

On the climate response to spatial patterns of absorbing aerosol and sea-surface temperatures



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*This thesis is dedicated to my mother, Susan Williams,
for the immense love you've shown me my whole life.*

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Abstract

Earth’s climate demonstrates rich regional variations; no two places on our planet are quite the same. From rain forests to deserts and doldrums to storm tracks, the character of the atmospheric circulation and its links to rainfall and cloudiness vary markedly across the globe. These regional features present a problem to the climate science community when faced with the problem of understanding how human activities (which also have a distinct spatial pattern) impact on climate. This thesis aims to bolster our understanding of how the geographical location and pattern of perturbations impact on their climate response. To do this, we focus on two problems; the climate response to localized absorbing aerosol and to sea-surface temperature anomalies. In both of these cases, we use an idealized modeling approach to isolate the key features of the problem and provide mechanistic, physical insights into the ‘spatial sensitivities’ we uncover.

This thesis begins by investigating the relationship between global-mean effective radiative forcing (ERF) and global-mean aerosol optical depth (AOD) for the case of absorbing aerosol, in Chapter 2. We use a large ensemble of atmosphere-only simulations with identical aerosol absorption perturbations in different geographical locations to show that the interactions between deep convection and diabatic heating resulting from aerosol absorption (such as occurs during large wildfire events) leads to an “aerosol pattern effect”, where the ERF from an identical burden of absorbing aerosol can vary by an order-of-magnitude and even change sign depending on where the aerosol heating is located relative to the region of tropical ascent.

Building on this work, we then investigate the implications of this “aerosol pattern effect” for understanding the global- and local- precipitation response to absorbing aerosol in Chapter 3. Contrary to the assumptions in recent work, we find that the relationship between global-mean AOD and global-mean precipitation changes depends strongly on the geographical location of the aerosol plume. We use an energetic decomposition of the global-mean precipitation changes into contributions from shortwave absorption, longwave cooling and sensible heat fluxes and link changes in these quantities to the dynamical and thermodynamical responses to localized aerosol heating. Further to this, we also explore the *local* precipitation

response to absorbing aerosol, confirming the ‘tropical-extratropical’ contrast found by previous studies. We then provide a robust physical mechanism for this contrast in terms of simple atmospheric dynamics.

In Chapter 4 we then turn our attention to the related question of how sea-surface temperature anomalies in different regions impact on the global-mean top-of-atmosphere (TOA) radiation budget, a phenomenon often referred to as the “SST pattern effect”. Previous work has attempted to quantify the SST pattern effect using a Green’s function approach, which assumes that TOA fluxes respond linearly to SST anomalies in a given region and that they combine linearly. However, using idealized model simulations, I demonstrate that the response of global-mean TOA fluxes to localized tropical SST perturbations is both non-linear and non-additive. I then explain that this behavior arises as a robust consequence of convective quasi-equilibrium and weak (but non-zero) temperature gradients in the tropical free-troposphere, which I encapsulate in a ‘circus tent’ model of the tropical atmosphere. These results demonstrate that the climate response to SST perturbations is fundamentally non-linear, due to the physics of deep convection.

All-in-all, this thesis highlights the crucial role that the spatial pattern of perturbations plays in determining their end climate response, which is an aspect which is often ignored in simple, ‘top-down’ models of the climate system. Furthermore, we have used an idealized modeling approach to find simple, physical mechanisms which explain how localized perturbations (in the form of absorbing aerosol and sea-surface temperature anomalies) generate different responses depending on their location. These ‘spatial sensitivities’ are often largest in the tropics, where the interactions between local perturbations and deep convection can lead to large, non-local responses; this suggests a deeper understanding of the links between deep convection and the large-scale atmosphere is needed to fully understand the climate response to spatial patterns of forcing.

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Manuel: Ah, your horse. It win! It win!

Basil Fawlty: [wanting him to keep quiet about his gambling venture] Shhh-shh-shh, Manuel. You - know - nothing.

Manuel: You ALWAYS say, Mr. Fawlty, but I learn.

Basil Fawlty: What?

Manuel: I learn. I learn.

Basil Fawlty: No-no-no-no-no.

Manuel: I get better.

Basil Fawlty: No-no. No-no, you don't understand.

Manuel: I do.

Basil Fawlty: No, you don't.

Manuel: Hey, I do understand that!

— Fawlty Towers

1

Introduction

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1.1 Motivation

Our planet is home to a rich and intricate diversity of climatic features. Ranging from the intense heat of the tropics to the cold of the polar regions, and from moist rainforests to arid deserts. Because the Earth is rotating, the character of the atmospheric circulation also varies hugely between the tropics and the higher latitudes. The tropical atmosphere generally consists of broad areas of dry descending air driven by radiative cooling and narrow, moist regions of intense ascent driven by towering cumulonimbus clouds (thunderstorms). The midlatitude atmosphere, on the other hand, is filled with vigorous weather systems known as

baroclinic eddies, for which convective dynamics are less important compared to the requisite conservation of potential vorticity.

The climate of any specific region is determined by the changing characteristics of the atmospheric circulation, with additional complications arising due to the local surface properties of the region, its relation to water sources and orography and the susceptibility of the region to abrupt seasonal transitions such as monsoons. The scale of this complexity is astounding, and a testament to the rich nonlinearities which arise when coupling together the basic physics of radiative transfer, thermodynamics and fluid flow.

Since the rise of the Industrial Revolution, human activity has become the preeminent factor in determining the long-term trajectory of Earth's climate. Understanding how such human (or 'anthropogenic') activities alter the climate is thus an extremely important question for the scientific community to answer. Understanding the influence of anthropogenic activity on Earth's climate boils down to understanding the 'forcing' applied to the climate, the 'feedbacks' at play when the climate is moving towards a new equilibrium state, and the downstream 'impacts' of these changes for human society. Understanding these factors is complicated, not just because Earth's present-day climate has complex regional variations, but also because *perturbations* to Earth's climate often also have a distinct geographical structure. One case for which this is particularly true of is anthropogenic aerosol, whose spatial distribution is extremely heterogeneous because of the relatively short atmospheric lifetime of aerosol particles. Another is sea-surface temperatures (SSTs), which also vary on the scale of hundreds to thousands of kilometres; exemplified in the El-Niño-Southern Oscillation, which is the largest mode of inter-annual variability in the tropics.

