling. In this chapter, we use experiments from seven CMIP6 models to assess the climate feedback parameter for net historical, historical GHG and AA forcings. The net radiative feedback is found to be more amplifying (higher effective climate sensitivity) for AA than GHG forcing. Owing to opposite forcing signs of GHG (positive) and AA (negative), feedbacks are thus less amplifying for net historical (GHG + AA) than GHG only. We demonstrate that this difference is consistent with their different latitudinal distributions. Historical AA forcing is most pronounced in northern extratropics, where the boundary layer is decoupled from the free troposphere, so the consequent temperature change is confined to low altitude and causes low-level cloud changes. Thus overall shortwave feedbacks are affected, through change in stability which also affects upper-tropospheric clearsky emission, affecting longwave radiative feedbacks also. This response is a feature of extratropical forcing generally, regardless of its sign or hemisphere.

¹This chapter is largely a recreation of published work (doi.org/10.1029/2022gl097766)

4.1 Introduction

Global warming due to future emissions of greenhouse gases and other climate forcing agents has been understood in recent years in terms of the energy imbalance that these forcings cause in the Earth system, to which this system responds through changing surface temperatures. The usual interpretive model is $N = F - \alpha T$, where α is the climate feedback parameter, N top-of-atmosphere energy imbalance, T global mean surface temperature change and F effective radiative forcing (Ramaswamy et al., 2001). (Note: we have chosen the sign convention where a positive α implies an increased positive-upwards radiative response for increasing surface temperatures.) Although this is helpful and intuitive as a simple model, there are several complications when using it (Andrews et al., 2015; Knutti & Rugenstein, 2015; Sherwood et al., 2015).

While there are generally confirmed differences in the feedback parameter across models (Andrews, Gregory, Webb, & Taylor, 2012; Becker & Wing, 2020; Zelinka et al., 2020), there is not a consensus on whether α depends on the different forcing agents that are relevant to past and future temperature changes. Comparing AAs and GHGs, the two dominant historical forcing agents (Smith et al., 2020), several studies have found differences in feedbacks (Marvel et al., 2016; Shindell, 2014; Smith & Forster, 2021). Gregory et al. (2020) presented evidence for a difference in α between anthropogenic forcing (GHGs and AAs) and natural forcing by volcanic aerosol. By contrast, Richardson et al. (2019) did not find significant differences among forcing agents, considering several models in the Precipitation Driver Model Intercomparison Project (PDRMIP) experiments. The focus of this study is the dependence of α on the nature of the forcing agent.

In the recently released data of the Coupled Model Intercomparison Project phase 6 (CMIP6), historical single-forcing experiments across several models have become available for analysis. These new experiments allow us to obtain the effective forcings for different agents, allowing us to accurately calculate the corresponding radiative feedbacks over the historical period. Previous work has found a dependence of these feedbacks on

SST patterns (Andrews et al., 2015; Ceppi & Gregory, 2019; Zhou et al., 2017), which affect the radiation budget via changes in stability and clouds (Andrews et al., 2018). Following this reasoning, this work considers how historical GHG and AA forcing affect SSTs and stability, and how these in turn modulate the radiative responses.

4.2 Data and methods

Data for several historical forcing experiments was obtained from the ESGF CEDA archive (`esgf-index1.ceda.ac.uk`) for seven models: CanESM5, CNRM-CM6-1, GISS-E2-1-G, HadGEM3-GC31-LL, IPSL-CM6A-LR, MIROC6, and NorESM2-LM. These models cover a range of climate sensitivities (2.56-5.64 K) with a mean of 4.07K that is close to the average of 3.86K across CMIP6 models (Zelinka et al., 2020). Furthermore, these models have been found to be relatively independent based on climatology biases (Brunner et al., 2020). These experiments include a control (piControl) with constant pre-industrial forcing agents, as well as experiments both with coupled atmosphere-ocean models (AOGCMs) and with atmosphere models (AGCMs) given prescribed sea surface conditions, for historical GHGs, AAs, and all historical forcings together.

The experiments, models and variants used are detailed in Table 4.1. The historical single-forcing AOGCM experiments hist-aer and hist-GHG were performed for the Detection and Attribution Model Intercomparison Project (DAMIP; Gillett et al., 2016), while the AGCM piClim simulations are part of the Radiative Forcing Model Intercomparison Project (RFMIP; Pincus et al., 2016). On a model by model basis, for each experiment, we sought multiple ensemble members where only the realisation indices (r labels) were different, which indicates only differences in initial conditions. Where possible, we chose variants with the same initialisation method (i1 variant label) such that initialisation procedures were generally consistent. Model physics (p label) were chosen to be consistent with piControl so as to avoid differences in parameterisations. Similarly, the forcing indices (f labels) were chosen to be consistent with piControl to avoid using different

Table 4.1: Experiment names and details for the CMIP6 experiments used in this chapter. The “piClim” prefix denotes fixed-SST experiments, while the rest are coupled atmosphere-ocean experiments. piClim-control is an AGCM experiment with SSTs and sea ice climatologies taken from piControl. The fourth column includes three prescribed-SST experiments: piClim-histall, piClim-histGHG and piClim-histaer.

	piClim-control	hist-GHG & hist-aer	historical	piClim-histxxx
CanESM5	r1i1p2f1	r[1-10]i1p2f1	r[1-10]i1p2f1	r1i1p2f1
CNRM-CM6-1	r1i1p1f2	r[1-10]i1p1f2	r[1-10]i1p1f2	r1i1p1f2
GISS-E2-1-G	r1i1p1f1	r[1-5]i1p1f1	r[1-10]i1p1f1	r1i1p1f1
HadGEM3-GC31-LL	r1i1p1f3	r[1-4]i1p1f3	r[1-5]i1p1f3	r[1-3]i1p1f3
IPSL-CM6A-LR	r1i1p1f1	r[1-10]i1p1f1	r[1-10]i1p1f1	r1i1p1f1
MIROC6	r1i1p1f1	r[1-3]i1p1f1	r[1-10]i1p1f1	r[1-3]i1p1f1
NorESM2-LM	r1i1p1f1	r[1-3]i1p1f1	r[1-3]i1p1f1	r[1-3]i1p1f1

forcing datasets per model (K. E. Taylor et al., 2018). We included all realisations that contained all of the required variables, and our results from each experiment of each model are ensemble averages. We calculated multi-model mean (MMM) values from the individual model ensemble averages, with equal weighting for each model.

We analyse the upward TOA radiative response R , the stability response S as measured by the estimated inversion strength (EIS; Wood & Bretherton, 2006), and the cloud-radiative effect (CRE) measured as the difference between allsky and clearsky downward fluxes at the TOA. We choose EIS as our measure of atmospheric stability in this chapter following related work (Ceppi & Gregory, 2019). Prior to analysis, all fields were regridded to a common T42 grid (corresponding to a grid resolution of approximately 2.8° in longitude and latitude), using conservative remapping for radiative fluxes, and bilinear interpolation for other fields. Monthly fields were aggregated into annual averages.

In order to separate out the surface warming (or cooling) driven feedbacks from the forcing and associated rapid adjustments (Hansen et al., 1997; Sherwood et al., 2015), we use the results from the AGCM experiments with fixed sea surface temperatures through the following equation:

$$X_{\text{response}} = X_{\text{AOGCM}} - X_{\text{AGCM}}, \quad (4.1)$$

where X_{response} represents the response of the variable of interest X . Radiative feedbacks and other responses per unit global warming, such as dS/dT , were calculated through linear least-squares regression of the desired variable against globally-averaged surface air temperature. The regressions are performed across the full length of the historical period (years 1850–2015) for all experiments, despite the possible time-dependence of α (Gregory & Andrews, 2016; Gregory et al., 2020) which we do not study in this chapter. Such time variations may be important for the evolution of historical climate change, and evolving historical feedbacks may vary significantly from the value calculated across the period, but we limit ourselves here to analysing the overall effect of different forcing agents across the historical period. Note that the impact of short-lived volcanic forcing should be minimal on the historical feedbacks calculated over this long time period.

Global averages in this chapter refer to area-weighted averages of the variables of interest, except for stability S where the average is taken over ocean regions 50°S–50°N. This choice follows Ceppi and Gregory (2019), who found EIS equatorward of 50° to be well-correlated with the global radiative response R in observations and in two GCMs. We expect stability to be related to low cloud cover over oceans, from previous work that has found such (Wood & Bretherton, 2006), but we omit polar regions since icy regions can have significant stability changes that are not reflected by cloud changes. Since stability correlates with low cloud cover, and hence leads to a more negative shortwave (SW) cloud response from enhanced low cloud amount and SW reflection, we expect that this near-global measure of stability over oceans should relate to the radiative response.

The confidence intervals for results derived from regressions combine two aspects of uncertainty. The first is the variability among different ensemble members, which we have calculated for each model as the variance across members of the historical experiment. This experiment was chosen since it generally had the most ensemble members, with the assumption that the magnitude of unforced variability is representative of other forcing scenarios. The second is the estimated error in the regression slope of the ensemble-averaged data. The combined error is calculated as the square root of the sum of variances

from these two aspects of uncertainty, and given as $\pm 1\sigma$ of the probability distribution.

Two idealised extratropical forcing experiments were run to investigate the impact of forcing localised away from the regions of deep convection on radiative feedbacks and tropospheric stability. These experiments are denoted as nh_extrop and sh_extrop for northern (30°N–90°N) and southern (30°S–90°S) hemisphere forcing respectively. A uniform “ghost” radiative forcing of -4.5 W m^{-2} was imposed as an extra term in the surface energy budget in the relevant regions for each experiment. The imposed instantaneous radiative flux values were chosen so that the global average forcing would be comparable to that of hist-aer over years 1995–2015 (around -1.13 W m^{-2}). Though the nh_extrop forcing is zonally uniform unlike the forcing for hist-GHG and hist-aer (Fig. 4.1a-b), the nh_extrop setup was chosen to capture the skew of the northern extratropical forcing that is seen in hist-aer relative to the more homogeneous hist-GHG (Fig. 4.1c). The idealised experiments were performed with the Hadley Centre Atmospheric Model version 3 (Pope et al., 2000) in both atmosphere-only (HadAM3) and slab ocean (HadSM3) configurations. The horizontal resolution is $2.5^\circ \times 3.75^\circ$; there are 19 levels in the atmosphere and the slab ocean has a thickness of 50 m. This model is one of the two used by Ceppi and Gregory (2019), and these experiments are the same as their $-\text{UNIF}_{\text{ET}}$ experiment for a single hemisphere at a time, except with a different forcing magnitude.

We also make use of tropical-only forcing experiments in HadSM3 and HadAM3 denoted as “tropical” in later figures. This is the same as the $+\text{UNIF}_{\text{T}}$ experiment in Ceppi and Gregory (2019), which involves a uniform forcing of 7 W m^{-2} over the tropics (30°S–30°N). Model outputs from these experiments can be found at <https://doi.org/10.6084/m9.figshare.17197748> and <https://doi.org/10.6084/m9.figshare.17182907>

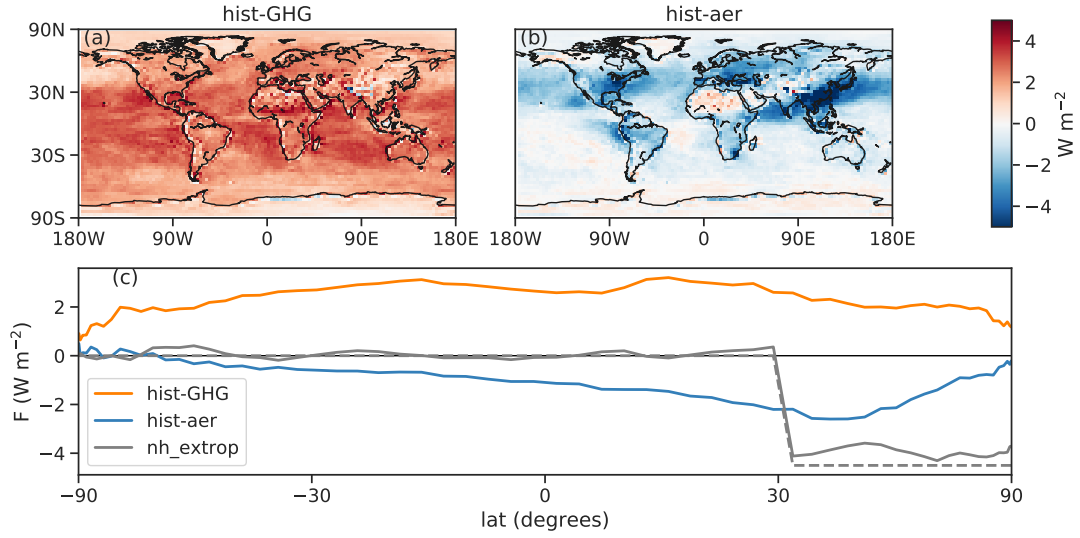


Figure 4.1: Top-of-atmosphere radiative forcing patterns over years 1995–2015 for hist-GHG (a) and hist-aer (b). The zonal-mean profiles of forcing are shown in (c), where values for hist-GHG (*orange*) and hist-aer (*blue*) are the average effective forcings of the relevant AGCM experiments (piClim-histghg and piClim-histaer, respectively) minus control. For nh_extrop (*grey*) the dashed line shows the prescribed instantaneous forcing whilst the solid line shows the effective forcing calculated from the HadAM3 (atmosphere-only) simulation. Note that the x -axis is scaled by geographical area.

4.3 Investigating radiative feedback differences in terms of stability differences

The radiative feedbacks for historical AA and all historical forcings, relative to those from GHGs, are shown in Fig. 4.2a, for each model analysed and for the MMM. Coloured bars show the allsky net feedback parameter α , whilst markers show the CRE feedback parameter α_{CRE} (minuses) and clearsky feedback parameter α_{CS} (pluses). α_{CRE} and α_{CS} are obtained as regressions in the same way as α but, where $\alpha = dR/dT$, $\alpha_{\text{CRE}} = dCRE/dT$ and $\alpha_{\text{CS}} = dR_{\text{CS}}/dT$. In the same way as $R = R_{\text{CS}} + \text{CRE}$, then, $\alpha = \alpha_{\text{CS}} + \alpha_{\text{CRE}}$. In all cases, positive numbers mean less amplifying feedbacks, i.e. a relatively larger upward radiative flux perturbation for positive T , which leads to lower climate sensitivity. The findings here show more amplifying feedbacks for hist-aer than hist-GHG in the MMM. On a model-by-model basis, hist-aer shows either significantly more

amplifying or very similar feedbacks to hist-GHG (except for CNRM-CM6-1 which, for unknown reasons, is the only model with a substantially less amplifying feedback for hist-aer).

Neither α_{CS} (correlation of 0.80 with allsky α , Fig. 4.3b) nor α_{CRE} (correlation of 0.85 with allsky α , Fig. 4.3d) entirely explains the differences of allsky α for AA (or the net historical forcing) compared to GHG forcing. Both correlations are highly significant ($p < 0.001$), which might indicate that both clearsky and CRE feedbacks are important in explaining the spread in α as neither α_{CS} nor α_{CRE} dominates the spread. However, the NorESM2-LM model seems to be responsible for much of the correlation seen here. Ignoring this model (not shown) preserves good correlation (0.79 for allsky α with α_{CS}) at high significance ($p < 0.001$), but greatly diminishes the correlation of allsky α with α_{CRE} (0.15, with low significance of $p = 0.527$). As such, clearsky differences in feedbacks seem to be dominating feedback differences between AA and GHG forcing.

We propose that differences in radiative feedback across forcing agents may be explained in terms of different tropospheric stability responses and their impact on cloud and lapse-rate feedbacks. Lapse-rate feedbacks result as decreases/increases in the lapse rate give a weaker/stronger greenhouse effect. This is tied to temperature change being different at the surface than higher up in the atmosphere, with a smaller ratio for decreased lapse rates (hence weaker greenhouse effect) and larger ratio for increased lapse rates (hence stronger greenhouse effect). This feedback can largely be cancelled out by water vapour feedbacks, which involves greater greenhouse effects from the warming of air as saturation water vapour increases, but the difference between lapse rate and water vapour feedbacks can still have a significant net effect on feedbacks (Andrews & Webb, 2018; Armour et al., 2013; Ceppi & Gregory, 2019; Po-Chedley et al., 2018).

Looking at Fig. 4.2b, we see that hist-aer causes smaller changes in stability per unit surface temperature change (dS/dT) than hist-GHG across all models (except for CNRM-CM6-1). That is, we see greater stability change per unit of GHG forcing than per unit of AA forcing. With EIS (equation 2.4), which measures the difference between potential

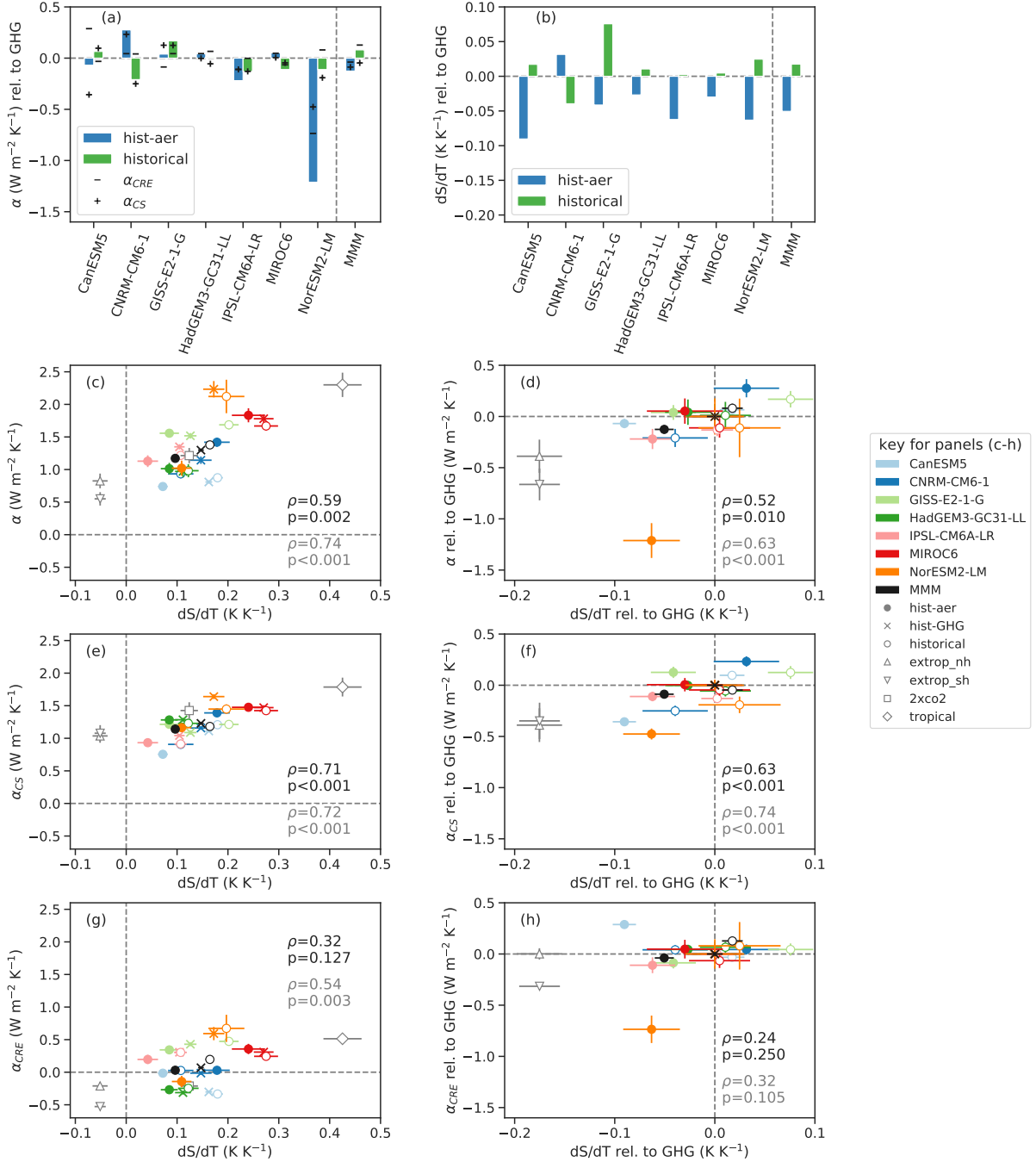


Figure 4.2: (a) allsky radiative feedback parameter (α , *bars*) alongside CRE (α_{CRE} , *minuses*) and clearsky radiative feedback parameters (α_{CS} , *pluses*) for each model and the multi-model mean, as the difference from hist-GHG values. (b) Difference of dS/dT from hist-GHG values. (c,e,g) Net (c) clearsky, (e) CRE, and (g) allsky radiative feedback parameters against net dS/dT . (d,f,h) as row above, except with values relative to hist-GHG. Confidence intervals in (c-h) denote $\pm$ one standard deviation based off combined regression and ensemble member uncertainties. The Pearson correlation ρ is shown in (c-h), across models and both excluding (*black*) and including (*grey*) the data points from the HadSM3 experiments. The tropical-only forcing experiment is included in (c,e,g), as part of the inter-model trend, but excluded (both in plotting and ρ) in (d,f,h) for clarity and to focus on the HadSM3 experiments that are more similar to hist-aer and hist-GHG. Also shown are p -values for the statistical significance of the correlations.

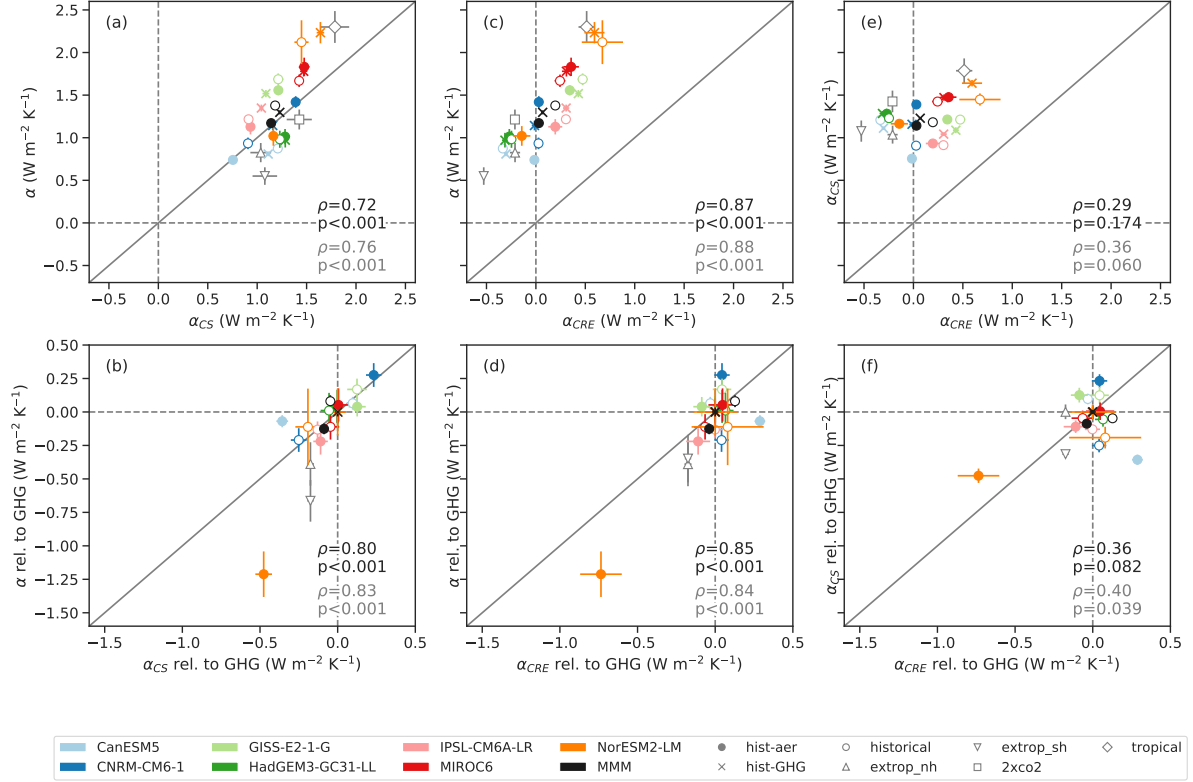


Figure 4.3: A similar plot as Fig. 4.2, comparing the all-sky radiative feedback parameter (α) to CRE (α_{CRE}) and clear-sky radiative feedback parameters (α_{CS}) for each model and the multi-model mean. Values are shown as absolutes (*top row*) and as the difference from hist-GHG values (*bottom row*). All panels contain a 1:1 line (*solid grey*)

temperatures at the surface and at 700hPa (with a moist adiabatic correction), as our metric of stability: a greater increase in stability (as in GHG compared with AA) suggests more warming in the troposphere than at the surface. Hence this suggests a positive (less amplifying) lapse-rate feedback (Andrews & Webb, 2018; Armour et al., 2013; Ceppi & Gregory, 2019), and an additional minor contribution from relative humidity changes (Fig. 6 in Ceppi & Gregory, 2019). Furthermore, the correlation of low cloud amount with stability (Zhou et al., 2016; Ceppi et al., 2017; Andrews & Webb, 2018; Ceppi & Gregory, 2019) and corresponding SW reflection (Hartmann, 2016) suggests that increased stability counteracts forcing. The feedbacks from increased low cloud, combined with clearsky feedbacks, are why we expect a positive correlation between net α and dS/dT , i.e. less amplifying feedbacks and smaller climate sensitivity for greater dS/dT .

Figure 4.2 (through its parts c–g) supports this. Considering all models and experiments

together, there is a positive correlation between net α and dS/dT (0.59 for the AOGCM experiments, Fig. 4.2c). This suggests that stability may explain much of the spread in α among this set of models, despite the implication of previous work that inter-model differences in climate feedback are dominated by cloud responses to mean SST warming (Ringer et al., 2014). This is still the case when feedbacks are broken down into α_{CS} (Fig. 4.2e) and, though to a lesser extent, α_{CRE} (Fig. 4.2g). We interpret the correlation in Fig. 4.2e, which involves only clearsky feedbacks, as being primarily driven by the linkage between stability and lapse-rate feedbacks since these are what has been found in previous work to be relevant amongst clearsky components, even when cancellation with water vapour feedbacks is accounted for (i.e. under a fixed relative humidity framework) (Andrews & Webb, 2018; Ceppi & Gregory, 2019).

By considering the difference in α between hist-aer and hist-GHG for each model, we remove part of the model spread in α which affects both forcing agents equally, revealing the positive correlation (Fig. 4.2d) between α and stability change in response to different forcing agents. Although the correlation across models is only moderate (0.52), it is nevertheless significant ($p=0.010$), and the difference is also significant in the MMM according to estimated error bars. The lack of correlation in Fig. 4.2h, both across models and in the MMM, despite such correlation in Fig. 4.2f, suggests that the impacts of stability on clearsky feedbacks are more robust than the impacts on α_{CRE} for explaining differences in α between historical AAs and GHGs. This may be because the relationship between S and CRE is not consistently simulated among climate models. Alternatively, it is possible that, contrary to the findings of Ceppi and Gregory (2019), global stability changes are not strongly physically linked to CRE in some of the models, and that regional changes in S would be a better explanatory factor. This disagreement could be related to Ceppi and Gregory (2019) diagnosing cloud-radiative changes through radiative kernels instead of using CRE. Alternatively, the findings of Ceppi and Gregory (2019) that SW cloud feedbacks drive forcing agent feedback differences could be model specific, and not be reflected across all models.

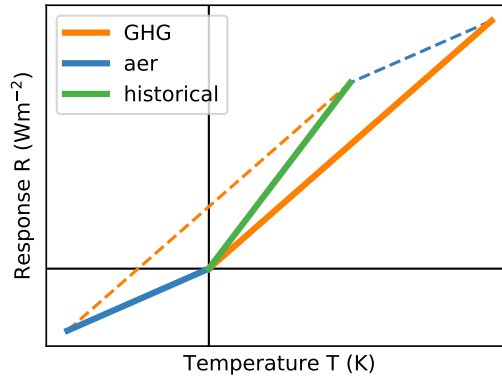


Figure 4.4: A geometric explanation of why the historical all-forcing feedbacks do not lie between the historical AA and historical GHG feedbacks. If the all historical is roughly a linear combination of the historical AA (*blue*) and historical GHG (*orange*) results, and the historical AA induces a negative temperature change with negative forcing unlike the positive forcing and temperature change from GHGs, then the all historical feedbacks parameter (the gradient of the *green* line) will be more positive than the hist-GHG feedback parameter.

The historical all-forcing case is dominated by responses to GHGs and AAs (Smith et al., 2020). Therefore, differences between historical and hist-GHG experiments are mostly due to differences between hist-aer and hist-GHG. Both the feedback parameter (Fig. 4.2a) and the stability response (Fig. 4.2b) are greater for historical than for hist-GHG in the MMM. This results from combining hist-GHG and hist-aer responses, given that AAs and GHGs forcing are of opposite sign (Appendix B in the online supporting information of Gregory and Andrews, 2016). A visual explanation of it can be found in Fig. 4.4.

The results here are in agreement with findings from previous studies of a greater transient climate response (indicative of a less positive α) from aerosols (Marvel et al., 2016) and extratropical forcing (Rose et al., 2014; Rose & Rayborn, 2016; Shindell, 2014) compared to forcing from well-mixed GHGs. By contrast, Richardson et al. (2019) found no significant differences in feedback between two kinds of aerosol (SO_4 and BC) and GHGs, regardless of whether they calculated ERF as in the present chapter, or additionally corrected for the impact of land surface temperature adjustments (Andrews et al., 2021). There are several possible explanations for our disagreement, including the

following. (1) Despite its statistical significance, the difference we find between feedbacks to AA and GHG forcing may be specific to our selection of models, which is smaller than theirs (7, versus 11 in PDRMIP models). (2) Historical AA and the 5xSO₄ forcing in PDRMIP might have important differences in feedback, both because of different SO₄ concentrations and because of contributions of aerosols other than SO₄ (Myhre et al., 2014). (3) The step forcing experiments analysed by Richardson et al. (2019) will be more representative of equilibrium responses than the historical simulations used here, which may affect the extent to which the feedbacks depend on forcing agent.

4.4 Explaining stability differences in hist-aer in terms of extratropical forcing

Next, we interpret the distinct radiative and stability responses to AAs and GHGs in terms of the latitudinal distribution of forcing. Ceppi and Gregory (2019) demonstrated that positive tropical forcing tends to increase global stability per unit global surface warming, while positive extratropical forcing has the opposite impact (and vice versa for negative forcing). To understand why, we recall that the tropics contain the warm pool which is well-coupled to the free troposphere, with the lapse rate closely following a moist adiabat due to moist convection (Flannaghan et al., 2014; Sobel, 2002). Consequently, homogeneous tropical warming has a relatively large impact on free-tropospheric temperatures since warming occurs in these deep convective regions. Mixing by atmospheric motions propagates the warming signal to the extratropical free troposphere, stabilising the atmosphere there (Fig. 4.5b,d). Conversely, positive forcing in the extratropics is expected to decrease stability, since surface temperatures in the extratropics are more weakly coupled to the free troposphere throughout, with deep convective regions absent. The effects of extratropical surface forcing tend to be more confined near to the surface (Fig. 4.5a,c), and since this forcing acts on a region that is (on average) climatologically stable, the stability response is similar to that found for warming in other stable regions

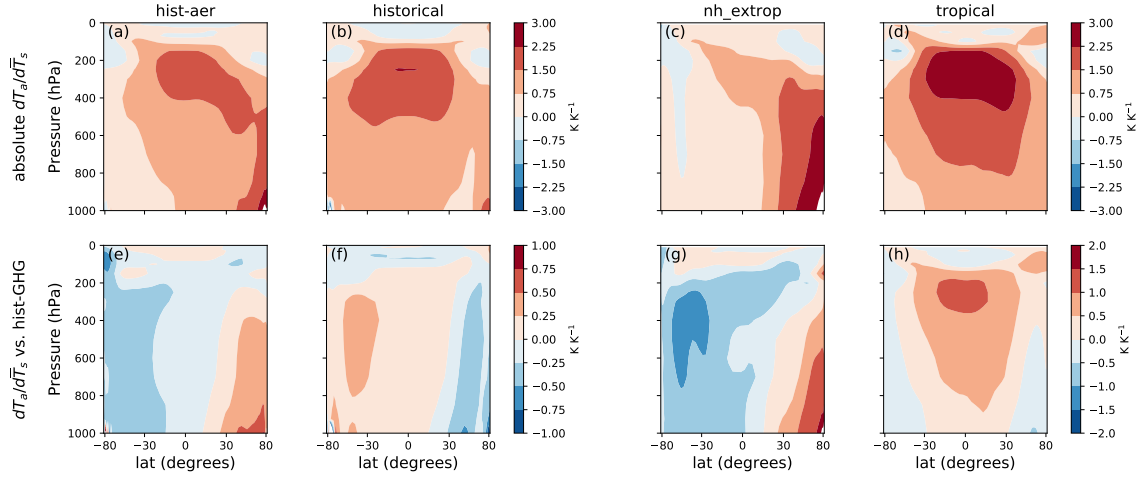


Figure 4.5: MMM zonal-mean profiles of local air temperature regressed onto global-mean surface air temperature. Values shown are absolute (a–b), and relative to hist-GHG (e–f). Also shown are the results from the HadSM3 Northern hemisphere (NH) extratropical forcing (c) and tropical-only forcing (d), and these relative to the HadSM3 2xco2 experiment (g–h). Note that the x -axis is scaled by geographical area.

such as in the tropical South-East Pacific (Andrews & Webb, 2018). This effect can be seen by comparing air temperature changes in the hist-aer and hist-GHG cases (Fig. 4.5e). (Note that aerosol forcing is *negative* and causes a surface *cooling*, but the patterns in Fig. 4.5 are normalised by regression against global mean surface temperature change.)

This reasoning could explain why the hist-aer case gives a less positive stability response per unit surface warming than the hist-GHG case. The skew of forcing towards the extratropics in the hist-aer case (blue line in Fig. 4.1c) means that a relatively larger fraction of the surface temperature response is in vertically decoupled regions, leading to the smaller dS/dT than in hist-GHG. In support of this hypothesis, we note that the pattern of tropospheric temperature change in the HadSM3 nh_extrop experiment compared to the 2xco2 experiment (Fig. 4.5g) is similar to the difference between hist-aer and hist-GHG (Fig. 4.5e). The pattern from tropical-only forcing (Fig. 4.5d) shows the propagation of warming to both the tropical and extratropical free tropospheres in accordance with an increase to stability as seen in Fig. 4.2c–g, although stability is further increased over the tropics than in the more homogeneous 2xco2 case (Fig. 4.5h). Fig. 4.2e shows that global-mean dS/dT is negative for both nh_extrop and sh_extrop, whereas it is positive in nearly every historical forcing experiment. This difference is probably

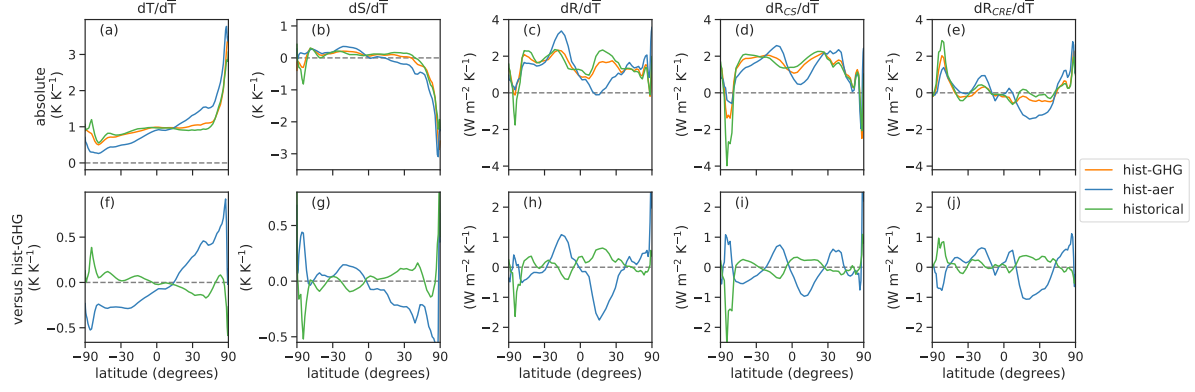


Figure 4.6: MMM zonal-mean values regressed onto global-mean surface air temperature for the historical cases (*top* row) and differenced relative to hist-GHG (*bottom* row) in terms of (a,f) temperature, (b,g) estimated inversion strength, (c,h) net upwards radiative response, (d,i) upwards radiative response in clearsky and (e,j) upwards radiative response from CRE. Note that the x -axis is scaled by geographical area.

related to the absence of tropical forcing in the idealised cases. That the negative dS/dT occurs for forcing in both hemispheres suggests that the essential characteristic is that the forcing is extratropical, rather than hemispheric. The historical all-forcing case shows the opposite pattern to hist-aer when compared to hist-GHG (Fig. 4.5e–f), indicating that AA has a similar effect on stability whether applied independently or jointly with GHGs.

To corroborate this reasoning, we consider the MMM zonal means of feedbacks and climate responses in Fig. 4.6. There is less meridional contrast in response to hist-GHG than hist-aer (top row). Subtracting the responses from hist-GHG (bottom row), we see a positive NH temperature response per unit global warming in hist-aer relative to hist-GHG, and the opposite in the Southern hemisphere (SH). This anti-correlates with the stability response per unit surface warming, which in turn correlates (in low latitudes) with the radiative response. The radiative response is then finally well explained, in terms of pattern, by the combination of clearsky and CRE feedbacks.

Just as the global average feedbacks for the historical case are more positive than those for hist-GHG, the zonal-mean curves for hist-GHG generally lie between those for the net historical and hist-aer cases (top row of Fig. 4.6), again consistent with the expected effect of AA. Likewise, the difference between historical and hist-GHG responses (bottom row of Fig. 4.6) can be interpreted as representing the effect of AA, but with the sign

reversed (see Appendix A).

4.5 Summary and conclusions

Our analysis of AOGCM historical experiments from CMIP6 (including new experiments which allow forcing to be diagnosed) reveals that climate feedback is more strongly amplifying (greater climate sensitivity) in response to AA forcing than GHG forcing. This difference is shown and is statistically significant in the MMM, though only seven AOGCMs have so far provided the required historical experiments and variables for this analysis, so it would be useful to repeat it with more. Our finding is consistent with those from past studies that also found greater climate sensitivity to AA than GHGs (Marvel et al., 2016; Shindell, 2014; Smith & Forster, 2021), but appears inconsistent with the recent study of Richardson et al. (2019). Further work is needed to explain the disagreement, which may relate to differences in the details of the prescribed aerosol forcing or the timescales of calculated feedbacks.

Furthermore, we find that the difference in net top-of-atmosphere radiative feedback parameter for AA and GHG forcing is positively correlated across AOGCMs with a difference in the response of tropospheric stability to the two kinds of forcing. We propose that the difference arises from the different latitudinal distributions of the forcing. An idealised slab model experiment with uniform surface forcing confined to the Northern extratropics qualitatively reproduces the near-surface extratropical temperature change that differentiates the historical AA experiment from the historical GHG experiment. The change in extratropical air temperatures in the former, in the boundary layer but not higher up, is explained by the lower proportion of forcing in the tropics, where the surface is relatively strongly coupled to the free troposphere by deep convection (Flannaghan et al., 2014; Sobel, 2002), compared to the higher proportion of forcing in the extratropics, where the coupling is weaker and the effect of forcing more confined to the surface.

Thus a positive extratropical forcing causes a near-surface warming, which reduces tro-

pospheric stability, whereas a positive forcing that is relatively homogeneous over the tropics contributes to a more positive change in stability per unit surface temperature change. A reduction in stability tends to reduce low-level cloudiness, which gives an anomalously negative (more amplifying) shortwave cloud feedback, whilst it also induces an anomalously negative clearsky feedback. Reasons for this anomalous clearsky feedback are unclear, but could be related to lapse rate and water vapour feedbacks (Ceppi & Gregory, 2019; Po-Chedley et al., 2018). In this way, the latitude of forcing is linked to the radiative feedback it produces. Historical AA forcing is negative, so the signs of temperature and stability change are reversed, but the feedback parameter, sensitivity of stability (change per unit warming), and hence the correlation with the feedback parameter have the same sign for either sign of forcing: extratropical forcing tends to give higher climate sensitivity. This link accords with previous works that have highlighted the impact of forcing patterns on radiative feedbacks (Ceppi & Gregory, 2019; Rose et al., 2014; Rose & Rayborn, 2016; Winton et al., 2010).

Overall, considering the historical period of 1850–2015, historical climate change is dominated by GHG forcing. Hence the net feedback simulated in the historical experiments with all forcings is nearer to that for GHG than for AA. The effective climate sensitivity for historical forcing is slightly *smaller* than for historical GHG forcing (the magnitude of α is larger), because combining negative hist-aer forcing with positive hist-GHG forcing reduces the warming by more as a proportion than it reduces the net forcing, due to the greater sensitivity to AA forcing. We find also that some of the spread across AOGCMs in the climate sensitivity to GHG forcing is correlated with the response of tropospheric stability to forcing (Fig. 4.2c), especially in clearsky (Fig. 4.2e); this aspect is intriguing and requires further investigation.

Chapter 5

TIME-EVOLUTION OF FEEDBACKS IN THE HISTORICAL PERIOD

The preceding chapter has established that stability serves as a basis for explaining different radiative feedbacks between AAs and GHGs, when calculating feedbacks over the full historical period. In this chapter, we investigate the time-dependence of radiative feedbacks in the historical case, and assess to what extent this time-dependence results from the time-varying mix of historical forcing agents. To do this, we use the individual historical GHG and AA forcing experiments, as in the previous chapter, but we also introduce the historical natural forcing experiment which involves strongly time-varying forcing. We find that the historical all-forcing experiment exhibits significantly time-evolving feedbacks, which can be explained through contributions of component forcings. We also highlight that volcanic forcing is especially important for understanding evolving historical feedbacks, as it exhibits particularly strong feedbacks that we relate to large stability changes due to enhanced surface temperature responses in the warm pool.

5.1 Introduction

The standard surface temperature change model of

$$N = F - R = F - \alpha T \tag{5.1}$$

was formerly assumed to have a constant α parameter i.e. one that does not vary in time. It is surprising a priori that a constant α could adequately describe the combination of non-linear feedbacks in the climate system, but much previous work has found this to be a good first approximation. However, analysis of experiments with general circulation models (GCMs) suggests that α is not truly constant, but time-dependent, both during historical climate change (Gregory & Andrews, 2016; Andrews et al., 2018; Gregory et al., 2020; Dong et al., 2020; Andrews et al., 2022) and climate change produced by assumed forcing scenarios, even in cases of fixed forcings such as quadrupled CO₂ (Andrews et al., 2015; Ceppi et al., 2017; Proistosescu & Huybers, 2017; M. Rugenstein et al., 2019; M. A. A. Rugenstein & Armour, 2021). The SST “pattern effect” (Stevens et al., 2016) has been identified as responsible for evolving feedbacks as different areas of the Earth’s surface warm at different rates (Andrews et al., 2015; Zhou et al., 2017). The physical mechanism driving this has been explained in previous work as tied to atmospheric stability changes, with this affecting the radiative response primarily through changes in low cloud amount, lapse rate, and water vapour (Ceppi & Gregory, 2017, 2019; Kamae et al., 2019; Mackie et al., 2021). Hence, as in chapter 4, it is desirable to understand the role of tropospheric stability for the radiative feedbacks and their time evolution in the historical period.