This thesis weaves together contributions I have made in our understanding of how the spatial *pattern* of perturbations influences their climate response, with

three chapters which focus on each of the three factors I mentioned earlier; forcing, feedback and impacts. The overarching approach is to use idealized climate model experiments to probe the sensitivity of the climate effect of regional aerosol or SSTs to their geographical location, which we will show illuminates interesting and previously unknown behaviour of the climate system. A recurring theme in these chapters is that the main determinant of these ‘spatial sensitivities’ is the way in which clouds respond to these idealized perturbations.

The remainder of the introduction will proceed as follows. Firstly I will introduce aerosol, clouds and precipitation and how these features are parameterized in the ICON global-circulation model. Then I will summarize the ‘forcing-feedback’ framework for understanding global-mean climate changes, which will involve further clarification of the terms ‘forcing’ and ‘feedback’, and a discussion of their uncertainties and open questions. Finally I will provide an outline of the thesis structure and goals.

1.2 Aerosol

Aerosol are solid or liquid particles suspended in the atmosphere, with sizes ranging from approximately 1 nm to 10s of micrometers (Seinfeld and Pandis, 1998). Aerosol enters the atmosphere through emission processes, which can directly inject aerosol particles into the atmosphere, generating so-called ‘primary aerosol’ such as mineral dust or sea-spray particles. Aerosol particles can also be formed by reactions *within* the atmosphere, generating so-called ‘secondary aerosol’ such as sulfates or nitrates, which are produced by oxidation of precursor emissions.

Emissions of aerosol can come from either natural or anthropogenic sources. ‘Natural’ aerosol refers to precursor emissions and aerosol particles which originate through processes such as volcanic activity, lightning, wildfires, and natural ocean-atmosphere interactions. Examples include sea salt aerosol (Ma et al., 2008), desert dust (Mahowald et al., 2014), or stratospheric aerosol from volcanic eruptions (Robock, 2000). On the other hand, ‘anthropogenic’ aerosol are particles generated by human activities such as industry, transportation or farming. The largest source of anthropogenic aerosol emissions is through fossil fuel combustion, which alongside generating CO₂ also acts as a source of aerosol precursors. It is also worth noting that this distinction between ‘natural’ and ‘anthropogenic’ aerosol can be somewhat blurry, due to the fact that some aerosol species can originate from both natural and anthropogenic sources, and that anthropogenic activities have enhanced the emission of natural aerosol such as desert dust, or black carbon from wildfires (Bond et al., 2013).

Once aerosol particles are added to the atmosphere, the next thing we need to know is how they are removed and on what timescale this occurs. The ultimate paths by which aerosol particles are removed from the atmosphere are ‘dry’ and ‘wet’ deposition. The ‘dry’ vs ‘wet’ distinction refers to whether the depositional processes involve atmospheric hydrometeors or not (Seinfeld and Pandis, 1998). Dry deposition largely occurs through gravitational sedimentation and can be

parameterized in models by using a bulk *deposition velocity*, although turbulent diffusion also acts to modulate the dry deposition flux near the surface (Seinfeld and Pandis, 1998). Wet deposition involves the ‘scavenging’ of aerosol particles by atmospheric hydrometeors such as rain, snow and cloud droplets. A broad distinction is generally made between ‘in-cloud’ scavenging and ‘below-cloud’ scavenging, where ‘in-cloud’ refers to the sink of atmospheric aerosol particles which occurs when they act as nuclei for water vapour to condense onto or ice to aggregate onto, thus generating cloud droplets/ice crystals. Separate but related to this is ‘below-cloud’ scavenging, which involves the capture of aerosol particles by *falling* hydrometeors, such as raindrops, snow, hail and graupel. Wet deposition dominates aerosol removal with high efficiency, such that the lifetimes of aerosol are usually no more than ten days (Textor et al., 2006).

1.2.1 Historical variations in the spatial pattern and composition of anthropogenic aerosol

As a result of these efficient aerosol removal processes in Earth’s troposphere, aerosol particles generally have atmospheric lifetimes which are quite short - on the order of days to weeks. This results in aerosol burdens which are highly localized to their source regions.

To illustrate this, Figure 1.1 shows decadal-mean maps of historic anthropogenic aerosol optical depth at 550nm for each decade from the 1920s to the 2000s, derived from the MACv2SP climatology of Stevens et al. (2017) with the anthropogenic contribution to the coarse mode assumed negligible. Anthropogenic aerosol emissions are negligible in the late 19th century but increase rapidly in the first half of the 20th century following the intense industrialization of North America and Western Europe (Persad and Caldeira, 2018). Following the passing of various clean-air acts in the USA and Europe during the 1980s, there has been a steady decrease in AOD, particularly over Western Europe. On the other hand, the AOD associated with

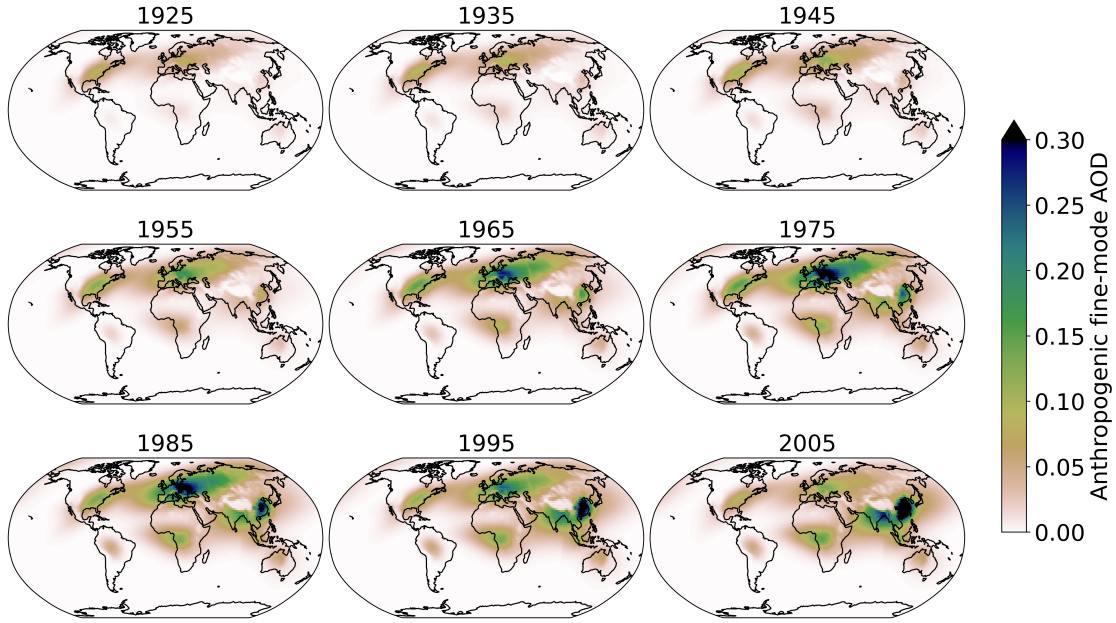


Figure 1.1: Decadal-mean maps of historic anthropogenic aerosol optical depth at 550nm from 1925-2005, derived from the MACv2SP climatology (Data from (Stevens et al., 2017)).

anthropogenic emissions over South and East Asia have steadily increased since around the 1940s, and currently dominate total anthropogenic aerosol emissions (Deser et al., 2020). Alongside industrial-driven increases, contributions from biomass-burning have also risen steadily over the period, evidenced by the rise in AOD over Central Africa, South America and Equatorial Asia (Brown et al., 2021).