Over this period, we expect radiative feedbacks to evolve not only as they do under constant forcing scenarios, but also due to the evolving combination of forcing agents responsible for historical climate change, as long as there is a forcing-agent dependence on feedbacks as suggested by chapter 4. By looking at experiments that involve only one historical forcing agent, it may be possible to understand the contributions of different agents to the evolution of historical feedbacks. Such work has already been done by Gregory et al. (2020) using CMIP5 data, who found that GHG feedbacks do not vary significantly in time (outside of unforced variability). They also identified natural forcing as important for determining the time-evolution of radiative feedbacks in the historical case, and found feedbacks from volcanic eruptions to be less amplifying than those from GHGs. Since this, CMIP6 data has become available containing the single-forcing DAMIP

(AOGCM; Gillett et al., 2016) and RFMIP (AGCM; Pincus et al., 2016) experiments. The latter involve fixed-SST simulations with time-evolving forcings, which allows for model-specific calculation of historical effective radiative forcings over time. This provides an advantage over Gregory et al. (2020) who had to estimate forcings over time, and used the same forcing per agent across models. Dong et al. (2021) have already used this new data to show that feedbacks in historical experiments feedbacks are mainly driven by GHG forcing from around 1940 onwards.

CMIP5 and CMIP6 AOGCMs generally fail to reproduce the observed historical SST pattern of recent decades, especially in the Pacific, and the consequent large positive α (Gregory & Andrews, 2016; Andrews et al., 2018; Dong et al., 2019; Gregory et al., 2020; Dong et al., 2021; Wills et al., 2022; Andrews et al., 2022). Since AOGCMs are essential for producing physically based estimates of future warming, we need to know why they do not reproduce the past in this respect. Not having a multi-centennial observational SST record for comparison, we cannot tell whether the relevant SST pattern is an unusual unforced variation, or a response to recent forcing. In either case, improved understanding of simulated historical feedbacks will be useful in explaining the difference from the observed record.

From this, there are a few questions that this chapter seeks to answer. The first considers how feedbacks vary over time in the historical case. With new data allowing for a better understanding of these time-evolving feedbacks, the second question becomes: can this time-evolution be understood by analysing component forcings of the historical period and their individual time-evolving feedbacks? Such insight is desired, in the hope that this will help determine what feedbacks will act in future climate change, as noted by the range of past works that have raised this question (Andrews et al., 2018; Dong et al., 2019; Gregory & Andrews, 2016). Finally, if a time-varying mix of forcing agents explains the time-evolving feedbacks of the historical case, then what are the relevant features of these independent forcings and how, through the relevant physical mechanisms, can their feedbacks be explained?

Table 5.1: Names and details of the CMIP6 experiments specific to this chapter. Data from experiments described in Table 4.1 are also used, with the same models as before for the piClim-histnat and hist-nat experiments. Unavailable data is marked by an x.

	piControl	piClim-histnat	hist-nat	amip-piForcing
CanESM5	r1i1p1f1	r1i1p2f1	r[1-9]i1p2f1	r1i1p2f1
CESM	r1i1p1f1	x	x	r1i1p1f1
CNRM-CM6-1	r1i1p1f2	r1i1p1f2	r[1-9]i1p1f2	r1i1p1f2
GISS-E2-1-G	r1i1p1f1	r1i1p1f1	r[1-9]i1p1f3	x
HadGEM3-GC31-LL	r1i1p1f1	r[1-3]i1p1f3	r[1-4]i1p1f3	r1i1p1f3
IPSL-CM6A-LR	r1i1p1f1	r1i1p1f1	r[1-9]i1p1f1	r1i1p1f1
MIROC6	r1i1p1f1	r[1-3]i1p1f1	r[1-3]i1p1f1	r1i1p1f1
MRI-ESM2-0	r1i1p1f1	x	x	r1i1p1f1
NorESM2-LM	r1i1p1f1	r[1-3]i1p1f1	r[1-3]i1p1f1	x

5.2 Data and methods

Most of the data referred to in this chapter is the same as that used in chapter 4, i.e. data from the ESGF CEDA archive for seven CMIP6 models. The same model output variables were required for this analysis (cf. Table 4.1). Some extra data is used here, which includes the historical natural forcing experiments for both coupled atmosphere-ocean (hist-nat, from DAMIP; Gillett et al., 2016) and fixed-SST experiments (piClim-histnat, from RFMIP; Pincus et al., 2016). Variants were chosen with the same criteria as for the data used in chapter 4. See section 4.2 for an explanation of variant labels (i,r,p,f). Where possible, variants were taken with the same initialisation (i1 variant label), physics and forcing definition as in piControl. Note that, for GISS-E2-1-G, the variant chosen for hist-nat was f3 since f1 was unavailable and f2 does not include solar nor volcanic forcings whereas f3 does. The difference between f1 and f3 for hist-nat is that f3 forcings include a correction to ozone-related heterogeneous reactions. We included all realisations that contained all of the required variables, and our results from each experiment of each model are ensemble averages. Available ensemble members for the new data are listed in Table 5.1.

Also included is data from the amip-piForcing experiments which prescribe SSTs according to the Atmosphere Model Intercomparison Project (AMIP) II data set (Gregory & Andrews, 2016; Webb et al., 2017; Andrews et al., 2018), which is based on historical

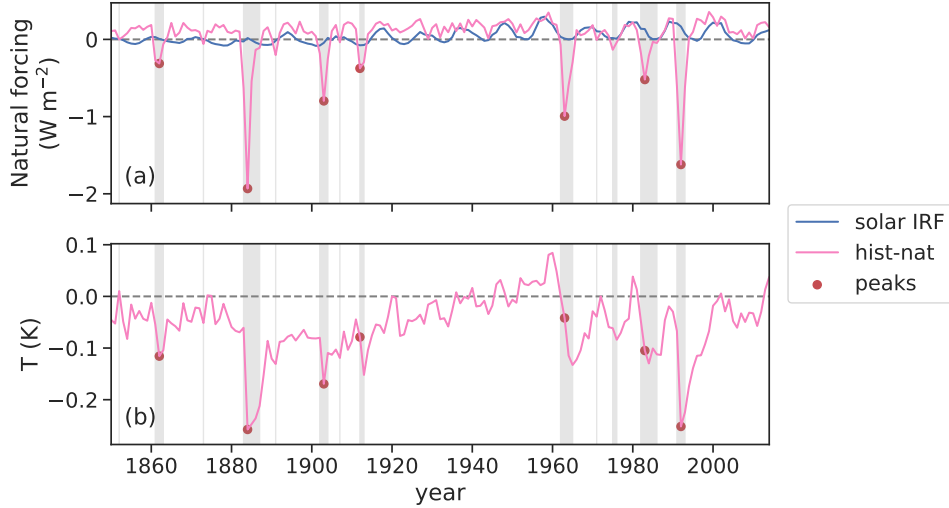


Figure 5.1: Annual-means of effective TOA radiative forcing over time (*top*) and global average surface temperature change (*bottom*) in hist-nat over the 1850–2015 historical period. The seven largest peaks of negative forcing of the period (corresponding to the seven most climatologically significant volcanic eruptions) are marked by red dots, whilst years of measured negative forcing (i.e. volcanic forcing, which makes up “hist-nat-volc”) are shaded in grey. Instantaneous radiative forcing (IRF) from changes in downwelling shortwave at the TOA is shown in blue.

observations. Only single ensemble members were included for each model for these experiments, but this is not expected to be a limiting factor as the noise in amip-piForcing results comes mostly from variation in prescribed SSTs which is not reduced by ensemble size (Gregory & Andrews, 2016). Note that models with amip-piForcing experiment outputs are not the same as those for individual agent historical forcing experiments, though this is unlikely important for the results discussed below.

Although there are lists of historically significant volcanic eruptions (Mass & Portman, 1989; Stenchikov et al., 1998), we chose to identify radiatively significant volcanic forcing events by looking at the time series of effective forcing in piClim-histnat. We used SciPy’s peak finding algorithm, which uses a simple comparison of neighbouring values, to find negative peaks of greater than 0.3 W m^{-2} magnitude in effective forcing. Identified peaks of negative forcing are shown in red in Fig. 5.1. The first identified eruption through this method is in 1861. This is unidentified in the dataset (Durack & Taylor, 2022), which was based on ice core data (Arfeuille et al., 2014), though this lines up roughly with the Dubbi (Wiert & Oppenheimer, 2000) and Makian (Picas & Grab, 2020) eruptions of

that year. As such, we simply call this the "Early" eruption. Later eruptions are also identified: Krakatoa (1883), Santa María/Mount Pelée/Soufrière which we refer to as "Santa María" (1902), Katmai/Novarupta which we refer to as "Novarupta" (1912), Agung (1963), El Chichón (1982), and Pinatubo (1991) (Mass & Portman, 1989; Stenchikov et al., 1998). There are many other eruptions that occurred over the 1850–2015 historical period, including some highly explosive ones (high VEI Newhall & Self, 1982) such as Mt. St. Helens and Cerro Hudson (A. Schmidt et al., 2018), but these have relatively little global average radiative impact.

With the most radiatively significant volcanic events identified, which demonstrates the accuracy of this method, we also look at all years involving radiatively significant volcanic forcing. To do this, we define volcanic forcing years as any years where effective forcing is below zero, which gives us the years highlighted by grey bars in Fig. 5.1. This is justified by the presence of weak background volcanic aerosol (equivalent to the historical average over 1850–2015) within piControl (Eyring et al., 2016), which imposes a positive forcing relative to control when this background is not present in the forced experiments (Gregory et al., 2013). This can be seen from the 1920–1960 period, free of significant eruptions, during which warming occurs. With solar forcing being evidently very small (*blue* in Fig. 5.1a), negative forcing values in hist-nat forcing relative to piControl are thus likely to be years where volcanic forcing is greater than the average over the 1850–2015 period.

As in chapter 4, global averages here refer to area-weighted global averages of the variables of interest, except for stability S which refers to the average EIS over oceans equatorwards of 50°S and 50°N. Again, we choose EIS as our measure of atmospheric stability in this chapter following related work (Ceppi & Gregory, 2019).

5.3 Dependence of climate feedback on forcing agent

In order to understand the dependence of α on historical forcing agents, we use the refined model of Ceppi and Gregory (2019). The Earth energy budget model of equation 5.1 gives

$$R = \alpha T = F - N \implies \frac{dR}{dT} = \alpha, \quad (5.2)$$

if α is constant. To account for the variation of α shown by GCM experiments, Ceppi and Gregory (2019) proposed a refined model, wherein R is described as a combined response to both T and global average atmospheric stability change S :

$$R = \tau T + \sigma S \implies \alpha = \frac{dR}{dT} = \tau + \sigma \frac{dS}{dT}, \quad (5.3)$$

where

$$\tau = \frac{\partial R}{\partial T} \quad (5.4)$$

and

$$\sigma = \frac{\partial R}{\partial S}. \quad (5.5)$$

Through uniform warming experiments in two models, Ceppi and Gregory (2019) found that τ and σ are approximately constant, i.e. the radiative response and stability change are both roughly linear with uniform global surface temperature change. Therefore, a change in $\alpha = dR/dT$ may be explained by a change in dS/dT (Eq. 5.3).

Before looking at time variations within historical feedbacks, we analyse overall radiative feedbacks of the 1850–2015 period. These best-fit constant feedbacks are not a suggestion that α is unchanging over the historical period, but an indication of the radiative changes caused overall across the historical period. We introduce two new experiments to the analysis done in chapter 4. The linear fits in Fig. 5.2a show that the positive relationship of α and dS/dT , found when comparing hist-aer and hist-GHG (Fig. 4.2), is maintained when both natural (hist-nat) and historical observed SST (amip-piForcing)

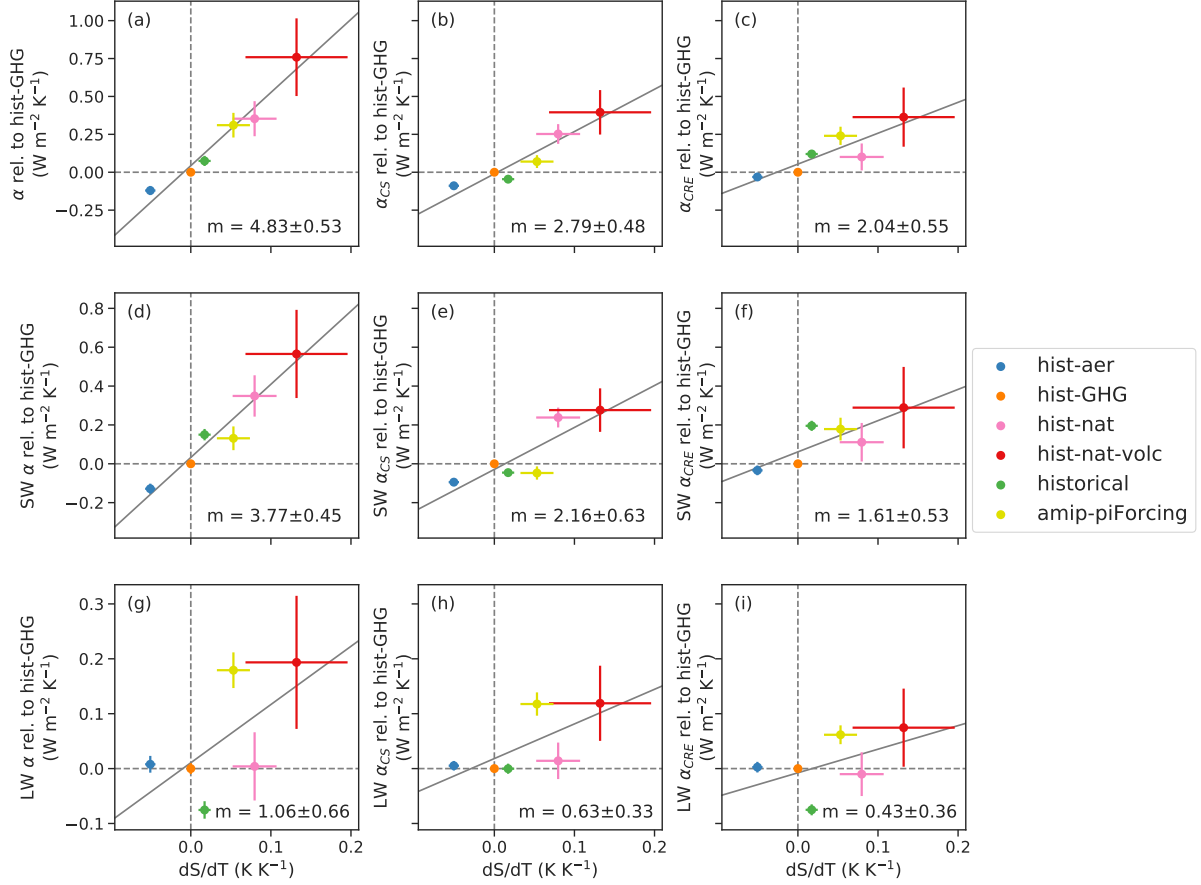


Figure 5.2: MMM radiative feedbacks α against dS/dT , with values calculated from regression over the full historical period (1850–2015) for hist-GHG, hist-aer, and hist-nat. Also shown are values for amip-piForcing over available years (1870–2015), and regression over hist-nat using only volcanic years of hist-nat (“hist-nat-volc”, see Fig. 5.1 for selection). Gradients m of least-squares fits are shown in each panel. Error bars denote estimated regression errors only.

forcings are considered. This suggests that feedbacks from historical natural forcings can be understood in terms of their effect on stability. We thus further investigate how stability changes affect LW, SW, clearsky, and CRE contributions to radiative feedbacks. CRE, which is just the difference between clearsky and allsky fluxes, is used here to understand feedbacks from cloud changes.

Similar to what was found in Fig. 4.2, clearsky (Fig. 5.2b) and CRE (Fig. 5.2c) radiative changes are both found to be required for explaining the overall feedback differences between forcing agents when these further cases are considered. Breaking things down instead into SW (Fig. 5.2d) and LW (Fig. 5.2g) contributions, SW appears to be dominant in explaining α differences, with both higher spread in and correlation between SW α

and dS/dT than LW α and dS/dT .

Although the accuracy of convection within models is limited by necessary parameterisations (Jones & Randall, 2011), in this case we seem to capture the relationship between stability and cloud as we expect. The weak relationship between LW CRE feedbacks and stability (Fig. 5.2i) is expected, given that stability should mainly affect low clouds (Zhou et al., 2016; Ceppi et al., 2017; Andrews & Webb, 2018; Ceppi & Gregory, 2019) that have little LW effect (Hartmann, 2016). Changing low cloud formation with stability is well related to SW reflection changes, and thus SW CRE feedbacks (Fig. 5.2f).

The substantial contribution of clearsky to overall feedback differences is less clear. Clearsky feedbacks are expected to be due to albedo, water vapour, and lapse rate changes, with the latter the most clearly tied to stability changes as lapse rate and stability are similar measures of vertical temperature gradients in the atmosphere. However, despite the lapse rate effect being a LW feedback, most of the relationship between clearsky feedbacks and stability originates from the SW component (Fig. 5.2e). Both water vapour and albedo feedbacks may contribute to the forcing agent spread in SW feedbacks. Since albedo feedbacks have no obvious relationship with atmospheric stability, it is possible that differences in albedo feedbacks are not caused by, but instead co-vary with, different stability changes, as noted by Ceppi and Gregory (2019).

Interestingly, the breakdown fails to reveal any component that dominates in explaining the overall good positive agreement between α and dS/dT . SW is more important than LW, but SW feedback differences are dominated by neither clearsky nor CRE. There are two possibilities: either this relationship works coincidentally well for the forcing agents analysed here, or there is some unexplained co-dependence between clearsky and CRE feedbacks in relation to stability change. We note here that more easily interpretable relationships might be obtained if feedbacks were broken down into non-cloud and cloud feedbacks instead of CRE, which aliases in some clearsky effects.

Having looked at overall feedbacks from the 1850–2015 period, we seek to justify trying to understand the time-evolution of feedbacks. Figure 5.3 highlights the evolving contri-

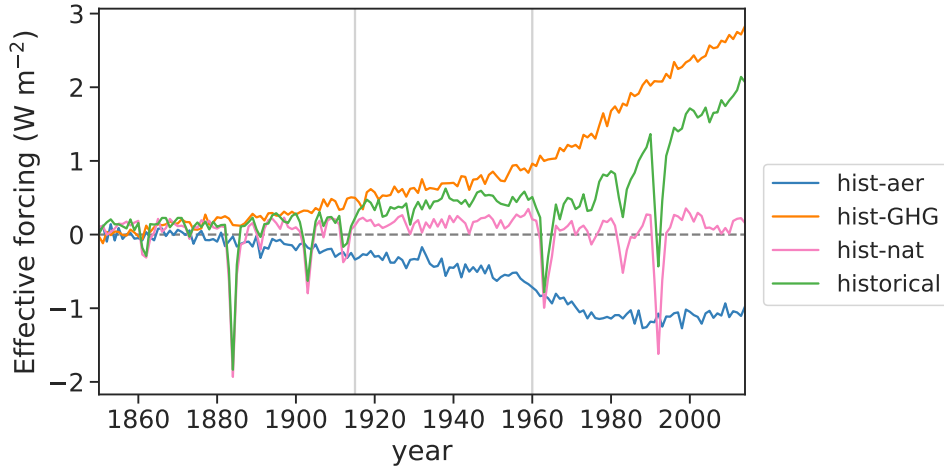


Figure 5.3: Forcing over time in the historical case and for simulations of component historical forcing agents. Values were calculated from RFMIP piClim-histxxx experiments with piClim-control as the control. Vertical grey lines demarcate the three different periods of different forcing flavours.

butions of different forcing agents to the historical case over the 1850–2015 period. For this analysis, we point out that the historical period can be roughly divided into three: an early one where both AA and GHG forcings are small but several significant eruptions occur (1850–1914), a middle period where both AA and GHG forcing are significant but no significant volcanic eruptions occur (1915–1959), and a late period which contains strong forcing from both AAs and GHGs alongside multiple significant volcanic eruptions (1960–2015).

Natural forcing exhibits a complex temporal structure compared to hist-GHG forcing. This is highlighted by the variability of natural forcing (*pink* in Fig. 5.3), as well as in Fig. 5.2 when comparing the values obtained by regression over the full period (*pink*) versus only significant volcanic forcing years (*red*). This complexity stems from natural forcing being composed of both volcanic and solar forcings, while the volcanic eruptions themselves are diverse in terms of forcing magnitude and spatial structure (cf. section 5.6). By contrast to highly homogeneous hist-GHG forcing, and the slowly evolving pattern of hist-aer forcing (Wang & Wen, 2021), natural forcing varies over very brief timescales as forcing from volcanic eruptions decays rapidly over a few years and many years can pass between significant volcanic eruptions (Ammann et al., 2003). Given such

variability in natural forcing, it is important to understand its evolution and the effects of volcanic eruptions on radiative feedbacks.