While future emissions scenarios are inherently uncertain, in all of the Shared Socioeconomic Pathways (SSPs) used in the IPCC’s Coupled Model Intercomparison Project Phase 6 (CMIP6), emissions of anthropogenic aerosol are projected to decline by the end of the 21st century, with marked declines in emissions over South Asia, North America and Europe (Lund et al., 2019).

Despite the strong degree of spatial inhomogeneity in the AOD distribution, previous studies have noted a number of consistent patterns. For example, it is now well-documented that there is a distinct hemispheric asymmetry in the pattern of anthropogenic AOD and aerosol forcing, driven by the greater population

growth and industrialization in the Northern Hemisphere (e.g., Voigt et al. (2017) and Wang (2004); Chung and Soden (2017)). Additionally, Chemke and Dagan (2018) decomposed the observed pattern of direct shortwave aerosol forcing into its dominant spatial modes using spherical harmonics as basis-functions. They showed that while the hemispheric asymmetry of aerosol forcing accounts for around 20% of the variance in the spatial pattern, the two subsequent strongest modes are an East-West asymmetry and a ‘midlatitudes-tropics’ contrast which each account for around 10% of the variance.

Alongside the rapid changes in the spatial pattern of anthropogenic aerosol which we saw in Figure 1.1, the *composition* of atmospheric aerosol has also varied significantly over the historical period. In this context, we take ‘composition’ to refer to the relative efficacy of aerosol in either absorbing or scattering incoming solar radiation. To quantify this, we introduce the ‘single-scattering albedo’ (SSA or ω), which is defined at a given wavelength as the ratio of extinction due to scattering to the total extinction. Using this we can calculate the total anthropogenic aerosol optical depth (AOD) which arises from absorbing aerosol, and both the total global-mean AOD and this absorbing component are plotted in Figure 1.2.

Figure 1.2 shows that since the 1980s there has been little change in global-mean anthropogenic aerosol emissions and subsequently the global-mean anthropogenic AOD has been relatively stable over the past 40 years (Lund et al., 2019; Samset, 2018). However, this stagnation of global-mean anthropogenic AOD masks many changes in the composition of the anthropogenic aerosol burden. As Figure 1.1 illustrated previously, the dominant emission regions have shifted markedly over the historical period, and one notable shift has been the increase in anthropogenic AOD from tropical biomass burning. Biomass burning, and more generally the process of incomplete combustion, generates carbonaceous aerosol which absorb significant amounts of incoming solar radiation and have a low SSA. This rise in carbonaceous aerosol is illustrated by the orange line in Figure 1.2, which shows

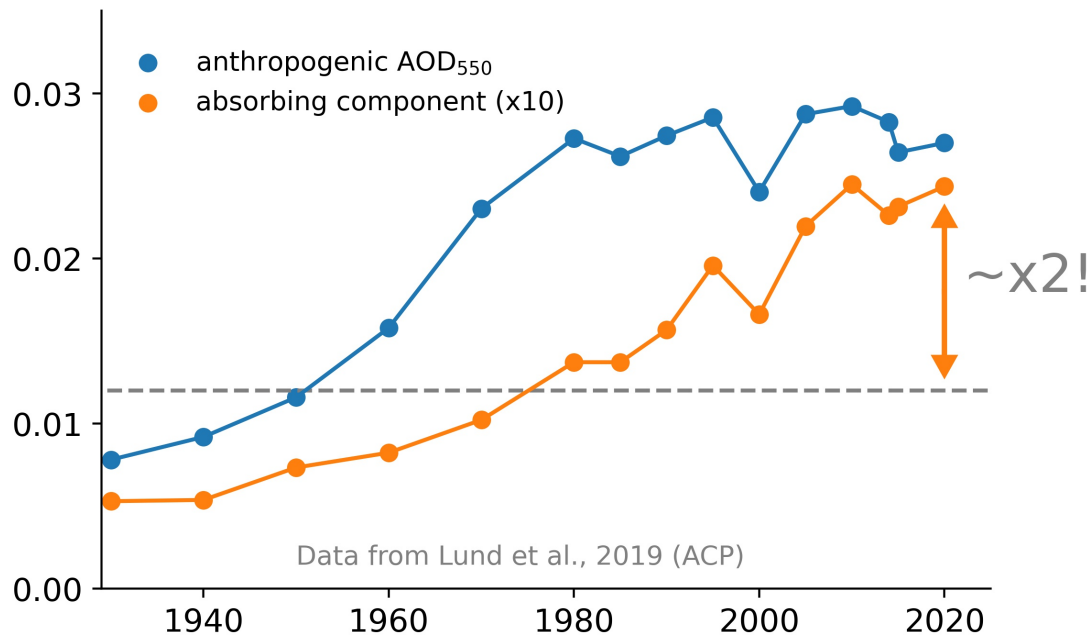


Figure 1.2: Time-series of global-mean anthropogenic aerosol optical depth (AOD) and the contribution from absorbing aerosol, multiplied by 10 so it can be plotted on the same axis. Data from Lund et al. (2019).

that the component of global-mean anthropogenic AOD arising from absorbing aerosol has roughly doubled since the 1980s. Put simply, since the 1980s the global aerosol distribution has become about twice as absorbing.

These trends towards a more ‘absorbing’ global aerosol distribution are likely to continue into the near-future, as a result of air quality policies which tackle sources of absorbing aerosol and non-absorbing aerosol (like sulfate) differently. Hence, even though the global-mean aerosol burden decreases in almost all future climate scenarios, absorbing aerosol is likely to become a significantly more important component of the aerosol problem, as opposed to sulfate aerosol which has dominated the historical evolution of anthropogenic aerosol (Samset, 2018). This increased prominence of absorbing aerosol as a player in the global climate system is a challenge because absorbing aerosol is poorly understood compared to non-absorbing aerosol such as sulfate. Specifically, it is due to the fact that absorbing aerosol like black carbon absorb large amounts of incoming solar radiation and are therefore a localized

adiabatic heating source to the atmosphere, altering the temperature structure and with it the patterns of humidity, circulation and cloudiness. We will now give an overview of these effects along with other processes by which aerosol interacts with clouds and radiation.

1.2.2 Aerosol-radiation interactions

The interactions of aerosol with radiation are typically broken broken up into so-called ‘direct’ and ‘semi-direct’ effects (though we will eventually drop this terminology in favour of the AR6 parlance). The ‘direct’ effect of aerosol on radiation refers to the fact that aerosol particles increase the total atmospheric extinction in a column through scattering or absorbing incoming solar radiation. The scattering of incoming solar radiation generally leads to a cooling influence on the atmosphere, whereas for carbonaceous or ‘absorbing’ aerosol such as black carbon, these scattering processes are accompanied by an absorption of the incident solar radiation, which heats the atmospheric column and acts as a direct diabatic heating perturbation to the atmosphere (Brown et al., 2021; Stier et al., 2007).

The balance between net cooling and heating of the atmosphere by the direct effect is thus determined by the degree to which the aerosol particles are scattering versus absorbing (i.e. their SSA). However, in addition to this the balance between net cooling and heating depends on a number of other factors such as the underlying surface albedo, distribution of particle sizes and the manner in which the aerosol are mixed (in the case where there are multiple species involved) (Seinfeld and Pandis, 1998). As it is more relevant to the work in this thesis, we will focus on illustrating the joint-dependence of net cooling versus heating on the SSA and underlying surface albedo.

To gain some intuition, it is helpful to consider the extreme limits of a perfectly scattering and a perfectly absorbing aerosol layer, located over both bright and dark

surfaces. For example, the direct radiative impact of ‘purely scattering’ (SSA=1) aerosol will be negligible when located over a sufficiently bright surface (such as desert or snow) yet will exert a negative TOA radiative impact (cooling) over a dark surface. Conversely, a hypothetical ‘purely absorbing’ (SSA=0) aerosol layer over a bright surface will exert a positive TOA radiative impact (warming) and will have little impact over dark surfaces. Interpolating between these extreme limits (and noting that for absorbing aerosols the SSA₅₅₀ is typically between 0.7 and 0.9) it is clear that the direct TOA radiative impact associated with an absorbing aerosol layer should change sign from negative to positive above a certain ‘critical albedo’ (Bellouin et al., 2020; Haywood and Shine, 1995).

This concept was made mathematically rigorous for clear-sky radiative transfer by Haywood and Shine (1995), where they equivalently defined a spectrally-weighted ‘critical SSA’, ω_c , at which aerosol have the same brightness as the underlying surface and thus exert no radiative perturbation in spite of interacting with radiation. Rearranging their equation (4) (and taking the negative root, as $0 < \alpha_c < 1$) yields an equivalent formula for the ‘critical albedo’, α_c , in terms of the spectrally-weighted single scattering albedo, ω , and the spectrally weighted backscattered fraction, $\bar{\beta}$, above which the TOA radiative impact switches from negative to positive:

$$\alpha_c = 1 + \left(\frac{1 - \omega}{\bar{\beta}\omega} \right) - \sqrt{\left(\frac{1 - \omega}{\bar{\beta}\omega} \right)^2 + 2 \left(\frac{1 - \omega}{\bar{\beta}\omega} \right)} \quad (1.1)$$

Equation 1.1 is plotted in Figure 1.3 for three illustrative values of $\bar{\beta}$. A typical value of $\omega = 0.8$ for strongly absorbing aerosol is highlight in the figure and demonstrates the the critical albedo above which the aerosol have a positive radiative impact at TOA is around 0.2, though it is dependent on the value of $\bar{\beta}$.

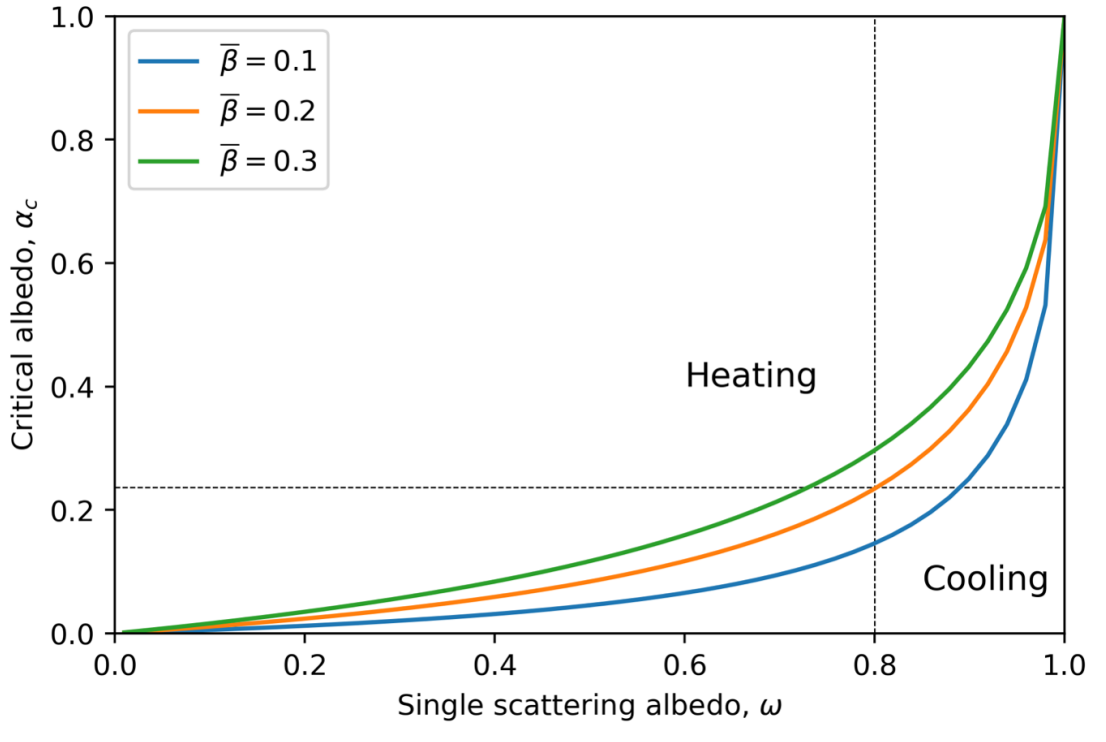


Figure 1.3: Critical albedo, α_c , at which the radiative forcing changes sign from negative to positive as a function of the single-scattering albedo, ω , for different values of the spectrally-weighted backscattered fraction, $\bar{\beta}$.