5.4 Time-evolution of radiative feedbacks on a forcing agent basis

Figure 5.4 shows the corresponding time-evolving feedbacks and dS/dT for regressions over 30-year sliding windows, following the approach of Gregory and Andrews (2016) and Gregory et al. (2020). Both α and dS/dT vary significantly over time in all cases except hist-GHG. In order to test the correlation between α and dS/dT we use consecutive non-overlapping 30-year windows from 1855–2015. We ignore the 1915–1945 window for hist-nat since no significant volcanic forcing occurs then (see Fig. 5.1). Doing this, we find a high correlation ($\rho > 0.75$) between α and dS/dT for each forcing, excepting hist-GHG. This suggests that stability variations explain time-variations in radiative feedbacks.

Historical AA forcing (Fig. 5.4a) causes a relatively large α between around 1920–1930, in contrast to the lower α (compared to under GHG forcing) that was calculated over the full historical period (Fig. 5.2). This feature is not obviously related to the evolving patterns of aerosol forcing, which are known to have shifted from being mainly over Europe and north America to their increasing prominence in east and south Asia (Lamarque et al., 2010; Wang & Wen, 2021), despite shifting forcing patterns being tied to different efficacies (Persad & Caldeira, 2018). The reason for the change in feedbacks for hist-aer could be related to changing relative concentrations of different AAs, as highlighted by an increase in radiative forcing from sulphates relative to black carbon over the historical period (Myhre et al., 2014, their Fig. 8.8). A shift from high α when BC dominates (around 1925) to low α as sulphates become more important (1950 onwards) would be consistent with the finding by Ceppi and Gregory (2019) of larger α (lower sensitivity) from BC than sulphate forcing. This is an uncertain explanation, however, as other works found opposite results (Nordling et al., 2021; Richardson et al., 2019).

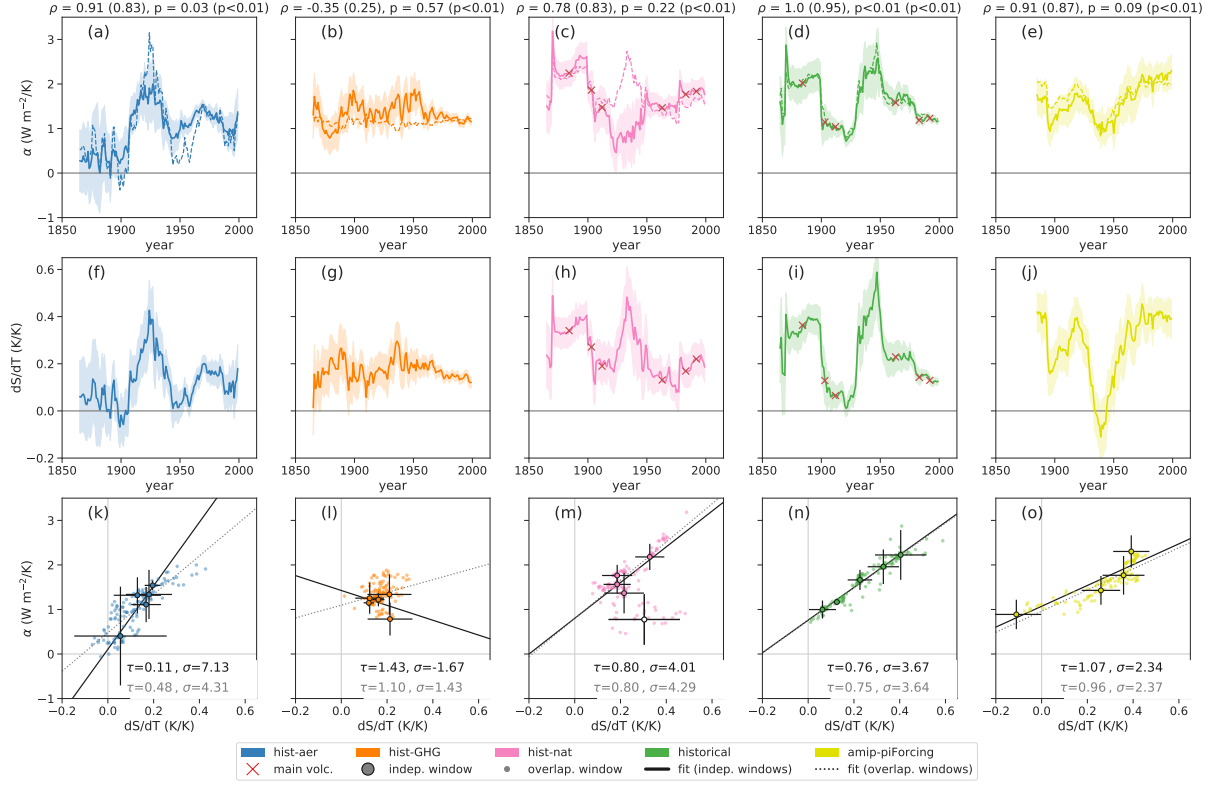


Figure 5.4: *Top and middle rows:* Time-evolving feedback parameter α (*top row*) and dS/dT (*middle row*) for various historical experiments as calculated using 30-year sliding windows. Coloured shaded regions denote regression errors. Dashed time-series in the top row show best linear fits of dS/dT to α using the non-overlapping windows from 1855–2015. Pearson correlation coefficients of these fits to α (ρ), as well as regression significance (p), are shown. In brackets, we show ρ and p for fits using overlapping windows. For hist-nat, both regressions ignore periods free from significant eruptions. Peaks of volcanic forcing are shown by red crosses for the hist-nat and historical cases. *Bottom row:* year-by-year scatter of α and dS/dT from 30-year windows. Circles with black edges show non-overlapping windows, whilst dots show windows that overlap with these. Linear fit coefficients for the gradient σ and intercept τ are shown, across independent windows (black, solid line fit) and all available windows (grey, dotted line fit). The window free of significant volcanic eruptions is ignored for hist-nat as above, and is shown by a white circle with a black edge in panel (m).

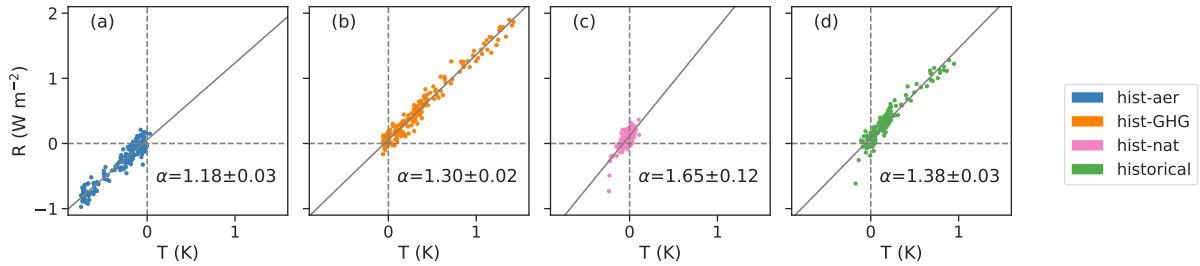


Figure 5.5: MMM responses R against surface temperature change T , for (a) hist-aer, (b) hist-GHG, (c) hist-nat, and (d) the historical experiments. Each dot marks a yearly average, with linear best-fits (of gradient α) in solid grey.

For hist-GHG (Fig. 5.4b,g), the low correlation ($\rho = -0.35$) across time between dS/dT and α is due to the lack of forced variations in these two parameters. With the hist-GHG forcing pattern remaining homogeneous throughout the historical period, only its magnitude slowly increasing, the inter-annual evolution of dS/dT and α is dominated by unforced variability, unlike the other cases where the forcing patterns and/or magnitudes vary more rapidly in time. This is highlighted by the good correlation of R with T over a larger T range than for the other experiments (Fig. 5.5), such that for hist-GHG both dR/dT and dS/dT are nearly constant, and their correlation is negligible in time. The lack of dS/dT and α variability in hist-GHG can be seen clearly from the tight scatter of points compared to the greater spread seen in other cases (Fig. 5.4l compared with other panels of the bottom row).

The hist-nat case is dominated by volcanic forcing (Fig. 5.4c,h). Low correlation between dS/dT and α in the middle period lines up with a lack of significant eruptions, which means there is no forced signal then. This suggests that stability is a good explanation of feedbacks as long as there is enough of a signal, i.e. enough variation in the forcing and thus response over time. In other periods, the hist-nat experiment shows significant variations in α and dS/dT which are possibly related to different eruptions causing different feedbacks, for example via differences in the spatial distributions of stratospheric aerosol and hence radiative forcing. For the full historical period, volcanic forcing gives less amplifying feedbacks (Fig. 5.2), and for the two eruptions with the strongest signal (Krakatoa in 1883 and Pinatubo in 1991) α tends to larger values than for hist-GHG.

We explain high correlation ($\rho = 1$ to 2 s.f., Fig. 5.4n) between α and dS/dT throughout the historical all-forcing case through significant forcing being present across the whole period. Before 1900 (Fig. 5.4d,i), variations in α and dS/dT very closely mirror those found in hist-nat, because AA and GHG forcing are relatively small. After 1980, α tends towards similar values as in hist-GHG. This could be due to a combination of factors, namely that hist-GHG forcing dominates towards the end of this historical period, whilst the strong Pinatubo eruption, with higher α , and aerosol forcing, with lower α , likely

cancel out in their effects on radiative feedbacks.

The reason for the large variations in α and dS/dT in the middle period in the historical all-forcing case is less clear. One possible explanation is that AA and GHG forcing tendencies roughly cancel out (Fig. 5.3), such that regressions in temperature over this period are sensitive to unforced variations in T . Alternatively, variations in α for hist-aer may explain it as follows: high α from hist-aer during 1910–1930 results in α being *less* positive in the historical all-forcing case than for hist-GHG over the same period, whilst low α for hist-aer during 1935–1955 results in *more* positive α for the historical all-forcing case (refer to Fig. 4.4).

The amip-piForcing experiment (Fig. 5.4e,j) shows a different time-evolution to the historical all-forcing case. While we do not expect the coupled multi-model mean to reproduce the variability found in the observed single realisation of historical climate, the disagreement may also reflect a failure of models to recreate historical SST patterns, whether forced or unforced (Gregory et al., 2020; Dong et al., 2021; Wills et al., 2022). Observational errors in the AMIP-II SST boundary condition dataset may also play a role (Fueglistaler & Silvers, 2021; Andrews et al., 2022).

The correlation is high ($\rho > 0.75$) between α and dS/dT in all cases except for hist-GHG forcing. This is true whether we use α and dS/dT values from overlapping windows or not. The p-value for the ordinary linear regressions is highly significant when using overlapping windows (always $p < 0.01$), but this may be a misrepresentation of significance. Conservatively, using only independent/non-overlapping 30-year windows to estimate the relationship between α and dS/dT , the regression slopes are significantly different from zero ($p < 0.05$) for hist-aer and hist-all. Importantly, though, we find a correlation between α and dS/dT across experiments, which is very unlikely to be fortuitous ($p < 0.01$, from multiplying p -values across experiments).

5.5 Reconstructing the historical case from individual historical forcing experiments

In order to quantify the contributions of different forcing agents to the historical radiative feedbacks, we attempt to reconstruct the historical values from the results of historical single-forcing experiments. The reconstruction is made thus: for the hist-aer and hist-GHG contributions to the historical, the temperature changes predicted by the relevant single forcing experiments are taken. These are multiplied by fixed α and dS/dT values calculated from the full 1850–2015 period from those experiments. This provides contributions to the historical R and S changes. The use of a fixed feedback parameter for the hist-GHG contribution is justified by little variability in radiative feedbacks from hist-GHG forcing over the historical period. For hist-aer, although radiative feedbacks do vary considerably over time, they vary more strongly earlier on when the AA forcing is relatively weak, so these large variations should not contribute strongly to the historical evolution of feedbacks. For the contribution from hist-nat, the model-diagnosed R and S are taken directly, since volcanic forcing is so variable that fixed α and dS/dT do not accurately apply.

Such a reconstruction of the historical case is shown in Fig. 5.6 (*red*), contrasted with the model output for the historical case (*black*) and the straightforward sum of changes from individual historical forcing cases (*grey*). The contributions to R , S , and T are summed, giving the left column of Fig. 5.6 after taking a 3-year running mean in order to remove short-term noise (for easier comparison). This process is summarised by the following equations, for the straightforward sums of model outputs:

$$R_{sum}(t) = R_{GHG}(t) + R_{AA}(t) + R_{nat}(t) \quad (5.6)$$

$$S_{sum}(t) = S_{GHG}(t) + S_{AA}(t) + S_{nat}(t) \quad (5.7)$$

$$T_{sum}(t) = T_{GHG}(t) + T_{AA}(t) + T_{nat}(t) \quad (5.8)$$

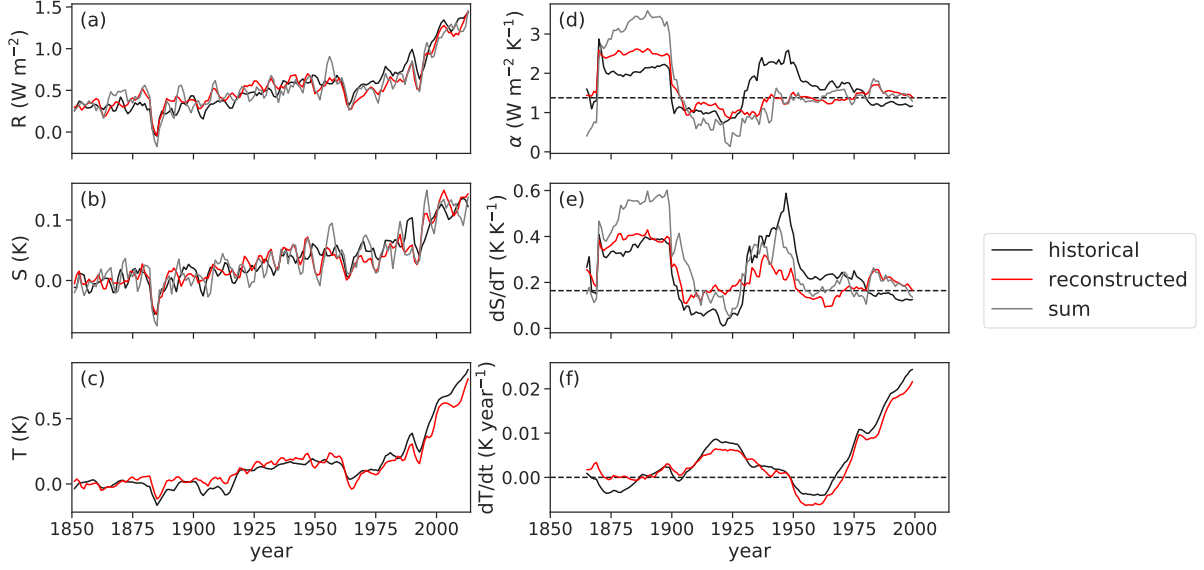


Figure 5.6: Model results from the historical (*black*) compared to the sum of model results from hist-nat, hist-GHG, and hist-aer cases (*grey*), along with the reconstruction using fixed α for AA and GHG forcings but time-varying α for natural forcing (*red*, detailed in the text). (a-c) 3-year running means of (a) the radiative response, (b) 50S-50N EIS changes over ocean, (c) surface air temperature change. (d-f) 30-year regressions of same variables as to their left (d-e) over surface temperature and (f) over time. Note that for T and dT/dt , the reconstruction and sum are identical as the reconstruction uses the sum of model output temperatures.

and the reconstructions:

$$R_{rec}(t) = \alpha_{GHG}T_{GHG}(t) + \alpha_{AA}T_{AA}(t) + R_{nat}(t) \quad (5.9)$$

$$S_{rec}(t) = \left(\frac{dS}{dT}\right)_{GHG} T_{GHG}(t) + \left(\frac{dS}{dT}\right)_{AA} T_{AA}(t) + S_{nat}(t) \quad (5.10)$$

$$T_{rec}(t) = T_{sum}(t) \quad (5.11)$$

where all of R , S , and T are changes from the control state. The reconstructed R and S changes are then regressed against the reconstructed T changes (i.e. the sum of model-diagnosed T contributions) over 30-year sliding windows to produce (d-e), whilst (f) is produced as a regression of T changes against time over 30-year windows.

The left column (a-c) shows that the reconstruction explains the overall R , S , and T changes well, with correlations of $\rho(R_{rec}, R_{hist}) = 0.90$, $\rho(S_{rec}, S_{hist}) = 0.79$, and $\rho(T_{sum}, T_{hist}) = 0.96$. These are higher correlations than between the summed single-

forcing results and the all-historical forcing results ($\rho(R_{sum}, R_{hist}) = 0.83$, $\rho(S_{sum}, S_{hist}) = 0.67$), and significantly so ($p < 0.002$ matching correlations, in both cases). This is likely fortuitous, as the reconstruction should at best closely match the sum of single-forcing outputs. Mismatches between the reconstruction and the model-diagnosed values can be attributed to missing forcings (ozone, land-use change) or non-linearities in responses to different forcings, though the high correlations suggest these factors are minor.

The right column of Fig. 5.6 shows that α and dS/dT are also well-represented by the reconstruction. Correlations are lower for the sliding regressions than for the raw values ($\rho(\alpha_{rec}, \alpha_{hist}) = 0.53$, $\rho(dS/dT_{rec}, dS/dT_{hist}) = 0.62$, $\rho(\alpha_{sum}, \alpha_{hist}) = 0.56$, $\rho(dS/dT_{sum}, dS/dT_{hist}) = 0.72$) but always significant ($p < 10^{-10}$). Looking at the plots qualitatively, the reconstruction fails in a few places. These include the early period where the reconstruction actually matches the diagnosed historical results well but not the sum of single-forcing results, indicating a match for the wrong reasons, and around 1940 where the α and dS/dT are underestimated. These periods also line up with periods of little T evolution (Fig. 5.6f), which suggests this could be a signal-to-noise issue as the regressions become sensitive to very small changes in the denominator. Overall, however, the reconstruction still captures the shape of how α evolves in the historical case. This supports the reasoning that AOGCM-derived historical feedbacks can be interpreted in terms of a time-varying mix of GHG, AA and volcanic forcings. Hence, any non-linearities that arise from the combination of these agents do not play a first-order role for the time-variation in historical feedbacks.

5.6 Explaining a stronger pattern effect from volcanic forcing

Some insight into the impact of volcanic forcing on time-evolving feedbacks might be gained by looking at particular eruptions, given that they exhibit very different forcing patterns (Fig. 5.7). For example, Novarupta (1912) is a high latitude eruption that

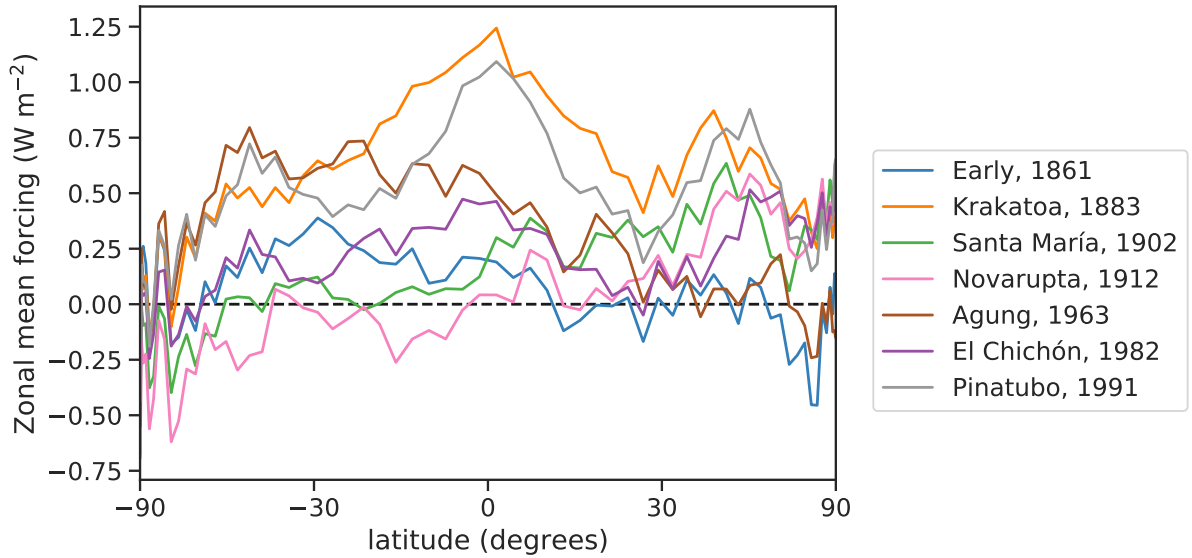


Figure 5.7: Zonal mean radiative forcing by latitude for various volcanic eruptions. Values are time-averages using year preceding, the year of, and the three years following the eruption.

causes NH extra-tropical forcing, unlike the larger Krakatoa (1883) and Pinatubo (1991) eruptions which display more homogeneous forcing with a slight focus near the equator. Given the effect of extra-tropical forcing found in chapter 4, we thus expect forcing from Novarupta to result in a very small α . Unfortunately, the small signal (relative to unforced variability) for all eruptions except for Krakatoa (1883) and Pinatubo (1991) means that analysing individual eruptions is unreliable, with values highly dependent on measurement specifics such as regression period (Fig. 5.8).

As highlighted by Fig. 5.2, the overall effect of volcanic forcing is to produce significantly stronger (less amplifying) radiative feedbacks than GHGs. This can be seen when looking at the overall α for hist-nat (*pink*) and is emphasised by regressions over only volcanic forcing years (hist-nat-volc, *red*). In order to understand this, we look at the two most significant volcanic eruptions of this period: Krakatoa (1883) and Pinatubo (1991). These eruptions have clear signal in all of the models analysed here, with correspondingly the largest responses in the MMM (Fig. 5.1).