This simple framework highlights the importance of the underlying surface albedo when considering the direct radiative impact of aerosol. However, this formulation was derived assuming clear-sky conditions, which is often an inappropriate assumption. While low-level clouds occurring below the aerosol layer can be incorporated into this model quite simply assuming that they simply increase the effective albedo of the underlying surface (Stier et al., 2007), the truth of the matter is that vertical layers of cloud and aerosol need not be so distinct (Herbert et al., 2020). Overlying clouds can also modulate this behaviour by reducing the incoming sunlight available for absorption. Thus, the vertical distribution of aerosol relative to clouds is another crucial factor in determining the cloudy-sky component of direct aerosol radiative impact.

In addition to their ‘direct’ impact on radiation, aerosol (primarily absorbing) can also impact cloud properties and abundance through changing the thermal structure

of the atmosphere, the so-called ‘semi-direct’ effect (e.g., Herbert et al. (2020); Stjern et al. (2017); Koch and Del Genio (2010)). Hansen et al. (1997) initially coined the term ‘semi-direct’ effect to describe how heating in the atmosphere due to absorption may reduce ambient relative humidity, suppress convective overturning, and lead to evaporation or ‘burn-off’ of low-clouds. However, it has been shown that this effect is sensitive to the height of the aerosol layer relative to the inversion. Johnson et al. (2004) showed with a series of large-eddy simulations that absorbing aerosol below the inversion tended to cause heating and drying of the boundary layer, whereas above the inversion the heating drives an increase in low-level static stability and a moistening of the boundary layer.

Additionally, although the semi-direct effect was originally defined as reduction of clouds due to increased evaporation, there have been studies which show that absorbing aerosols can drive increases in cloud-cover (corresponding to a negative radiative forcing) through interactions with the local meteorology and large-scale circulation. For example, Randles and Ramaswamy (2008), in a sensitivity study conducted with the GFDL General Circulation Model (GCM) using fixed SSTs, found that for sufficiently large AOD, cloud amount increased as absorption increased (i.e., with decreasing SSA). This increase in cloudiness was found to be caused by an increase in low-level convergence, driven by heating of the atmospheric column, which increased upward motion and overcame the stabilization effect of the absorbing aerosol. These changes further drove increases in atmospheric water vapor, low-level cloudiness and precipitation.

Similar results were also found by Perlwitz and Miller (2010) where, using a slab ocean model, they found that for strongly absorbing aerosols, the diabatic heating in the column could drive low-level moisture convergence which was sufficient to overcome the warming effect of the absorbing aerosols on relative humidity and cause increases in specific humidity and low-level cloudiness. The authors further hypothesize that diabatic heating by absorbing aerosols in the rising branch of a

direct circulation can enhance the ascent and increase moisture convergence, nudging the system towards a state of increased cloud cover and an increased likelihood of precipitation.

Koch and Del Genio (2010), in their synthesis of the semi-direct effect, have further suggested that for convergent regions, absorbing aerosols can drive increases in cloud amount, though they admit that this is still largely uncertain due to the non-linear interactions between heating and instability in the tropics and that further sensitivity experiments are required to elucidate this effect.

In summary, while the semi-direct effect is usually thought of in terms of the local thermodynamic changes (local changes in stability and humidity driving changes at the cloud-scale), diabatic heating from absorbing aerosol can also cause dynamical changes in the large-scale circulation. When this heating is concentrated in a tropical convergence area (for example, the ascending branch of the Walker circulation), the impact of this diabatic heating is amplified by allowing the perturbation to be communicated to different regions through thermally direct circulations.

1.2.3 Aerosol-cloud interactions

Under certain meteorological conditions, some aerosol are able to be ‘activated’ and act as cloud condensation nuclei (CCN). As a result, aerosol can also ‘indirectly’ affect the properties and abundance of clouds (Bellouin et al., 2020). Generally, particles which are more hygroscopic are the most readily activated (Köhler, 1936). The efficiency with which aerosol particles can act as CCN is a function of their local environment and the ambient aerosol concentrations, N_a . For a fixed updraft velocity, w , the CCN- N_a relationship is linear in pristine conditions (where N_a is low), but as N_a increases the CCN- N_a relationship becomes sublinear and eventually levels off in sufficiently polluted environments (Chen et al., 2016). Using an ensemble of simulations with a parcel model, Reutter et al. (2009) decomposed the CCN- N_a

relationship into regimes based on the ratio of w to N_a . Briefly, the ‘aerosol-limited’ regime is characterized by high w/N_a , high supersaturation and a linear dependence of CCN on N_a , but little dependence of CCN on w . On the other hand, the ‘updraft-limited’ regime is characterized by a low w/N_a ratio, low supersaturation, a weak CCN- N_a relationship but a strong dependence of CCN on w .

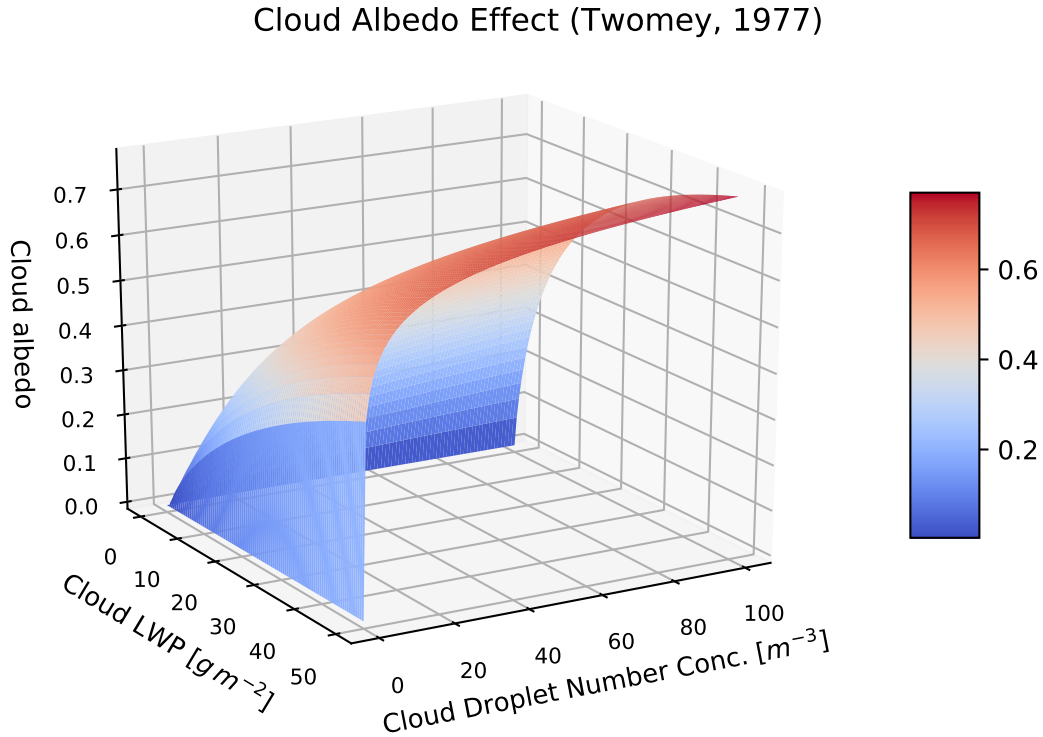


Figure 1.4: Schematic representation of the ‘cloud albedo effect’ or ‘Twomey effect’ (Twomey, 1977). The cloud optical depth and albedo are calculated following the approximations of Brenguier et al. (2011) and Seifert and Heus (2013), respectively.

One example of how aerosol may ‘indirectly’ affect Earth’s energy budget through acting as CCN is the ‘cloud albedo’ or ‘Twomey’ effect (Twomey, 1977). This states that higher CCN may lead to a greater number of cloud droplets (CDNC), which lead to a reduction in the effective radius of cloud droplets (r_e), assuming no changes in liquid water content. Figure 1.3 illustrates this for the case of an adiabatic, liquid cloud varies as a function of LWP and the cloud droplet number concentration (CDNC). Looking closely at the figure, one can see that the gradient of albedo with respect to CDNC changes is highest in pristine environments, with

low ambient number concentration (for example, over the remote oceans), whereas the albedo saturates at higher CDNC values.

In addition to indirect impacts of aerosol on cloud albedo through decreasing r_e , the same changes to CDNC and r_e may also alter cloud lifetime and extent through changes in precipitation processes (Albrecht, 1989). Because of the smaller droplet radius, cloud droplets are less likely to grow large enough to rain out as precipitation. The ‘cloud lifetime’ hypothesis states that this delayed onset of precipitation leads to longer-lasting clouds and possibly an increased cloud cover (Albrecht, 1989; Rosenfeld et al., 2002). There is some observational evidence for these effects (e.g., Christensen et al. (2020)), but the overall magnitude and sign of these effects on a global-scale is uncertain and likely dependent on the ambient environmental conditions (Ackerman et al., 2004; Dagan and Stier, 2020; Gryspeerdt et al., 2017).

1.3 Clouds

Clouds are a ubiquitous feature of Earth’s atmosphere, covering about 70% of the planet at any given time (King et al, 2013). This preponderance of cloudiness has led recent authors to remark that while the Earth is generally regarded as the ‘blue planet’, it might equally be considered the ‘white planet’ (Stephens et al., 2022). The diversity of planetary cloudiness is also astonishing, as evidence in Figure 1.5 which shows a view of the Earth taken from the geostationary satellite GOES-18. Visible in the image are frontal systems in the midlatitudes, organized deep convection along the inter-tropical convergence zone (ITCZ), trade cumulus in the tropical Atlantic and Central Pacific, and prominent stratocumulus decks which blanket the Western coasts of California and Chile.

Clouds are not just important for their beauty, they also play a number of crucial roles in the Earth system. For example, the distribution and morphology of clouds

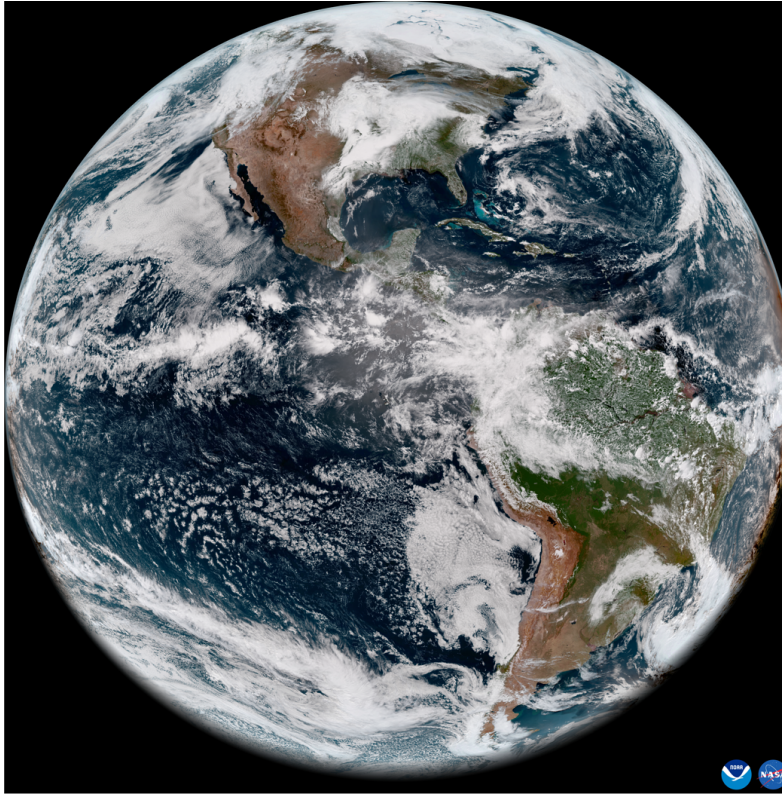


Figure 1.5: View of Earth using the Advanced Baseline Imager instrument on GOES-18. Image credit: NASA

has a strong impact on the top-of-atmosphere (TOA) radiative budget of the Earth, with their net effect involving a large degree of cancellation between their shortwave (SW) and longwave (LW) effects. In the shortwave, clouds scatter incoming solar radiation, resulting in a cooling effect of around $-40\text{--}50 \text{ Wm}^{-2}$ (Figure 1.6a). In the longwave, clouds trap infrared radiation emitted from the Earth's surface, and thus leads to a global-mean warming effect of around $25\text{--}30 \text{ Wm}^{-2}$ (Figure 1.6b). The net balance between the SW and LW impact of clouds on the TOA balance is termed the 'net cloud radiative effect' (CRE) and represents a cooling influence of around -20 Wm^{-2} in our current climate (Figure 1.6c). Due to the large magnitudes of the LW and SW CRE, there is significant potential for changes in cloud properties to alter the TOA energy balance in response both to variability and long-term climate change (termed the 'cloud feedback', which we will return to later).