In the analysis that follows, the “main volcanic” years that are discussed are years 1882–1886 (for Krakatoa) and 1990–1994 (for Pinatubo), i.e. for each of the two main eruptions

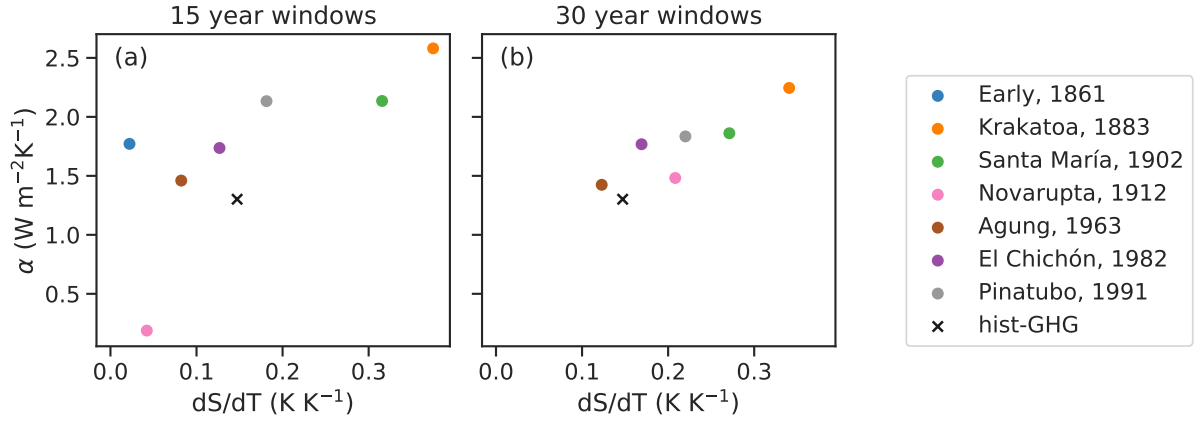


Figure 5.8: Radiative feedback parameter α against dS/dT , both calculated using regressions centred on various volcanic eruptions, using 15-year windows (*left*) and 30-year windows (*right*). Also shown are values calculated across the full 1850–2015 time series for hist-GHG. All values are for the MMM.

we have taken the year before the eruption, the year of the eruption, and the three years afterwards. These ranges were chosen to ensure that the full eruptions were captured, with the year before included since the eruption dates are defined from yearly peaks in forcing which can occur in the year after the actual eruption year (i.e. if the eruption happened late in the year). The three years following the peak forcing year are included to allow the volcanic forcing signal to decay to below noise (which happens within around 2–3 years). The Krakatoa and Pinatubo eruptions themselves are qualitatively similar in terms of radiative forcing, as they are equatorial and occurred over NH summer months (Mass & Portman, 1989; Stenchikov et al., 1998).

Unlike AA forcing (chapter 4), volcanic forcing from the two main eruptions of the 1850–2014 period does not show a noticeable tropical–extratropical contrast (Fig. 5.9a). The significantly greater dS/dT for volcanic forcing is instead tied to warm pool temperature change.

We make use of the SST# measure laid out by Fueglistaler (2019) to understand the relevance of the SST change pattern to stability. We refer to the warm pool as the 30% warmest tropical SSTs, as is done for the SST# index. This SST# measure is related to stability, with warming in the warm pool (a deep convective region) tied to warming in the troposphere as warming propagates easily from the surface to the atmosphere. Warming

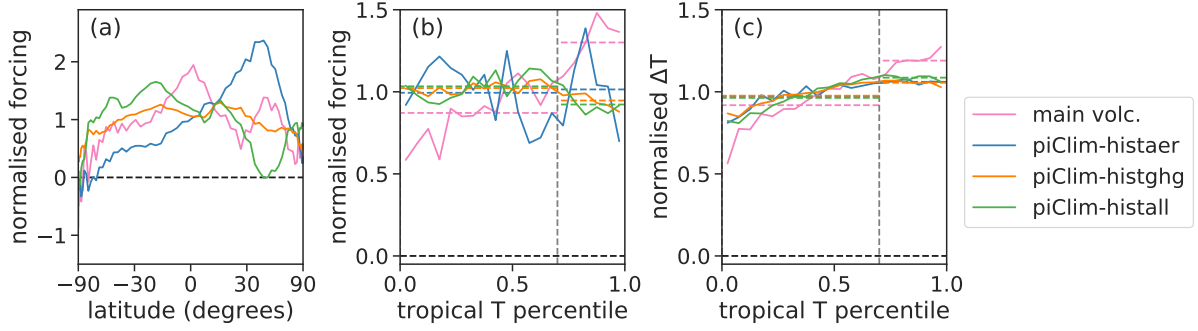


Figure 5.9: (a) Zonal patterns of forcing, normalised by global average forcing as in Fig. 4.1 but including forcing from the two main volcanic events in the 1850–2015 period. (b-c) Patterns of global average TOA effective forcing (b) and surface temperature change (c) in climatological tropical SST space. y -axis values are normalised for each experiment by the mean value over tropical SSTs i.e. such that all values would be unity for uniform tropical forcing/temperature change. Vertical grey dashed lines delimit the 30th percentile bounds as used in the SST# measure, whilst horizontal dashed lines denote average values on each side of the delimitation.

in less convective regions, by contrast, causes little change in tropospheric temperatures. Thus the difference between warming between the warm pool and elsewhere is indicative of tropospheric stability. Using SST# here therefore provides insight into why dS/dT is larger under selected volcanic forcing by showing that warm pool warming is significantly enhanced relative to average tropical SST warming. This is consistent with other work that has found similarly enhanced warm pool SST warming when looking at Pinatubo-style eruptions (Günther et al., 2022). Panel (d) seems to suggest that we should also see greater warm pool warming for AA than GHG forcing, but the lack of this may be down to the relative weakness of AA forcing over the tropics.

Apart from the forcing pattern, we propose another mechanism that may contribute to the significantly larger dS/dT for volcanic forcing than for GHG forcing. This mechanism is based on the short timescales on which volcanic forcing occurs, generally over only a few years. Such short timescales suggest that upper ocean heat capacity cannot be ignored as the upper ocean warms more quickly than if it has more time to transfer heat to the deeper ocean (Gregory et al., 2015). This is not likely an issue for historical AA and GHG forcings, since these evolve slowly over time, but in the volcanic case the forcings appear and disappear abruptly, lasting only a few years at most. Given varying effective

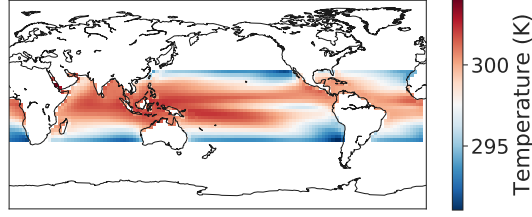


Figure 5.10: Map of climatological SSTs within the tropics, calculated from piControl experiments.

ocean depths across the globe, which implies varying ocean heat capacities, it is expected that forcing over very short timescales may result in a particular pattern of SST change. Depending on this pattern (i.e. if it correlates with climatological SSTs), stability changes per unit of temperature change may be larger on short timescales before substantial heat transfer occurs from the upper ocean downwards.

M. A. A. Rugenstein et al. (2016) provide some insight into what this pattern might look like (see their Fig. 4). For abrupt $4\times\text{CO}_2$ forcing using CMIP5 data, they find an SST change pattern that positively projects onto the climatological SST distribution within the tropics (see Fig. 5.10), with greater warming in the warm pool and less warming outside of deep convective regions. If the short timescale warming pattern found by M. A. A. Rugenstein et al. (2016) is mostly due to the abrupt nature of the forcing, then this could provide further explanation as to why strong short-lived volcanic events show greater dS/dT (and hence larger α) than other historical forcing agents.

5.7 Summary and conclusions

The findings of this chapter can be summarised as follows:

- In the MMM, historical radiative feedbacks vary significantly over time, driven by volcanic forcing and changing ratios of AA and GHG forcing.
- The amip-piForcing experiments show different α evolution than the historical experiments, possibly as coupled GCMs fail to recreate observed SST patterns.
- dS/dT explains variations in α , outside of unforced variability, in each experiment

as long as the forcing signal is sufficient and variations in dS/dT and α are above noise (natural variability).

- SW feedbacks, both in CRE and clearsky, play a significant role for net feedback differences among forcing agents. SW clearsky feedbacks differences are possibly due to water vapour and albedo feedbacks.
- Experiments with individual historical forcing agents explain the evolution of historical radiative feedbacks. Volcanic forcing dominates early on, while AA and GHGs both contribute in the middle period. All three forcings contribute in more recent years.
- Analysis of the strongest two eruptions of the 1850–2015 period (Pinatubo and Krakatoa) highlight very large dS/dT for volcanic eruptions, which results in lower climate sensitivity from their forcing.
- Large dS/dT for volcanic eruptions is tied to strong warm pool SST responses. This is linked to volcanic forcing being stronger over the warm pool. The short-term ocean response to forcing may also project more strongly onto the warm pool SST, compared to the longer-term response.

Findings here using CMIP6 data are broadly consistent with those found by Gregory et al. (2020) using CMIP5 data, with some exceptions. Significant time-variations in MMM historical feedbacks are reaffirmed, and GHG forcing is similarly shown to have little effect on decadal variations in feedbacks. That the time-evolution of feedbacks from historical forcing can be explained through its component forcings also reaffirms that the evolution of historical feedbacks is to some extent a result of evolving forcings, at least in the MMM. A main difference in findings is in the variations in the hist-nat case, which shows significantly more variation early on in CMIP6 than in CMIP5 data. Stronger measured variation may be tied to the knowledge of evolving effective forcing, from the CMIP6 piClim experiments, where Gregory et al. (2020) had to estimate this from aerosol optical depth data. Knowledge of the strong forcing from Pinatubo (1991) also leads to a different explanation of the final years of the historical simulation. Whereas Gregory et al. (2020) posited that GHG forcing purely dominated towards the end of the period, we

suggest that the less positive α for AA forcing and more positive α for volcanic forcing (both relative to GHG α) counteract each other such that the historical feedbacks are similar to GHG feedbacks for the final years of the simulation.

Our results are mostly consistent with Dong et al. (2021), who also used CMIP6 data, though we find more significant time-variation in feedbacks for the historical experiment, especially during the period of slowly varying T around 1940. This is likely related to our use of regressions on the MMM as opposed to Dong et al. taking the mean of model regressions, which we expect to result in regression dilution (Gregory et al., 2020), giving more amplifying feedbacks in Dong et al.’s method. We do not expect the inclusion/absence of data from GFDL-CM4’s historical experiment, which Dong et al. (2021) included but we did not, to explain the difference.

The results from our work augment the body of research that points to the SST pattern effect (Stevens et al., 2016), through stability and SW cloud feedbacks (Ceppi & Gregory, 2017, 2019; Kamae et al., 2019; Mackie et al., 2021), being related to radiative feedbacks. We further confirm the work of Ceppi and Gregory (2019) and Salvi et al. (2022) that points to a forcing agent dependence on radiative feedbacks, explained through this stability mechanism. We also affirm that the evolution of historical feedbacks, as well as the evolution of feedbacks for historical single-forcing agents, is explained by stability changes. This is a similar result to what has been found in works that highlight an evolution of feedbacks after the introduction of forcing, related to the pattern effect (Gregory & Andrews, 2016; Paynter et al., 2018; Rohrschneider et al., 2019; M. Rugenstein et al., 2020).

In determining the contributions of clouds, clearsky, SW, and LW to the feedback dependence between forcing agents, we find that SW cloud feedbacks are significant similar to the findings of Ceppi and Gregory (2019). Our findings disagree moderately with Ceppi and Gregory (2019), however, by suggesting that SW clearsky is also important for explaining the agent feedback dependence as tied to stability changes. This disagreement could be related to our simplistic breakdown into clearsky and CRE, as opposed to the

breakdown into cloud and non-cloud components that Ceppi and Gregory (2019) used via radiative kernels. Our analysis should also ideally be done with radiative kernels (Soden et al., 2008; Pendergrass et al., 2018) for better comparison. If the disagreement persists, this could indicate that the MMM behaviour is not the same as was found in the two models used by Ceppi and Gregory (2019).

More experiments would be helpful in confirming this, alongside analysis using other abrupt experiments such as abrupt-4xCO₂. These experiments could include brief step-forcing experiments in various patterns in order to identify how climatological SSTs correlate with SST changes under such forcings. This would inform on whether differences in forcing timescale are the main driver of differences in dS/dT between GHG and volcanic forcings, by disentangling the contributions of forcing patterns and forcing timescales from SST pattern changes.

The limited availability of volcanic forcing events in the historical period makes it unclear whether strong volcanic eruptions would generally cause high dS/dT or whether this is a quirk of the large events that have been observed. Further insight could be gained from prescribed forcing experiments that mimic volcanic eruptions, but for example with these eruptions set at different latitudes or times of the year (as done in Zhuo et al., 2021), or by simulating stronger versions, or larger ensembles, of historically weaker volcanic eruptions.

Further experiments involving historical forcing agents could also be useful. These could involve experiments that only include historical volcanic, solar, land use, or ozone forcings, which would fill in gaps of forcing contributions to the historical all-forcing case. Other experiments could be performed to provide a breakdown of AA forcing into different components such as historical BC and sulphate aerosols, since shifts in their relative contributions to historical forcings may be important for the historical evolution of feedbacks. For some of these, corresponding coupled AOGCM experiments already exist (e.g. hist-volc and hist-sol; Gillett et al., 2016) and as such only fixed-SST AGCM experiments would be needed in order to get the evolving adjustments required to calculate feedbacks.

Other agents may require both AOGCM and AGCM experiments.

Chapter 6

CONCLUSION

6.1 Summary of Thesis Achievements

The main goal of this project has been to improve our understanding of how radiative forcing drives climate change, through improved physical process understanding which can inform radiative forcing both within models. This work has demonstrated various mechanisms, all tied to atmospheric stability, that contribute to each aspect of the forcing–feedback–adjustment framework for understanding climate change. The key results from the analysis done here are as follows:

1. **Tropospheric stability is an important factor in determining radiative adjustments and radiative feedbacks due to forcing in the model world.**

Stability describes rapid cloud fraction changes unrelated to surface warming, with increasing stability tied to suppressed vertical mixing and thus increased low cloud formation. Conversely, decreased stability encourages convection and thus encourages formation of higher clouds whilst removing low clouds. Although models are limited in their representation of convection, they seem to capture this expected relationship between stability and clouds. This relates stability increases to negative (cooling) adjustments as increased low clouds reflect incoming SW, whilst stability decreases result in positive (warming) adjustments as some low cloud SW reflection is removed whilst increased high cloud increases LW trapping. Stability changes

are also tied to radiative feedbacks, driven by surface temperature change. For feedbacks, the pattern of surface warming drives stability changes that are related to changes in low cloud amount and thus mainly SW instead of LW CRE. Positive stability trends with surface temperature are tied to increased SW cooling from increased low cloud reflection of incoming solar radiation. We find that stability changes are not only linked to cloud feedbacks, but also clearsky feedbacks. This could be due to lapse rate feedbacks, which occur as increased stability leads to reduced lapse rates such that the greenhouse effect becomes weaker. Relative humidity and albedo feedbacks may also play a role here, though for unclear reasons as tied to stability changes.

2. **Stability explains a forcing agent dependence on adjustments.** Different forcing agents cause different radiative adjustments, tied to differences in how they affect atmospheric stability. Stability thus works as a way of explaining radiative adjustment differences between forcing agents. This is related to the vertical profile of atmospheric heating caused by given forcing agents, particularly the pressure centroid (i.e. average height) of where heating is applied. Forcing agents that cause heating focused at higher altitudes increase stability and thus positive cloud-radiative adjustments, whilst heating at lower altitudes causes negative cloud-radiative adjustments.

3. **Stability explains a forcing agent dependence on feedbacks.** This is tied to the warming patterns caused by different agents, related back to their forcing patterns and perhaps timescales of their forcing. We point to forcing patterns, with increased stability change and thus lower effective climate sensitivity resulting from agents that force more, and thus cause more warming, over convective regions whilst higher effective climate sensitivity results from agents that force more in non-convective regions.

6.2 Future Work

It is our hope that the findings in this thesis will be expanded upon, with future research to deal with the open questions raised here. The first of these questions relates to energy input in the earth system, as discussed in chapter 3. We have looked at vertical heating profiles within one model, but it would be interesting to see such an analysis using a variety of models. Furthermore, it is of interest to understand the contribution to radiative adjustments from the horizontal patterns of forcing. The analysis done here used horizontally uniform atmospheric heating, but stability is not globally uniform, and neither are the distributions of realistic forcing agents. Horizontally inhomogeneous atmospheric forcing may therefore reveal regions where forcing contributes more or less strongly to radiative adjustments.

Experiments with $\cos^2$ heating profiles centred at different levels in the atmosphere may be useful when done in AOGCMs, to provide insight into how the vertical structure of atmospheric heating contributes to radiative feedbacks instead of just adjustments. Given that our analysis of radiative feedbacks was centred on atmospheric stability as related to surface temperature change, such analysis of atmospheric heating may provide insight into contributions to stability change that have not been covered here. This work was originally intended, but was not performed due to a lack of access to the CESM model shortly after the experiments from chapter 4 had been completed.

Our approach of using prescribed forcing patterns (section 4.4) has shown its merits for understanding differences between tropical and extratropical forcing, and may be applicable to other forcing agents. Volcanic forcing, for example, remains difficult to analyse owing to significant forcing differences between eruptions. A prescribed forcing approach might be taken whereby an idealised forcing pattern is used that mimics forcing from different eruptions. This may elucidate surface warming pattern differences, and in turn radiative feedback differences, between eruptions. Such experiments, alongside other abrupt forcing experiments, may also provide insight into whether short forcing timescales

are a driver of warming patterns that lead to higher effective climate sensitivity.

While stability provides much physical insight into radiative feedbacks and adjustments, there does remain a potential issue in how stability is defined so that it accounts for the effects we wish to explain. Throughout this work we referred to several measures, from “bulk tropospheric stability” to near-global EIS, and also made use of indices that are related to stability such as SST#. Each of these measures accounts for stability in slightly different ways, and has been used in this thesis according to relevance, but a unified measure would be ideal. New works have looked at other measures of stability such as the “Estimated Cloud-Top Entrainment Index” (ECTEI) which may provide better insight into low cloud changes (Koshiro et al., 2022). Such an analysis of indices for stability remains as future work. Depending on how much these measures relate to cloud and clearsky radiative effects, such work may also provide better insight into which components are most important for describing the forcing agent dependence on radiative feedbacks. A kernel decomposition of radiative feedbacks, which would allow for radiative components to be broken down into contributions from clouds and clearsky components such as lapse rate, relative humidity, and albedo changes (Soden et al., 2008; Pendergrass et al., 2018), was absent here but would also likely be useful for this.

Though we have been successful in explaining much of the historical evolution of radiative feedbacks within models, using the main forcing agents of the period, additional accuracy could be afforded if the remaining forcing agents (land use change and ozone) could be accounted for. Hence, our hope is that future modelling efforts may include experiments with these individual historical forcing agents, so that the contribution from these can be accounted for in our analysis. Ideally, solar and volcanic forcing would also be disentangled with separate fixed SST experiments to be used with the pre-existing hist-volc and hist-sol coupled experiments (Gillett et al., 2016). Obtaining these will likely involve making proposals to CMIP for the next phase. Such experiments are likely to be helpful for dealing with the failure of models to recreate observed historical SSTs and feedbacks, by improving our understanding of the model world. Models still give us

our best (only) estimates of future warming, and the limited observation record may be unhelpful as an incomplete constraint on future warming. As such, pending many more years of observations which can help better constrain models, the future work outlined here should improve our understanding of the components that drive the evolution of historical climate change within models so that these can be more accurately corrected in the future as needed.