The geographical distribution of the SW and LW CRE highlights two dominant

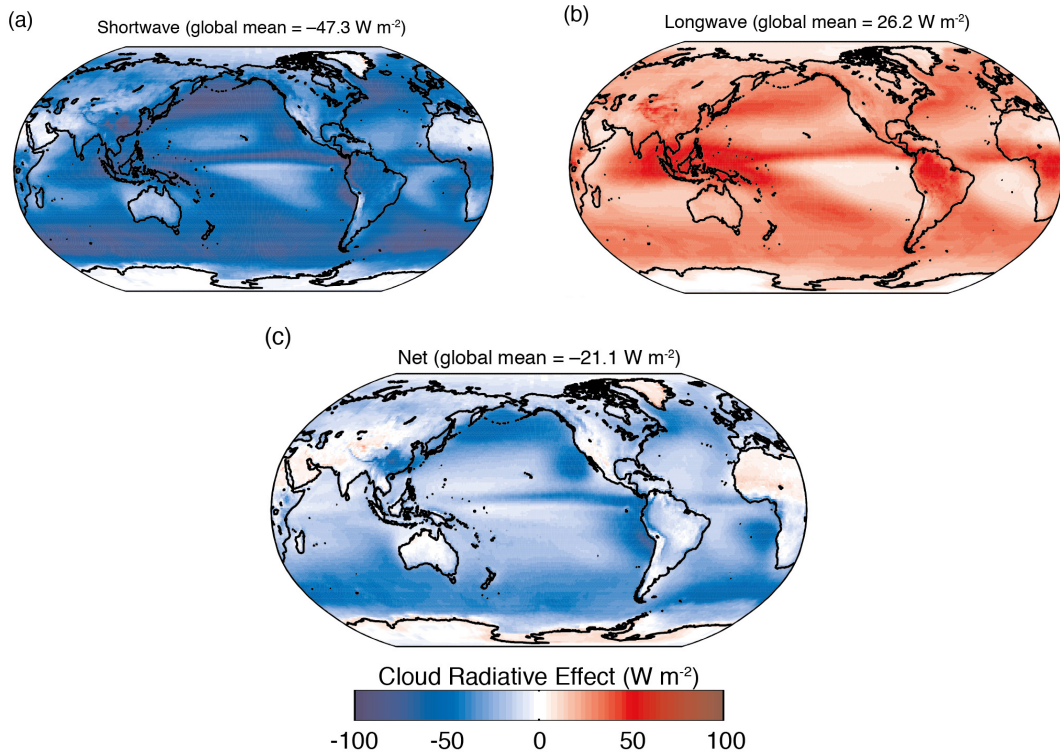


Figure 1.6: Distribution of annual-mean top of the atmosphere (a) shortwave, (b) longwave, (c) net cloud radiative effects averaged over the period 2001–2011 from the Clouds and the Earth’s Radiant Energy System (CERES) Energy Balanced and Filled (EBAF) Ed2.6r data set (Loeb et al., 2009). Adapted from Boucher et al. (2013).

cloud morphologies which are important when considering the climatological TOA balance and perturbations therein. Deep convection is associated with high clouds with cold tops, which radiate to space at much colder temperatures than the surrounding clear air and explain most of the variation in LW CRE in the tropics. The SW CRE on the other hand is sensitive to optically dense clouds at all heights, and thus exhibits a broader and less well-defined distribution. Notably, in regions of deep convection the LW and SW components of the CRE approximately cancel (Kiehl, 1994; Wall and Hartmann, 2018), such that the net global CRE is primarily set by low clouds (particularly marine stratocumulus; Figure 1.6c).

It is worth briefly discussing what processes give rise to such a diversity of cloud types, particularly in the tropics. In the tropics, the distribution of cloudiness is strongly linked to the large-scale circulation, which itself is dominated by overturning

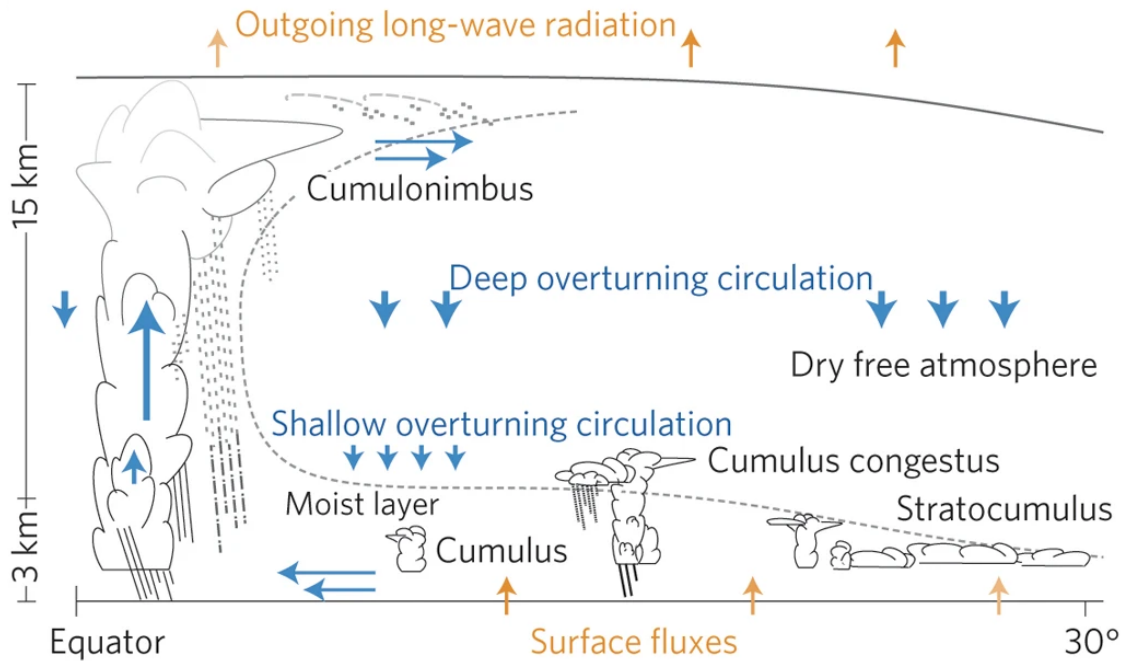


Figure 1.7: Schematic distribution of clouds and circulation in the tropics. Adapted from Bony et al. (2015).