Appendix A: Why $\frac{X_{hist}}{\bar{T}_{hist}} - \frac{X_{GHG}}{\bar{T}_{GHG}}$ anti-correlates well with $\frac{X_{aer}}{\bar{T}_{aer}} - \frac{X_{GHG}}{\bar{T}_{GHG}}$

Historical forcing over the years 1850–2015 was mainly comprised of AA and GHG forcing (Forster et al., 2021). Though other forcings are significant at times within this historical period, they contribute little on average to the overall period. As such, we expect that the results from the historical case can be approximated as a sum of hist-GHG and hist-aer cases:

$$X_{hist} \approx X_{GHG} + X_{aer} \quad (\text{A.1})$$

$$\bar{T}_{hist} \approx \bar{T}_{GHG} + \bar{T}_{aer} \quad (\text{A.2})$$

Given that the global average surface air temperature changes $\bar{T}$ are, by definition, constants in latitude and longitude, we can relate $\bar{T}_{GHG}$ and $\bar{T}_{aer}$ by some function of time a :

$$\bar{T}_{aer} = a \cdot \bar{T}_{GHG} \quad (\text{A.3})$$

This allows us to rewrite the expression, for the difference between feedbacks in the historical case and the GHG case, in the following way:

$$\begin{aligned}
\frac{X_{hist}}{\bar{T}_{hist}} - \frac{X_{GHG}}{\bar{T}_{GHG}} &= \frac{X_{GHG} + X_{aer}}{\bar{T}_{GHG} + \bar{T}_{aer}} - \frac{X_{GHG}}{\bar{T}_{GHG}} \\
&= \frac{X_{GHG}}{\bar{T}_{GHG} + \bar{T}_{aer}} + \frac{X_{aer}}{\bar{T}_{GHG} + \bar{T}_{aer}} - \frac{X_{GHG}}{\bar{T}_{GHG}} \\
&= \frac{X_{GHG}}{(a+1) \cdot \bar{T}_{GHG}} + \frac{X_{aer}}{(\frac{1}{a} + 1) \cdot \bar{T}_{aer}} - \frac{X_{GHG}}{\bar{T}_{GHG}} \\
&= \frac{X_{GHG}}{(a+1) \cdot \bar{T}_{GHG}} + \frac{a \cdot X_{aer}}{(a+1) \cdot \bar{T}_{aer}} - \frac{X_{GHG}}{\bar{T}_{GHG}} \\
&= \frac{a \cdot X_{aer}}{(a+1)\bar{T}_{aer}} - \frac{a \cdot X_{GHG}}{(a+1) \cdot \bar{T}_{GHG}} \\
&= \frac{a}{(a+1)} \left(\frac{X_{aer}}{\bar{T}_{aer}} - \frac{X_{GHG}}{\bar{T}_{GHG}} \right)
\end{aligned}$$

We expect that $a < 0$ for all points in time, since we expect aerosol to reduce surface air temperatures where GHGs increase surface air temperatures. We also generally expect that $|a| < 1$ since the historical case gives rising temperatures i.e. the aerosol forcing does not outweigh the GHG forcing. As such, we expect that $\frac{a}{a+1} < 0$ to provide the observed anti-correlation.

Bibliography

- Allen, R. J., Amiri-Farahani, A., Lamarque, J.-F., Smith, C., Shindell, D., Hassan, T., & Chung, C. E. (2019, December). Observationally constrained aerosol–cloud semi-direct effects. *npj Climate and Atmospheric Science*, 2(1), 16. Retrieved 2020-02-04, from <http://www.nature.com/articles/s41612-019-0073-9> doi: 10.1038/s41612-019-0073-9
- Amiri-Farahani, A., Allen, R. J., Neubauer, D., & Lohmann, U. (2017, May). Impact of Saharan dust on North Atlantic marine stratocumulus clouds: importance of the semidirect effect. *Atmospheric Chemistry and Physics*, 17(10), 6305–6322. Retrieved 2021-01-25, from <https://acp.copernicus.org/articles/17/6305/2017/> (Publisher: Copernicus GmbH) doi: <https://doi.org/10.5194/acp-17-6305-2017>
- Ammann, C. M., Meehl, G. A., Washington, W. M., & Zender, C. S. (2003). A monthly and latitudinally varying volcanic forcing dataset in simulations of 20th century climate. *Geophysical Research Letters*, 30(12). Retrieved 2022-06-07, from <https://onlinelibrary.wiley.com/doi/abs/10.1029/2003GL016875> (_eprint: <https://onlinelibrary.wiley.com/doi/pdf/10.1029/2003GL016875>) doi: 10.1029/2003GL016875
- Andrews, T., Bodas-Salcedo, A., Gregory, J. M., Dong, Y., Armour, K. C., Paynter, D., ... Liu, C. (2022). On the Effect of Historical SST Patterns on Radiative Feedback. *Journal of Geophysical Research: Atmospheres*, 127(18), e2022JD036675. Retrieved 2022-10-21, from <https://onlinelibrary.wiley.com/doi/abs/10.1029/2022JD036675> (_eprint: <https://onlinelibrary.wiley.com/doi/pdf/10.1029/2022JD036675>) doi: 10.1029/2022JD036675
- Andrews, T., & Forster, P. M. (2008). CO₂ forcing induces semi-direct effects with consequences for climate feedback interpretations. *Geophysical Research Letters*, 35(4). Retrieved 2021-01-12, from <https://agupubs.onlinelibrary.wiley.com/doi/abs/10.1029/2007GL032273> (_eprint: <https://agupubs.onlinelibrary.wiley.com/doi/pdf/10.1029/2007GL032273>) doi: <https://doi.org/10.1029/2007GL032273>
- Andrews, T., Gregory, J. M., Forster, P. M., & Webb, M. J. (2012). Cloud Adjustment and its Role in CO₂ Radiative Forcing and Climate Sensitivity: A Review. *Surveys in Geophysics*, 33(3), 619–635. Retrieved 2019-07-15, from <https://doi.org/10.1007/s10712-011-9152-0> doi: 10.1007/s10712-011-9152-0
- Andrews, T., Gregory, J. M., Paynter, D., Silvers, L. G., Zhou, C., Mauritsen, T., ... Titchner, H. (2018). Accounting for Changing Temperature Patterns Increases Historical Estimates of Climate Sensitivity. *Geophysical Research Letters*, 45(16),

- 8490–8499. Retrieved 2019-07-15, from <https://agupubs.onlinelibrary.wiley.com/doi/abs/10.1029/2018GL078887> doi: 10.1029/2018GL078887
- Andrews, T., Gregory, J. M., & Webb, M. J. (2015, February). The Dependence of Radiative Forcing and Feedback on Evolving Patterns of Surface Temperature Change in Climate Models. *Journal of Climate*, 28(4), 1630–1648. Retrieved 2021-12-09, from <https://journals.ametsoc.org/view/journals/clim/28/4/jcli-d-14-00545.1.xml> (Publisher: American Meteorological Society Section: Journal of Climate) doi: 10.1175/JCLI-D-14-00545.1
- Andrews, T., Gregory, J. M., Webb, M. J., & Taylor, K. E. (2012). Forcing, feedbacks and climate sensitivity in CMIP5 coupled atmosphere-ocean climate models. *Geophysical Research Letters*, 39(9). Retrieved 2019-07-23, from <https://agupubs.onlinelibrary.wiley.com/doi/abs/10.1029/2012GL051607> doi: 10.1029/2012GL051607
- Andrews, T., Smith, C. J., Myhre, G., Forster, P. M., Chadwick, R., & Ackerley, D. (2021). Effective Radiative Forcing in a GCM With Fixed Surface Temperatures. *Journal of Geophysical Research: Atmospheres*, 126(4), e2020JD033880. Retrieved 2021-10-29, from <https://onlinelibrary.wiley.com/doi/abs/10.1029/2020JD033880> (_eprint: <https://onlinelibrary.wiley.com/doi/pdf/10.1029/2020JD033880>) doi: 10.1029/2020JD033880
- Andrews, T., & Webb, M. J. (2018, January). The Dependence of Global Cloud and Laplace Rate Feedbacks on the Spatial Structure of Tropical Pacific Warming. *Journal of Climate*, 31(2), 641–654. Retrieved 2021-10-12, from <https://journals.ametsoc.org/view/journals/clim/31/2/jcli-d-17-0087.1.xml> (Publisher: American Meteorological Society Section: Journal of Climate) doi: 10.1175/JCLI-D-17-0087.1
- Arfeuille, F., Weisenstein, D., Mack, H., Rozanov, E., Peter, T., & Brönnimann, S. (2014, February). Volcanic forcing for climate modeling: a new microphysics-based data set covering years 1600–present. *Climate of the Past*, 10(1), 359–375. Retrieved 2022-12-12, from <https://cp.copernicus.org/articles/10/359/2014/> doi: 10.5194/cp-10-359-2014
- Armour, K. C., Bitz, C. M., & Roe, G. H. (2013, July). Time-Varying Climate Sensitivity from Regional Feedbacks. *Journal of Climate*, 26(13), 4518–4534. Retrieved 2022-06-16, from <https://journals.ametsoc.org/view/journals/clim/26/13/jcli-d-12-00544.1.xml> (Publisher: American Meteorological Society Section: Journal of Climate) doi: 10.1175/JCLI-D-12-00544.1
- Ban-Weiss, G. A., Cao, L., Bala, G., & Caldeira, K. (2012, March). Dependence of climate forcing and response on the altitude of black carbon aerosols. *Climate Dynamics*, 38(5), 897–911. Retrieved 2020-10-21, from <https://doi.org/10.1007/s00382-011-1052-y> doi: 10.1007/s00382-011-1052-y
- Becker, T., & Wing, A. A. (2020). Understanding the Extreme Spread in Climate Sensitivity within the Radiative-Convective Equilibrium Model Intercomparison Project. *Journal of Advances in Modeling Earth Systems*, 12(10), e2020MS002165. Retrieved 2020-11-11, from <https://agupubs.onlinelibrary.wiley.com/doi/abs/10.1029/2020MS002165> (_eprint: <https://agupubs.onlinelibrary.wiley.com/doi/pdf/10.1029/2020MS002165>) doi: <https://doi.org/10.1029/2020MS002165>
- Bellouin, N., Quaas, J., Gryspeerdt, E., Kinne, S., Stier, P., Watson-

- Parris, D., ... Stevens, B. (2020). Bounding Global Aerosol Radiative Forcing of Climate Change. *Reviews of Geophysics*, 58(1), e2019RG000660. Retrieved 2021-02-01, from <https://agupubs.onlinelibrary.wiley.com/doi/abs/10.1029/2019RG000660> (_eprint: <https://agupubs.onlinelibrary.wiley.com/doi/pdf/10.1029/2019RG000660>) doi: <https://doi.org/10.1029/2019RG000660>
- Beucler, T., Cronin, T., & Emanuel, K. (2018, August). A Linear Response Framework for Radiative-Convective Instability. *Journal of Advances in Modeling Earth Systems*, 10(8), 1924–1951. Retrieved 2021-06-17, from <http://doi.wiley.com/10.1029/2018MS001280> doi: 10.1029/2018MS001280
- Bloch-Johnson, J., Rugenstein, M., Stolpe, M. B., Rohrschneider, T., Zheng, Y., & Gregory, J. M. (2021). Climate Sensitivity Increases Under Higher CO₂ Levels Due to Feedback Temperature Dependence. *Geophysical Research Letters*, 48(4), e2020GL089074. Retrieved 2022-06-16, from <https://onlinelibrary.wiley.com/doi/abs/10.1029/2020GL089074> (_eprint: <https://onlinelibrary.wiley.com/doi/pdf/10.1029/2020GL089074>) doi: 10.1029/2020GL089074
- Boucher, O., Randall, D., Artaxo, P., Bretherton, C., Feingold, G., Forster, P., ... Zhang, X.-Y. (2013). Clouds and Aerosols. *IPCC, 2013: Climate Change 2013: The Physical Science Basis. Contribution of Working Group I to the Fifth Assessment Report of the Intergovernmental Panel on Climate Change*. Retrieved from https://www.ipcc.ch/site/assets/uploads/2018/02/WG1AR5_Chapter07_FINAL-1.pdf
- Brunner, L., Pendergrass, A. G., Lehner, F., Merrifield, A. L., Lorenz, R., & Knutti, R. (2020, November). Reduced global warming from CMIP6 projections when weighting models by performance and independence. *Earth System Dynamics*, 11(4), 995–1012. Retrieved 2022-05-09, from <https://esd.copernicus.org/articles/11/995/2020/> (Publisher: Copernicus GmbH) doi: 10.5194/esd-11-995-2020
- Caldwell, P. M., Zelinka, M. D., Taylor, K. E., & Marvel, K. (2016, January). Quantifying the Sources of Intermodel Spread in Equilibrium Climate Sensitivity. *Journal of Climate*, 29(2), 513–524. Retrieved 2022-07-06, from <https://journals.ametsoc.org/view/journals/clim/29/2/jcli-d-15-0352.1.xml> (Publisher: American Meteorological Society Section: Journal of Climate) doi: 10.1175/JCLI-D-15-0352.1
- Ceppi, P., Brient, F., Zelinka, M. D., & Hartmann, D. L. (2017). Cloud feedback mechanisms and their representation in global climate models. *Wiley Interdisciplinary Reviews: Climate Change*, 8(4), e465. Retrieved 2019-07-15, from <https://onlinelibrary.wiley.com/doi/abs/10.1002/wcc.465> doi: 10.1002/wcc.465
- Ceppi, P., & Gregory, J. M. (2017, December). Relationship of tropospheric stability to climate sensitivity and Earth’s observed radiation budget. *Proceedings of the National Academy of Sciences*, 114(50), 13126–13131. Retrieved 2019-07-15, from <https://www.pnas.org/content/114/50/13126> doi: 10.1073/pnas.1714308114
- Ceppi, P., & Gregory, J. M. (2019, October). A refined model for the Earth’s global energy balance. *Climate Dynamics*, 53(7), 4781–4797. Retrieved 2020-12-03, from <https://doi.org/10.1007/s00382-019-04825-x> doi: 10.1007/s00382-019-04825-x
- Charlson, R. J., Schwartz, S. E., Hales, J. M., Cess, R. D., Coakley, J. A., Hansen, J. E., & Hofmann, D. J. (1992, January). Climate Forcing by Anthropogenic Aerosols. *Science*, 255(5043), 423–430. Retrieved 2022-12-07, from <https://www>

- .science.org/doi/abs/10.1126/science.255.5043.423 (Publisher: American Association for the Advancement of Science) doi: 10.1126/science.255.5043.423
- Charney, J. G., Arakawa, A., Baker, D. J., Bolin, B., Dickinson, R. E., Goody, R. M., ... Wunsch, C. I. (1979). *Carbon dioxide and climate: a scientific assessment*. National Academy of Sciences, Washington, DC.
- Chung, E.-S., & Soden, B. J. (2015, March). An Assessment of Direct Radiative Forcing, Radiative Adjustments, and Radiative Feedbacks in Coupled Ocean–Atmosphere Models. *Journal of Climate*, 28(10), 4152–4170. Retrieved 2019-07-15, from <https://journals.ametsoc.org/doi/full/10.1175/JCLI-D-14-00436.1> doi: 10.1175/JCLI-D-14-00436.1
- Collins, W. D. (2001, November). Parameterization of Generalized Cloud Overlap for Radiative Calculations in General Circulation Models. *Journal of the Atmospheric Sciences*, 58(21), 3224–3242. Retrieved 2023-02-08, from https://journals.ametsoc.org/view/journals/atasc/58/21/1520-0469_2001_058_3224_pogcof_2.0.co_2.xml (Publisher: American Meteorological Society Section: Journal of the Atmospheric Sciences) doi: 10.1175/1520-0469(2001)058<3224:POGCOF>2.0.CO;2
- Colman, R. A., & Hanson, L. (2017, September). On the relative strength of radiative feedbacks under climate variability and change. *Climate Dynamics*, 49(5), 2115–2129. Retrieved 2022-07-06, from <https://doi.org/10.1007/s00382-016-3441-8> doi: 10.1007/s00382-016-3441-8
- Colman, R. A., & McAvaney, B. J. (2011, April). On tropospheric adjustment to forcing and climate feedbacks. *Climate Dynamics*, 36(9), 1649. Retrieved 2019-09-20, from <https://doi.org/10.1007/s00382-011-1067-4> doi: 10.1007/s00382-011-1067-4
- Conley, A. J., Lamarque, J.-F., Vitt, F., Collins, W. D., & Kiehl, J. (2013, April). PORT, a CESM tool for the diagnosis of radiative forcing. *Geoscientific Model Development*, 6(2), 469–476. Retrieved 2019-07-15, from <https://www.geoscientific-model-dev.net/6/469/2013/> doi: <https://doi.org/10.5194/gmd-6-469-2013>
- Cook, J., Oreskes, N., Doran, P. T., Anderegg, W. R. L., Verheggen, B., Maibach, E. W., ... Rice, K. (2016, April). Consensus on consensus: a synthesis of consensus estimates on human-caused global warming. *Environmental Research Letters*, 11(4), 048002. Retrieved 2019-07-23, from <https://doi.org/10.1088/1748-9326/11/4/048002> doi: 10.1088/1748-9326/11/4/048002
- Dinh, T., & Fueglistaler, S. (2017). Mechanism of Fast Atmospheric Energetic Equilibration Following Radiative Forcing by CO₂. *Journal of Advances in Modeling Earth Systems*, 9(7), 2468–2482. Retrieved 2019-07-15, from <https://agupubs.onlinelibrary.wiley.com/doi/abs/10.1002/2017MS001116> doi: 10.1002/2017MS001116
- Dong, Y., Armour, K. C., Proistosescu, C., Andrews, T., Battisti, D. S., Forster, P. M., ... Shiogama, H. (2021). Biased Estimates of Equilibrium Climate Sensitivity and Transient Climate Response Derived From Historical CMIP6 Simulations. *Geophysical Research Letters*, 48(24), e2021GL095778. Retrieved 2022-05-13, from <https://onlinelibrary.wiley.com/doi/abs/10.1029/2021GL095778> (_eprint: <https://onlinelibrary.wiley.com/doi/pdf/10.1029/2021GL095778>) doi: 10.1029/2021GL095778
- Dong, Y., Armour, K. C., Zelinka, M. D., Proistosescu, C., Battisti, D. S., Zhou, C., & Andrews, T. (2020, September). Intermodel Spread in the Pattern Effect and Its Con-

- tribution to Climate Sensitivity in CMIP5 and CMIP6 Models. *Journal of Climate*, 33(18), 7755–7775. Retrieved 2022-06-16, from <https://journals.ametsoc.org/view/journals/clim/33/18/jcliD191011.xml> (Publisher: American Meteorological Society Section: Journal of Climate) doi: 10.1175/JCLI-D-19-1011.1
- Dong, Y., Proistosescu, C., Armour, K. C., & Battisti, D. S. (2019, September). Attributing Historical and Future Evolution of Radiative Feedbacks to Regional Warming Patterns using a Green’s Function Approach: The Preeminence of the Western Pacific. *Journal of Climate*, 32(17), 5471–5491. Retrieved 2021-06-17, from <https://journals.ametsoc.org/view/journals/clim/32/17/jcli-d-18-0843.1.xml> (Publisher: American Meteorological Society Section: Journal of Climate) doi: 10.1175/JCLI-D-18-0843.1
- Doutriaux-Boucher, M., Webb, M. J., Gregory, J. M., & Boucher, O. (2009). Carbon dioxide induced stomatal closure increases radiative forcing via a rapid reduction in low cloud. *Geophysical Research Letters*, 36(2). Retrieved 2019-09-20, from <https://agupubs.onlinelibrary.wiley.com/doi/abs/10.1029/2008GL036273> doi: 10.1029/2008GL036273
- Durack, P. J., & Taylor, K. E. (2022, June). *CMIP6 Forcing Datasets Summary v6.2.43*. Retrieved from <http://goo.gl/r8up31>
- Etheridge, D. M., D.M., & Steele, L. P. (1996, February). Natural and anthropogenic changes in atmospheric CO₂ over the last 1000 years from air in Antarctic ice and firn. *Journal of Geophysical Research*, 14.
- Eyring, V., Bony, S., Meehl, G. A., Senior, C. A., Stevens, B., Stouffer, R. J., & Taylor, K. E. (2016, May). Overview of the Coupled Model Intercomparison Project Phase 6 (CMIP6) experimental design and organization. *Geoscientific Model Development*, 9(5), 1937–1958. Retrieved 2022-06-28, from <https://gmd.copernicus.org/articles/9/1937/2016/> doi: 10.5194/gmd-9-1937-2016
- Flannaghan, T. J., Fueglistaler, S., Held, I. M., Po-Chedley, S., Wyman, B., & Zhao, M. (2014). Tropical temperature trends in Atmospheric General Circulation Model simulations and the impact of uncertainties in observed SSTs. *Journal of Geophysical Research: Atmospheres*, 119(23), 13,327–13,337. Retrieved 2021-10-12, from <https://onlinelibrary.wiley.com/doi/abs/10.1002/2014JD022365> (_eprint: <https://onlinelibrary.wiley.com/doi/pdf/10.1002/2014JD022365>) doi: 10.1002/2014JD022365
- Forster, P. M., Richardson, T., Maycock, A. C., Smith, C. J., Samset, B. H., Myhre, G., ... Schulz, M. (2016). Recommendations for diagnosing effective radiative forcing from climate models for CMIP6. *Journal of Geophysical Research: Atmospheres*, 121(20), 12,460–12,475. Retrieved 2020-01-21, from <https://agupubs.onlinelibrary.wiley.com/doi/abs/10.1002/2016JD025320> doi: 10.1002/2016JD025320
- Forster, P. M., Storelvmo, T., Armour, K., Collins, W., Dufresne, J., Frame, D., ... Zhang, H. (2021). The Earth’s Energy Budget, Climate Feedbacks, and Climate Sensitivity. *Climate Change 2021: The Physical Science Basis. Contribution of Working Group I to the Sixth Assessment Report of the Intergovernmental Panel on Climate Change*. Retrieved from https://www.ipcc.ch/report/ar6/wg1/downloads/report/IPCC_AR6_WGI_Chapter07.pdf
- Fueglistaler, S. (2019). Observational Evidence for Two Modes of Coupling Between Sea Surface Temperatures, Tropospheric Temperature Profile, and Shortwave Cloud Radiative Effect in the Tropics. *Geophysical Re-*

- search Letters*, 46(16), 9890–9898. Retrieved 2021-05-26, from <https://agupubs.onlinelibrary.wiley.com/doi/abs/10.1029/2019GL083990> (_eprint: <https://agupubs.onlinelibrary.wiley.com/doi/pdf/10.1029/2019GL083990>) doi: <https://doi.org/10.1029/2019GL083990>
- Fueglistaler, S., & Silvers, L. G. (2021). The Peculiar Trajectory of Global Warming. *Journal of Geophysical Research: Atmospheres*, 126(4), e2020JD033629. Retrieved 2021-05-26, from <https://agupubs.onlinelibrary.wiley.com/doi/abs/10.1029/2020JD033629> (_eprint: <https://agupubs.onlinelibrary.wiley.com/doi/pdf/10.1029/2020JD033629>) doi: <https://doi.org/10.1029/2020JD033629>
- Gillett, N. P., Shiogama, H., Funke, B., Hegerl, G., Knutti, R., Matthes, K., ... Tebaldi, C. (2016, April). *Detection and Attribution Model Intercomparison Project (DAMIP)* (preprint). Climate and Earth System Modeling. Retrieved 2022-02-04, from <https://gmd.copernicus.org/preprints/gmd-2016-74/gmd-2016-74.pdf> doi: 10.5194/gmd-2016-74
- Gregory, J. M., & Andrews, T. (2016). Variation in climate sensitivity and feedback parameters during the historical period. *Geophysical Research Letters*, 43(8), 3911–3920. Retrieved 2021-02-22, from <https://agupubs.onlinelibrary.wiley.com/doi/abs/10.1002/2016GL068406> (_eprint: <https://agupubs.onlinelibrary.wiley.com/doi/pdf/10.1002/2016GL068406>) doi: <https://doi.org/10.1002/2016GL068406>
- Gregory, J. M., Andrews, T., Ceppi, P., Mauritsen, T., & Webb, M. J. (2020, January). How accurately can the climate sensitivity to CO₂ be estimated from historical climate change? *Climate Dynamics*, 54(1), 129–157. Retrieved 2021-02-22, from <https://doi.org/10.1007/s00382-019-04991-y> doi: 10.1007/s00382-019-04991-y
- Gregory, J. M., Andrews, T., & Good, P. (2015, November). The inconstancy of the transient climate response parameter under increasing CO₂. *Philosophical Transactions of the Royal Society A: Mathematical, Physical and Engineering Sciences*, 373(2054), 20140417. Retrieved 2022-06-10, from <https://royalsocietypublishing.org/doi/full/10.1098/rsta.2014.0417> (Publisher: Royal Society) doi: 10.1098/rsta.2014.0417
- Gregory, J. M., Andrews, T., Good, P., Mauritsen, T., & Forster, P. M. (2016, December). Small global-mean cooling due to volcanic radiative forcing. *Climate Dynamics*, 47(12), 3979–3991. Retrieved 2022-04-22, from <http://link.springer.com/10.1007/s00382-016-3055-1> doi: 10.1007/s00382-016-3055-1
- Gregory, J. M., Bi, D., Collier, M., Dix, M., Hirst, A., Hu, A., ... Church, J. (2013). Climate models without preindustrial volcanic forcing underestimate historical ocean thermal expansion. *Geophysical Research Letters*, 40(8), 1600–1604. Retrieved 2022-07-26, from <https://onlinelibrary.wiley.com/doi/abs/10.1002/grl.50339> (_eprint: <https://onlinelibrary.wiley.com/doi/pdf/10.1002/grl.50339>) doi: 10.1002/grl.50339
- Gregory, J. M., Ingram, W. J., Palmer, M. A., Jones, G. S., Stott, P. A., Thorpe, R. B., ... Williams, K. D. (2004). A new method for diagnosing radiative forcing and climate sensitivity. *Geophysical Research Letters*, 31(3). Retrieved 2019-07-23, from <https://agupubs.onlinelibrary.wiley.com/doi/abs/10.1029/2003GL018747> doi: 10.1029/2003GL018747
- Gregory, J. M., & Webb, M. (2008, January). Tropospheric Adjustment Induces a Cloud

- Component in CO₂ Forcing. *Journal of Climate*, 21(1), 58–71. Retrieved 2019-07-15, from <https://journals.ametsoc.org/doi/full/10.1175/2007JCLI1834.1> doi: 10.1175/2007JCLI1834.1
- Günther, M., Schmidt, H., Timmreck, C., & Toohey, M. (2022, November). Climate Feedback to Stratospheric Aerosol Forcing: The Key Role of the Pattern Effect. *Journal of Climate*, 35(24), 4303–4317. Retrieved 2022-12-09, from <https://journals.ametsoc.org/view/journals/clim/35/24/JCLI-D-22-0306.1.xml> (Publisher: American Meteorological Society Section: Journal of Climate) doi: 10.1175/JCLI-D-22-0306.1
- Hansen, J., Sato, M., Nazarenko, L., Ruedy, R., Lacis, A., Koch, D., ... Christy, J. (2002). Climate forcings in Goddard Institute for Space Studies SI2000 simulations. *Journal of Geophysical Research: Atmospheres*, 107(D18), ACL 2–1–ACL 2–37. Retrieved 2019-07-25, from <https://agupubs.onlinelibrary.wiley.com/doi/abs/10.1029/2001JD001143> doi: 10.1029/2001JD001143
- Hansen, J., Sato, M., & Ruedy, R. (1997). Radiative forcing and climate response. *Journal of Geophysical Research: Atmospheres*, 102(D6), 6831–6864. Retrieved 2019-08-09, from <https://agupubs.onlinelibrary.wiley.com/doi/abs/10.1029/96JD03436> doi: 10.1029/96JD03436
- Hartmann, D. L. (2016). *Global physical climatology* (2nd ed.). Elsevier.
- Jones, T. R., & Randall, D. A. (2011). Quantifying the limits of convective parameterizations. *Journal of Geophysical Research: Atmospheres*, 116(D8). Retrieved 2023-01-09, from <https://onlinelibrary.wiley.com/doi/abs/10.1029/2010JD014913> (_eprint: <https://onlinelibrary.wiley.com/doi/pdf/10.1029/2010JD014913>) doi: 10.1029/2010JD014913
- Kamae, Y., Chadwick, R., Ackerley, D., Ringer, M., & Ogura, T. (2019, May). Seasonally variant low cloud adjustment over cool oceans. *Climate Dynamics*, 52(9-10), 5801–5817. Retrieved 2019-09-16, from <http://link.springer.com/10.1007/s00382-018-4478-7> doi: 10.1007/s00382-018-4478-7
- Kamae, Y., & Watanabe, M. (2012). On the robustness of tropospheric adjustment in CMIP5 models. *Geophysical Research Letters*, 39(23). Retrieved 2020-04-28, from <https://agupubs.onlinelibrary.wiley.com/doi/abs/10.1029/2012GL054275> (_eprint: <https://agupubs.onlinelibrary.wiley.com/doi/pdf/10.1029/2012GL054275>) doi: 10.1029/2012GL054275
- Kamae, Y., & Watanabe, M. (2013, December). Tropospheric adjustment to increasing CO₂: its timescale and the role of land–sea contrast. *Climate Dynamics*, 41(11-12), 3007–3024. Retrieved 2019-07-04, from <http://link.springer.com/10.1007/s00382-012-1555-1> doi: 10.1007/s00382-012-1555-1
- Kamae, Y., Watanabe, M., Ogura, T., Yoshimori, M., & Shiogama, H. (2015, June). Rapid Adjustments of Cloud and Hydrological Cycle to Increasing CO₂: a Review. *Current Climate Change Reports*, 1(2), 103–113. Retrieved 2019-07-04, from <http://link.springer.com/10.1007/s40641-015-0007-5> doi: 10.1007/s40641-015-0007-5
- Klein, S. A., & Hartmann, D. L. (1993, August). The Seasonal Cycle of Low Stratiform Clouds. *Journal of Climate*, 6(8), 1587–1606. Retrieved 2020-12-18, from https://journals.ametsoc.org/view/journals/clim/6/8/1520-0442_1993_006_1587_tscols_2_0_co_2.xml (Publisher: American Meteorological Society Section: Journal of Climate) doi: 10.1175/1520-0442(1993)006<1587:

TSCOLS>2.0.CO;2

- Knutti, R., & Rugenstein, M. A. A. (2015, November). Feedbacks, climate sensitivity and the limits of linear models. *Philosophical Transactions of the Royal Society A: Mathematical, Physical and Engineering Sciences*, 373(2054), 20150146. Retrieved 2019-07-15, from <https://royalsocietypublishing.org/doi/full/10.1098/rsta.2015.0146> doi: 10.1098/rsta.2015.0146
- Knutti, R., Rugenstein, M. A. A., & Hegerl, G. C. (2017, October). Beyond equilibrium climate sensitivity. *Nature Geoscience*, 10(10), 727–736. Retrieved 2019-07-15, from <https://www.nature.com/articles/ngeo3017> doi: 10.1038/ngeo3017
- Koch, D., & Del Genio, A. D. (2010, August). Black carbon semi-direct effects on cloud cover: review and synthesis. *Atmospheric Chemistry and Physics*, 10(16), 7685–7696. Retrieved 2021-01-25, from <https://acp.copernicus.org/articles/10/7685/2010/> doi: 10.5194/acp-10-7685-2010
- Koshiro, T., Kawai, H., & Noda, A. T. (2022, July). Estimated cloud-top entrainment index explains positive low-cloud-cover feedback. *Proceedings of the National Academy of Sciences*, 119(29), e2200635119. Retrieved 2022-07-28, from <https://www.pnas.org/doi/abs/10.1073/pnas.2200635119> (Publisher: Proceedings of the National Academy of Sciences) doi: 10.1073/pnas.2200635119
- Kuang, Z. (2010, April). Linear Response Functions of a Cumulus Ensemble to Temperature and Moisture Perturbations and Implications for the Dynamics of Convectively Coupled Waves. *Journal of the Atmospheric Sciences*, 67(4), 941–962. Retrieved 2021-06-17, from <https://journals.ametsoc.org/view/journals/atsc/67/4/2009jas3260.1.xml> (Publisher: American Meteorological Society Section: Journal of the Atmospheric Sciences) doi: 10.1175/2009JAS3260.1
- Lamarque, J.-F., Bond, T. C., Eyring, V., Granier, C., Heil, A., Klimont, Z., ... van Vuuren, D. P. (2010, August). Historical (1850–2000) gridded anthropogenic and biomass burning emissions of reactive gases and aerosols: methodology and application. *Atmospheric Chemistry and Physics*, 10(15), 7017–7039. Retrieved 2022-06-24, from <https://acp.copernicus.org/articles/10/7017/2010/> doi: 10.5194/acp-10-7017-2010
- Mackie, A., Brindley, H. E., & Palmer, P. I. (2021). Contrasting Observed Atmospheric Responses to Tropical Sea Surface Temperature Warming Patterns. *Journal of Geophysical Research: Atmospheres*, 126(7), e2020JD033564. Retrieved 2022-05-19, from <https://onlinelibrary.wiley.com/doi/abs/10.1029/2020JD033564> (_eprint: <https://onlinelibrary.wiley.com/doi/pdf/10.1029/2020JD033564>) doi: 10.1029/2020JD033564
- Mandl, F. (2013). *Statistical Physics* (2nd ed.). Chichester, UNITED KINGDOM: John Wiley & Sons, Incorporated. Retrieved 2019-08-09, from <http://ebookcentral.proquest.com/lib/imperial/detail.action?docID=1211897>
- Marvel, K., Schmidt, G. A., Miller, R. L., & Nazarenko, L. S. (2016, April). Implications for climate sensitivity from the response to individual forcings. *Nature Climate Change*, 6(4), 386–389. Retrieved 2021-10-08, from <https://www.nature.com/articles/nclimate2888> (Bandiera_abtest: a Cg_type: Nature Research Journals Number: 4 Primary_atype: Research Publisher: Nature Publishing Group Subject_term: Climate and Earth system modelling;Projection and prediction Subject_term_id: climate-and-earth-system-modelling;projection-and-prediction) doi: 10.1038/nclimate2888
- Mass, C. F., & Portman, D. A. (1989, June). Major Volcanic Eruptions and Climate:

- A Critical Evaluation. *Journal of Climate*, 2(6), 566–593. Retrieved 2022-06-08, from https://journals.ametsoc.org/view/journals/clim/2/6/1520-0442_1989_002_0566_mveaca_2_0_co_2.xml (Publisher: American Meteorological Society Section: Journal of Climate) doi: 10.1175/1520-0442(1989)002<0566:MVEACA>2.0.CO;2
- Meraner, K., Mauritsen, T., & Voigt, A. (2013). Robust increase in equilibrium climate sensitivity under global warming. *Geophysical Research Letters*, 40(22), 5944–5948. Retrieved 2022-09-22, from <https://onlinelibrary.wiley.com/doi/abs/10.1002/2013GL058118> (_eprint: <https://onlinelibrary.wiley.com/doi/pdf/10.1002/2013GL058118>) doi: 10.1002/2013GL058118
- Murphy, J. M. (1995, March). Transient Response of the Hadley Centre Coupled Ocean-Atmosphere Model to Increasing Carbon Dioxide. Part III: Analysis of Global-Mean Response Using Simple Models. *Journal of Climate*, 8(3), 496–514. Retrieved 2022-08-03, from https://journals.ametsoc.org/view/journals/clim/8/3/1520-0442_1995_008_0496_trothc_2_0_co_2.xml (Publisher: American Meteorological Society Section: Journal of Climate) doi: 10.1175/1520-0442(1995)008<0496:TROTHC>2.0.CO;2
- Myhre, G., Forster, P. M., Samset, B. H., Hodnebrog, O., Sillmann, J., Aalbergstjø, S. G., ... Zwiers, F. (2017, June). PDRMIP A Precipitation Driver and Response Model Intercomparison Project—Protocol and Preliminary Results. *Bulletin of the American Meteorological Society*, 98(6), 1185–1198. Retrieved 2019-08-01, from <http://journals.ametsoc.org/doi/10.1175/BAMS-D-16-0019.1> doi: 10.1175/BAMS-D-16-0019.1
- Myhre, G., Shindell, D., Bréon, F.-M., Collins, W., Fuglestad, J., Huang, J., ... Zhang, H. (2014). *Anthropogenic and Natural Radiative Forcing. In: Climate Change 2013: The Physical Science Basis. Contribution of Working Group I to the Fifth Assessment Report of the Intergovernmental Panel on Climate Change* (Tech. Rep. No. AR5). Retrieved from <https://www.ipcc.ch/report/ar5/wg1/> (Publisher: Cambridge University Press)
- Neale, R. B., Richter, J. H., Conley, A. J., Park, S., Lauritzen, P. H., Gettelman, A., ... Lin, S.-J. (2010, April). *Description of the NCAR Community Atmosphere Model (CAM 4.0)*.
- Newhall, C. G., & Self, S. (1982). The volcanic explosivity index (VEI) an estimate of explosive magnitude for historical volcanism. *Journal of Geophysical Research: Oceans*, 87(C2), 1231–1238. Retrieved 2022-07-26, from <https://onlinelibrary.wiley.com/doi/abs/10.1029/JC087iC02p01231> (_eprint: <https://onlinelibrary.wiley.com/doi/pdf/10.1029/JC087iC02p01231>) doi: 10.1029/JC087iC02p01231
- Nordling, K., Korhonen, H., Räisänen, J., Partanen, A.-I., Samset, B. H., & Merikanto, J. (2021, October). Understanding the surface temperature response and its uncertainty to CO₂, CH₄, black carbon, and sulfate. *Atmospheric Chemistry and Physics*, 21(19), 14941–14958. Retrieved 2022-07-11, from <https://acp.copernicus.org/articles/21/14941/2021/> (Publisher: Copernicus GmbH) doi: 10.5194/acp-21-14941-2021
- Paynter, D., Frölicher, T. L., Horowitz, L. W., & Silvers, L. G. (2018). Equilibrium Climate Sensitivity Obtained From Multimillennial Runs of Two GFDL Climate Models. *Journal of Geophysical Research: Atmospheres*, 123(4), 1921–1941. Re-

- trieved 2019-07-15, from <https://agupubs.onlinelibrary.wiley.com/doi/abs/10.1002/2017JD027885> doi: 10.1002/2017JD027885
- Peixoto, J. P., & Oort, A. H. (1992). *Physics of climate*. American Institute of Physics. Retrieved 2019-07-04, from <https://www.osti.gov/biblio/7287064>
- Pendergrass, A. G., Conley, A., & Vitt, F. M. (2018, February). Surface and top-of-atmosphere radiative feedback kernels for CESM-CAM5. *Earth System Science Data*, 10(1), 317–324. Retrieved 2023-01-06, from <https://essd.copernicus.org/articles/10/317/2018/> (Publisher: Copernicus GmbH) doi: 10.5194/essd-10-317-2018
- Persad, G. G., & Caldeira, K. (2018, December). Divergent global-scale temperature effects from identical aerosols emitted in different regions. *Nature Communications*, 9(1), 3289. Retrieved 2022-06-24, from <http://www.nature.com/articles/s41467-018-05838-6> doi: 10.1038/s41467-018-05838-6
- Picas, J., & Grab, S. (2020, April). Potential impacts of major nineteenth century volcanic eruptions on temperature over Cape Town, South Africa: 1834–1899. *Climatic Change*, 159(4), 523–544. Retrieved 2022-12-12, from <https://doi.org/10.1007/s10584-020-02678-6> doi: 10.1007/s10584-020-02678-6
- Pincus, R., Forster, P. M., & Stevens, B. (2016, September). The Radiative Forcing Model Intercomparison Project (RFMIP): experimental protocol for CMIP6. *Geoscientific Model Development*, 9(9), 3447–3460. Retrieved 2022-02-04, from <https://gmd.copernicus.org/articles/9/3447/2016/> doi: 10.5194/gmd-9-3447-2016
- Po-Chedley, S., Armour, K. C., Bitz, C. M., Zelinka, M. D., Santer, B. D., & Fu, Q. (2018, January). Sources of Intermodel Spread in the Lapse Rate and Water Vapor Feedbacks. *Journal of Climate*, 31(8), 3187–3206. Retrieved 2019-07-15, from <https://journals.ametsoc.org/doi/full/10.1175/JCLI-D-17-0674.1> doi: 10.1175/JCLI-D-17-0674.1
- Pope, V. D., Gallani, M. L., Rowntree, P. R., & Stratton, R. A. (2000, February). The impact of new physical parametrizations in the Hadley Centre climate model: HadAM3. *Climate Dynamics*, 16(2), 123–146. Retrieved 2021-10-22, from <https://doi.org/10.1007/s003820050009> doi: 10.1007/s003820050009
- Portner, H.-O., Roberts, D., Poloczanska, E., Mintenbeck, K., Tignor, M., Alegría, A., ... Okem, A. (2022). *IPCC, 2022: Summary for Policymakers* (Tech. Rep.). Retrieved from https://www.ipcc.ch/report/ar6/wg2/downloads/report/IPCC_AR6_WGII_SummaryForPolicymakers.pdf
- Proistosescu, C., & Huybers, P. J. (2017, July). Slow climate mode reconciles historical and model-based estimates of climate sensitivity. *Science Advances*, 3(7), e1602821. Retrieved 2022-06-16, from <https://www.science.org/doi/10.1126/sciadv.1602821> (Publisher: American Association for the Advancement of Science) doi: 10.1126/sciadv.1602821
- Ramaswamy, V., Boucher, O., Haigh, J., Hauglustaine, D., Haywood, J., Myhre, G., ... Srinivasan, J. (2001). *Radiative Forcing of Climate Change* (Tech. Rep.). IPCC, Geneva, Switzerland. Retrieved from <https://www.ipcc.ch/site/assets/uploads/2018/03/TAR-06.pdf>
- Richardson, T. B., Forster, P. M., Smith, C. J., Maycock, A. C., Wood, T., Andrews, T., ... Watson-Parris, D. (2019). Efficacy of Climate Forcings in PDRMIP Models. *Journal of Geophysical Research: Atmospheres*, 124(23), 12824–12844. Retrieved 2020-01-17, from <https://agupubs.onlinelibrary.wiley.com/doi/abs/10.1029/2019JD030581> doi: 10.1029/2019JD030581

- Ringer, M. A., Andrews, T., & Webb, M. J. (2014). Global-mean radiative feedbacks and forcing in atmosphere-only and coupled atmosphere-ocean climate change experiments. *Geophysical Research Letters*, 41(11), 4035–4042. Retrieved 2021-10-22, from <https://onlinelibrary.wiley.com/doi/abs/10.1002/2014GL060347> (_eprint: <https://onlinelibrary.wiley.com/doi/pdf/10.1002/2014GL060347>) doi: 10.1002/2014GL060347
- Roeckner, E., Schlese, U., Biercamp, J., & Loewe, P. (1987, September). Cloud optical depth feedbacks and climate modelling. *Nature*, 329(6135), 138–140. Retrieved 2023-01-11, from <https://www.nature.com/articles/329138a0> (Number: 6135 Publisher: Nature Publishing Group) doi: 10.1038/329138a0
- Rohrschneider, T., Stevens, B., & Mauritsen, T. (2019, September). On simple representations of the climate response to external radiative forcing. *Climate Dynamics*, 53(5), 3131–3145. Retrieved 2023-01-09, from <https://doi.org/10.1007/s00382-019-04686-4> doi: 10.1007/s00382-019-04686-4
- Rose, B. E. J., Armour, K. C., Battisti, D. S., Feldl, N., & Koll, D. D. B. (2014). The dependence of transient climate sensitivity and radiative feedbacks on the spatial pattern of ocean heat uptake. *Geophysical Research Letters*, 41(3), 1071–1078. Retrieved 2021-11-01, from <https://onlinelibrary.wiley.com/doi/abs/10.1002/2013GL058955> (_eprint: <https://onlinelibrary.wiley.com/doi/pdf/10.1002/2013GL058955>) doi: 10.1002/2013GL058955
- Rose, B. E. J., & Rayborn, L. (2016, December). The Effects of Ocean Heat Uptake on Transient Climate Sensitivity. *Current Climate Change Reports*, 2(4), 190–201. Retrieved 2021-11-01, from <http://link.springer.com/10.1007/s40641-016-0048-4> doi: 10.1007/s40641-016-0048-4
- Rotstayn, L. D., & Penner, J. E. (2001, July). Indirect Aerosol Forcing, Quasi Forcing, and Climate Response. *Journal of Climate*, 14(13), 2960–2975. Retrieved 2019-07-25, from <https://journals.ametsoc.org/doi/full/10.1175/1520-0442%282001%29014%3C2960%3AIAFQFA%3E2.0.CO%3B2> doi: 10.1175/1520-0442(2001)014<2960:IAFQFA>2.0.CO;2
- Rugenstein, M., Bloch-Johnson, J., Abe-Ouchi, A., Andrews, T., Beyerle, U., Cao, L., ... Yang, S. (2019, December). LongRunMIP: Motivation and Design for a Large Collection of Millennial-Length AOGCM Simulations. *Bulletin of the American Meteorological Society*, 100(12), 2551–2570. Retrieved 2022-10-21, from <https://journals.ametsoc.org/view/journals/bams/100/12/bams-d-19-0068.1.xml> (Publisher: American Meteorological Society Section: Bulletin of the American Meteorological Society) doi: 10.1175/BAMS-D-19-0068.1
- Rugenstein, M., Bloch-Johnson, J., Gregory, J., Andrews, T., Mauritsen, T., Li, C., ... Knutti, R. (2020). Equilibrium Climate Sensitivity Estimated by Equilibrating Climate Models. *Geophysical Research Letters*, 47(4), e2019GL083898. Retrieved 2022-10-21, from <https://onlinelibrary.wiley.com/doi/abs/10.1029/2019GL083898> (_eprint: <https://onlinelibrary.wiley.com/doi/pdf/10.1029/2019GL083898>) doi: 10.1029/2019GL083898
- Rugenstein, M. A. A., & Armour, K. C. (2021). Three Flavors of Radiative Feedbacks and Their Implications for Estimating Equilibrium Climate Sensitivity. *Geophysical Research Letters*, 48(15), e2021GL092983. Retrieved 2022-06-16, from <https://onlinelibrary.wiley.com/doi/abs/10.1029/2021GL092983> (_eprint:

- <https://onlinelibrary.wiley.com/doi/pdf/10.1029/2021GL092983>) doi: 10.1029/2021GL092983
- Rugenstein, M. A. A., Gregory, J. M., Schaller, N., Sedláček, J., & Knutti, R. (2016, May). Multiannual Ocean–Atmosphere Adjustments to Radiative Forcing. *Journal of Climate*, 29(15), 5643–5659. Retrieved 2019-07-15, from <https://journals.ametsoc.org/doi/full/10.1175/JCLI-D-16-0312.1> doi: 10.1175/JCLI-D-16-0312.1
- Salvi, P., Ceppi, P., & Gregory, J. M. (2021). Interpreting the Dependence of Cloud-Radiative Adjustment on Forcing Agent. *Geophysical Research Letters*, 48(18), e2021GL093616. Retrieved 2022-04-25, from <https://onlinelibrary.wiley.com/doi/abs/10.1029/2021GL093616> (_eprint: <https://onlinelibrary.wiley.com/doi/pdf/10.1029/2021GL093616>) doi: 10.1029/2021GL093616
- Salvi, P., Ceppi, P., & Gregory, J. M. (2022). Interpreting Differences in Radiative Feedbacks From Aerosols Versus Greenhouse Gases. *Geophysical Research Letters*, 49(8), e2022GL097766. Retrieved 2022-04-25, from <https://onlinelibrary.wiley.com/doi/abs/10.1029/2022GL097766> (_eprint: <https://onlinelibrary.wiley.com/doi/pdf/10.1029/2022GL097766>) doi: 10.1029/2022GL097766
- Samset, B. H., & Myhre, G. (2015). Climate response to externally mixed black carbon as a function of altitude. *Journal of Geophysical Research: Atmospheres*, 120(7), 2913–2927. Retrieved 2019-10-04, from <https://agupubs.onlinelibrary.wiley.com/doi/abs/10.1002/2014JD022849> doi: 10.1002/2014JD022849
- Sand, M., Samset, B. H., Tsigaridis, K., Bauer, S. E., & Myhre, G. (2020). Black Carbon and Precipitation: An Energetics Perspective. *Journal of Geophysical Research: Atmospheres*, 125(13), e2019JD032239. Retrieved 2021-06-17, from <https://agupubs.onlinelibrary.wiley.com/doi/abs/10.1029/2019JD032239> (_eprint: <https://agupubs.onlinelibrary.wiley.com/doi/pdf/10.1029/2019JD032239>) doi: 10.1029/2019JD032239
- Schlesinger, M. E. (1988, September). Negative or positive cloud optical depth feedback? *Nature*, 335(6188), 303–304. Retrieved 2023-02-06, from <https://www.nature.com/articles/335303c0> (Number: 6188 Publisher: Nature Publishing Group) doi: 10.1038/335303c0
- Schmidt, A., Mills, M. J., Ghan, S., Gregory, J. M., Allan, R. P., Andrews, T., ... Toon, O. B. (2018). Volcanic Radiative Forcing From 1979 to 2015. *Journal of Geophysical Research: Atmospheres*, 123(22), 12491–12508. Retrieved 2022-07-26, from <https://onlinelibrary.wiley.com/doi/abs/10.1029/2018JD028776> (_eprint: <https://onlinelibrary.wiley.com/doi/pdf/10.1029/2018JD028776>) doi: 10.1029/2018JD028776
- Schmidt, G. A., Ruedy, R., Hansen, J. E., Aleinov, I., Bell, N., Bauer, M., ... Yao, M.-S. (2006, January). Present-Day Atmospheric Simulations Using GISS ModelE: Comparison to In Situ, Satellite, and Reanalysis Data. *Journal of Climate*, 19(2), 153–192. Retrieved 2019-07-25, from <https://journals.ametsoc.org/doi/full/10.1175/JCLI3612.1> doi: 10.1175/JCLI3612.1
- Schmidt, G. A., Ruedy, R. A., Miller, R. L., & Lacis, A. A. (2010, October). Attribution of the present-day total greenhouse effect. *Journal of Geophysical Research*, 115(D20), D20106. Retrieved 2019-08-09, from <http://doi.wiley.com/10.1029/2010JD014287> doi: 10.1029/2010JD014287

- Sherwood, S. C., Bony, S., Boucher, O., Bretherton, C., Forster, P. M., Gregory, J. M., & Stevens, B. (2015, February). Adjustments in the Forcing-Feedback Framework for Understanding Climate Change. *Bulletin of the American Meteorological Society*, 96(2), 217–228. Retrieved 2020-12-10, from <https://journals.ametsoc.org/view/journals/bams/96/2/bams-d-13-00167.1.xml> (Publisher: American Meteorological Society Section: Bulletin of the American Meteorological Society) doi: 10.1175/BAMS-D-13-00167.1
- Shindell, D. T. (2014, April). Inhomogeneous forcing and transient climate sensitivity. *Nature Climate Change*, 4(4), 274–277. Retrieved 2021-09-10, from <http://www.nature.com/articles/nclimate2136> doi: 10.1038/nclimate2136
- Shine, K. P., Cook, J., Highwood, E. J., & Joshi, M. M. (2003). An alternative to radiative forcing for estimating the relative importance of climate change mechanisms. *Geophysical Research Letters*, 30(20). Retrieved 2021-01-11, from <https://agupubs.onlinelibrary.wiley.com/doi/abs/10.1029/2003GL018141> (_eprint: <https://agupubs.onlinelibrary.wiley.com/doi/pdf/10.1029/2003GL018141>) doi: <https://doi.org/10.1029/2003GL018141>
- Shine, K. P., Fouquart, Y., Ramaswamy, V., Solomon, S., & Srinivasan, J. (1995). *Radiative Forcing. In: IPCC Report on Radiative Forcing of Climate Change and An Evaluation of the IPCC IS92 Emission Scenarios* (Tech. Rep.). Retrieved 2021-01-14, from <https://www.ipcc.ch/report/climate-change-1994-radiative-forcing-of-climate-change-and-an-evaluation-of-the-ipcc-is92-emission-scenarios-2/>
- Smith, C. J., & Forster, P. M. (2021). Suppressed Late-20th Century Warming in CMIP6 Models Explained by Forcing and Feedbacks. *Geophysical Research Letters*, 48(19), e2021GL094948. Retrieved 2022-01-24, from <https://onlinelibrary.wiley.com/doi/abs/10.1029/2021GL094948> (_eprint: <https://onlinelibrary.wiley.com/doi/pdf/10.1029/2021GL094948>) doi: 10.1029/2021GL094948
- Smith, C. J., Kramer, R. J., Myhre, G., Alterskjær, K., Collins, W., Sima, A., ... Forster, P. M. (2020, January). *Effective radiative forcing and adjustments in CMIP6 models* (preprint). Radiation/Atmospheric Modelling/Troposphere/Physics (physical properties and processes). Retrieved 2020-02-05, from <https://www.atmos-chem-phys-discuss.net/acp-2019-1212/> doi: 10.5194/acp-2019-1212
- Smith, C. J., Kramer, R. J., Myhre, G., Forster, P. M., Soden, B. J., Andrews, T., ... Watson-Parris, D. (2018). Understanding Rapid Adjustments to Diverse Forcing Agents. *Geophysical Research Letters*, 45(21), 12,023–12,031. Retrieved 2019-07-15, from <https://agupubs.onlinelibrary.wiley.com/doi/abs/10.1029/2018GL079826> doi: 10.1029/2018GL079826
- Sobel, A. H. (2002). The ENSO Signal in Tropical Tropospheric Temperature. *JOURNAL OF CLIMATE*, 15, 5.
- Soden, B. J., Held, I. M., Colman, R., Shell, K. M., Kiehl, J. T., & Shields, C. A. (2008, July). Quantifying Climate Feedbacks Using Radiative Kernels. *Journal of Climate*, 21(14), 3504–3520. Retrieved 2019-07-15, from <https://journals.ametsoc.org/doi/full/10.1175/2007JCLI2110.1> doi: 10.1175/2007JCLI2110.1
- Stenchikov, G. L., Kirchner, I., Robock, A., Graf, H.-F., Antuña, J. C., Grainger, R. G., ... Thomason, L. (1998). Radiative forcing from the 1991 Mount Pinatubo volcanic eruption. *Journal of Geophysical Research: Atmospheres*, 103(D12), 13837–13857. Retrieved 2022-06-08, from

- <https://onlinelibrary.wiley.com/doi/abs/10.1029/98JD00693> (_eprint: <https://onlinelibrary.wiley.com/doi/pdf/10.1029/98JD00693>) doi: 10.1029/98JD00693
- Stevens, B., Sherwood, S. C., Bony, S., & Webb, M. J. (2016, November). Prospects for narrowing bounds on Earth’s equilibrium climate sensitivity. *Earth’s Future*, 4(11), 512–522. Retrieved 2019-08-05, from <https://agupubs.onlinelibrary.wiley.com/doi/full/10.1002/2016EF000376> doi: 10.1002/2016EF000376
- Stjern, C. W., Samset, B. H., Boucher, O., Iversen, T., Lamarque, J.-F., Myhre, G., ... Takemura, T. (2020, November). How aerosols and greenhouse gases influence the diurnal temperature range. *Atmospheric Chemistry and Physics*, 20(21), 13467–13480. Retrieved 2021-02-01, from <https://acp.copernicus.org/articles/20/13467/2020/> (Publisher: Copernicus GmbH) doi: <https://doi.org/10.5194/acp-20-13467-2020>
- Stjern, C. W., Samset, B. H., Myhre, G., Forster, P. M., Hodnebrog, O., Andrews, T., ... Voulgarakis, A. (2017, November). Rapid Adjustments Cause Weak Surface Temperature Response to Increased Black Carbon Concentrations. *Journal of Geophysical Research: Atmospheres*. Retrieved 2019-07-15, from <https://agupubs.onlinelibrary.wiley.com/doi/abs/10.1002/2017JD027326> doi: 10.1002/2017JD027326
- Stouffer, R. J. (2004, January). Time Scales of Climate Response. *Journal of Climate*, 17(1), 209–217. Retrieved 2019-08-15, from <https://journals.ametsoc.org/doi/full/10.1175/1520-0442%282004%29017%3C0209%3ATSOCR%3E2.0.CO%3B2> doi: 10.1175/1520-0442(2004)017<0209:TSOCR>2.0.CO;2
- Swales, D. J., Pincus, R., & Bodas-Salcedo, A. (2018, January). The Cloud Feedback Model Intercomparison Project Observational Simulator Package: Version 2. *Geoscientific Model Development*, 11(1), 77–81. Retrieved 2021-01-12, from <https://gmd.copernicus.org/articles/11/77/2018/> (Publisher: Copernicus GmbH) doi: <https://doi.org/10.5194/gmd-11-77-2018>
- Taylor, K. E., Juckes, M., Balaji, V., Cinquini, L., Denvil, S., Durack, P. J., ... Stockhause, M. (2018, September). *CMIP6 Global Attributes, DRS, Filenames, Directory Structure, and CV’s*. Retrieved 2023-02-02, from https://docs.google.com/document/d/1h0r8RZr_f3-8egBMMh7aqLwy3snpD6_MrDz1q8n5XUk/edit?usp=embed_facebook ((v6.2.7))
- Taylor, P. C. (2012, July). The Role of Clouds: An Introduction and Rapporteur Report. *Surveys in Geophysics*, 33(3), 609–617. Retrieved 2019-08-09, from <https://doi.org/10.1007/s10712-012-9182-2> doi: 10.1007/s10712-012-9182-2
- Vial, J., Dufresne, J.-L., & Bony, S. (2013, December). On the interpretation of inter-model spread in CMIP5 climate sensitivity estimates. *Climate Dynamics*, 41(11), 3339–3362. Retrieved 2021-01-13, from <https://doi.org/10.1007/s00382-013-1725-9> doi: 10.1007/s00382-013-1725-9
- Wang, H., & Wen, Y.-J. (2021, November). Climate response to the spatial and temporal evolutions of anthropogenic aerosol forcing. *Climate Dynamics*. Retrieved 2022-06-07, from <https://link.springer.com/10.1007/s00382-021-06059-2> doi: 10.1007/s00382-021-06059-2
- Watkiss, P., Troeltzsch, J., & McGlade, K. (2018, May). *The Economic Cost of Climate Change in Europe: Synthesis Report on State of Knowledge and Key Research Gaps. Policy brief by the COACCH project*. (Tech. Rep.). COACCH. Retrieved 2019-07-26, from <https://www.ecologic.eu/sites/files/publication/>

- 2018/2811-coacch-review-synthesis-updated-june-2018.pdf
- Webb, M. J., Andrews, T., Bodas-Salcedo, A., Bony, S., Bretherton, C. S., Chadwick, R., ... Watanabe, M. (2017, January). The Cloud Feedback Model Intercomparison Project (CFMIP) contribution to CMIP6. *Geoscientific Model Development*, 10(1), 359–384. Retrieved 2022-07-25, from <https://gmd.copernicus.org/articles/10/359/2017/> (Publisher: Copernicus GmbH) doi: 10.5194/gmd-10-359-2017
- Wiart, P., & Oppenheimer, C. (2000, April). Largest known historical eruption in Africa: Dubbi volcano, Eritrea, 1861. *Geology*, 28(4), 291–294. Retrieved 2022-12-12, from [https://doi.org/10.1130/0091-7613\(2000\)28<291:LKHEIA>2.0.CO;2](https://doi.org/10.1130/0091-7613(2000)28<291:LKHEIA>2.0.CO;2) doi: 10.1130/0091-7613(2000)28<291:LKHEIA>2.0.CO;2
- Wild, M., Folini, D., Hakuba, M. Z., Schär, C., Seneviratne, S. I., Kato, S., ... König-Langlo, G. (2015, June). The energy balance over land and oceans: an assessment based on direct observations and CMIP5 climate models. *Climate Dynamics*, 44(11-12), 3393–3429. Retrieved 2022-12-05, from <http://link.springer.com/10.1007/s00382-014-2430-z> doi: 10.1007/s00382-014-2430-z
- Wills, R. C. J., Dong, Y., Proistosescu, C., Armour, K. C., & Battisti, D. S. (2022). Systematic Climate Model Biases in the Large-Scale Patterns of Recent Sea-Surface Temperature and Sea-Level Pressure Change. *Geophysical Research Letters*, 49(17), e2022GL100011. Retrieved 2022-10-21, from <https://onlinelibrary.wiley.com/doi/abs/10.1029/2022GL100011> (_eprint: <https://onlinelibrary.wiley.com/doi/pdf/10.1029/2022GL100011>) doi: 10.1029/2022GL100011
- Winton, M., Takahashi, K., & Held, I. M. (2010, May). Importance of Ocean Heat Uptake Efficacy to Transient Climate Change. *Journal of Climate*, 23(9), 2333–2344. Retrieved 2022-02-04, from <https://journals.ametsoc.org/view/journals/clim/23/9/2009jcli3139.1.xml> (Publisher: American Meteorological Society Section: Journal of Climate) doi: 10.1175/2009JCLI3139.1
- Wood, R., & Bretherton, C. S. (2006, December). On the Relationship between Stratiform Low Cloud Cover and Lower-Tropospheric Stability. *Journal of Climate*, 19(24), 6425–6432. Retrieved 2019-09-16, from <https://journals.ametsoc.org/doi/full/10.1175/JCLI3988.1> doi: 10.1175/JCLI3988.1
- Wood, R., & Hartmann, D. L. (2006). Spatial variability of liquid water path in marine low cloud: The importance of mesoscale cellular convection. *Journal of Climate*, 19(9), 1748–1764. Retrieved 2022-07-01, from <https://journals.ametsoc.org/view/journals/clim/19/9/jcli3702.1.xml> (Publisher: American Meteorological Society)
- Wu, Y., Seager, R., Ting, M., Naik, N., & Shaw, T. A. (2011, December). Atmospheric Circulation Response to an Instantaneous Doubling of Carbon Dioxide. Part I: Model Experiments and Transient Thermal Response in the Troposphere. *Journal of Climate*, 25(8), 2862–2879. Retrieved 2019-07-26, from <https://journals.ametsoc.org/doi/full/10.1175/JCLI-D-11-00284.1> doi: 10.1175/JCLI-D-11-00284.1
- Zelinka, M. D., Klein, S. A., & Hartmann, D. L. (2012a, June). Computing and Partitioning Cloud Feedbacks Using Cloud Property Histograms. Part I: Cloud Radiative Kernels. *Journal of Climate*, 25(11), 3715–3735. Retrieved 2020-03-17, from <https://journals.ametsoc.org/doi/full/10.1175/JCLI-D-11-00248.1> doi: 10.1175/JCLI-D-11-00248.1
- Zelinka, M. D., Klein, S. A., & Hartmann, D. L. (2012b, June). Computing and Par-

- tioning Cloud Feedbacks Using Cloud Property Histograms. Part II: Attribution to Changes in Cloud Amount, Altitude, and Optical Depth. *Journal of Climate*, 25(11), 3736–3754. Retrieved 2020-04-28, from <https://journals.ametsoc.org/doi/full/10.1175/JCLI-D-11-00249.1> (Publisher: American Meteorological Society) doi: 10.1175/JCLI-D-11-00249.1
- Zelinka, M. D., Klein, S. A., Taylor, K. E., Andrews, T., Webb, M. J., Gregory, J. M., & Forster, P. M. (2013, January). Contributions of Different Cloud Types to Feedbacks and Rapid Adjustments in CMIP5. *Journal of Climate*, 26(14), 5007–5027. Retrieved 2019-07-15, from <https://journals.ametsoc.org/doi/full/10.1175/JCLI-D-12-00555.1> doi: 10.1175/JCLI-D-12-00555.1
- Zelinka, M. D., Myers, T. A., McCoy, D. T., Po-Chedley, S., Caldwell, P. M., Ceppi, P., ... Taylor, K. E. (2020). Causes of Higher Climate Sensitivity in CMIP6 Models. *Geophysical Research Letters*, 47(1), e2019GL085782. Retrieved 2020-05-12, from <https://agupubs.onlinelibrary.wiley.com/doi/abs/10.1029/2019GL085782> (_eprint: <https://agupubs.onlinelibrary.wiley.com/doi/pdf/10.1029/2019GL085782>) doi: 10.1029/2019GL085782
- Zhou, C., Zelinka, M. D., & Klein, S. A. (2016, December). Impact of decadal cloud variations on the Earth’s energy budget. *Nature Geoscience*, 9(12), 871–874. Retrieved 2021-10-12, from <http://www.nature.com/articles/ngeo2828> doi: 10.1038/ngeo2828
- Zhou, C., Zelinka, M. D., & Klein, S. A. (2017). Analyzing the dependence of global cloud feedback on the spatial pattern of sea surface temperature change with a Green’s function approach. *Journal of Advances in Modeling Earth Systems*, 9(5), 2174–2189. Retrieved 2021-06-17, from <https://agupubs.onlinelibrary.wiley.com/doi/abs/10.1002/2017MS001096> (_eprint: <https://agupubs.onlinelibrary.wiley.com/doi/pdf/10.1002/2017MS001096>) doi: 10.1002/2017MS001096
- Zhuo, Z., Kirchner, I., Pfahl, S., & Cubasch, U. (2021, September). Climate impact of volcanic eruptions: the sensitivity to eruption season and latitude in MPI-ESM ensemble experiments. *Atmospheric Chemistry and Physics*, 21(17), 13425–13442. Retrieved 2022-02-07, from <https://acp.copernicus.org/articles/21/13425/2021/> doi: 10.5194/acp-21-13425-2021