cells such as the Hadley and Walker Circulations (Bony et al., 2006). These overturning cells are generally associated with broad regions of radiatively-driven subsidence associated with a dry free-troposphere. On the other hand, the warm and moist regions tend to be narrower and linked to large scale ascent through cloudy updrafts (Emanuel et al., 1994a). A schematic representation of the large-scale circulation of the tropical atmosphere and its connection to cloudiness is shown in Figure 1.7. The descending branches of the large-scale tropical overturning circulation tend to occur over the cold subtropical oceans, and this juxtaposition of cold SSTs and warm free-tropospheric air sets up strong inversions which promote the growth of extensive stratocumulus decks (cf Figure 1.5) which reflect large amounts of incoming solar radiation to space (Wood, 2012).

On the one hand, clouds are ‘indicators’ of the overturning atmospheric circulation. The occurrence of stratocumulus decks or deep cumulonimbus clouds tells us something about the thermal and dynamical structure of the atmosphere. However, it is increasingly understood that clouds do not simply trace out atmospheric

motions passively, but that they actively play a role in shaping these circulations (Voigt et al., 2021). Clouds primarily interact with the atmospheric circulation through releasing latent heat (either when water vapor condenses or when cloud droplets freeze), or by interacting with longwave or shortwave radiation and altering heating rates in the atmosphere (Bony et al., 2015). Through these effects, clouds influence the local temperature, stability and humidity of the atmosphere. In the tropics, because the Coriolis force is weak the atmosphere is unable to sustain large pressure gradients, and as a result any local perturbation to temperature gets rapidly homogenized across the tropics through the action of gravity waves (Bretherton and Smolarkiewicz, 1989; Pierrehumbert, 1995). This is often termed the ‘weak temperature gradient’ approximation (WTG; Sobel et al., 2001) and provides a strong constraint on the response of the tropical atmosphere to perturbations.

The strength of the WTG constraint means that clouds not only alter their *local* environment, but also the environment of the whole tropical free-troposphere (e.g., Zhang and Fueglistaler, 2020). For example, it has long been observed that deep convective clouds forming near the equator tend to relax the temperature profile of the local atmosphere to a moist adiabatic temperature profile set by the properties of the subcloud-layer air. This is the heart of the ‘convective quasi-equilibrium’ hypothesis (CQE; Betts, 1982; Raymond, 1995; Emanuel, 2007). The CQE and WTG results predict that the temperature structure of the entire tropical free-troposphere should approximately equal a moist adiabat set by the subcloud properties of deeply-convecting regions (Emanuel, 2007). Recent work has shown this prediction holds true to within 1-3 K across the tropics (Bao et al., 2022; Fueglistaler et al., 2009; Zhang and Fueglistaler, 2020), which means that by reasoning about the behaviour of deep convection near the equator we can arrive at constraints on the temperature structure of the whole tropics.

The astute reader may have noticed some slight-of-hand in the last two paragraphs, I started by talking about how clouds impact on the atmospheric circulation

and ended by concluding that deep convection provides a strong constraint on the free-tropospheric temperature profile. Worry not though, this is all for a good reason. The essential insight behind this framing is that one may conceptualize the tropical atmosphere as two ‘boxes’ (à la Pierrehumbert, 1995). In these idealized box models the circulation is partitioned into a moist, precipitating area covered by convective clouds and a dry, non-precipitating area associated with sinking motion in the mid-troposphere. The rate of subsidence in the descending box, w_{sub} , can be derived from the thermodynamic energy equation assuming no condensational heating (e.g., Jenney et al., 2020, Jeevanjee, 2022),

$$w_{\text{sub}} = \frac{\mathcal{H}_{\text{rad}} + \mathcal{H}_e}{\Gamma_d - \Gamma}, \quad (1.2)$$

where $\mathcal{H}_{\text{rad}}$ and $\mathcal{H}_e$ are the radiative and evaporative cooling rates, both negative and in units of K/s. The Γ_d and Γ terms are the dry and actual lapse rates. Many studies often ignore the $\mathcal{H}_e$ term, which is equivalent to assuming a unit ‘precipitation efficiency’ (Lutsko et al., 2022; Lutsko and Cronin, 2018). All-in-all, Eq. 1.2 demonstrates that in the absence of convective heating, the subsidence velocity in the dry, clear-sky regions of the tropical free-troposphere is largely controlled by the clear-sky radiative cooling rate and the lapse rate of temperature. Hence, given that $\mathcal{H}_{\text{rad}}$ is relatively constant at around 1-2 K day⁻¹, variations in the temperature structure of the atmosphere (and thus Γ) have a strong impact on the subsidence strength in the dry regions of the tropics. It is this reasoning which fully connects deep convection to the strength and character of the large-scale tropical circulation. Deep convection locally sets the temperature structure of the atmosphere (CQE), which is effectively transmitted across the tropics through the action of gravity waves (WTG), which then impacts the subsidence velocities in descending regions.

1.4 Precipitation

As we just saw, precipitation and the associated latent heat release in cloudy-sky is an essential element of both the dynamics of deep convection and its relationship with the tropical circulation. In light of this, we will now take a few pages to review the mechanisms by which precipitation is formed *within* clouds, and the climatological character of precipitation in observations and in the ICON global climate model which is used in this thesis.

“Precipitation forms within clouds”; an almost trivial statement at first glance, but one which hides a wealth of complexity. Even in ‘warm’ clouds (i.e. clouds without ice), the processes by which precipitation forms are myriad and complex. The initial formation of a cloud droplet depends on the availability of CCN, moisture and the attainment of a critical supersaturation. Only once these conditions are met will a cloud droplet form (Köhler, 1936). However, the typical size of a cloud drop is only around $10\mu\text{m}$, whereas raindrops are typically around 1mm in size; so, how do they grow? One option is through condensational growth, but this is a game of diminishing returns as the condensational growth rate is inversely proportional to the droplet radius, and hence this process alone is too slow to grow raindrops (Lohmann et al., 2016). Indeed the main pathways by which such large raindrops are formed in warm clouds is through either ‘autoconversion’ and ‘accretion’, which are both ‘collision-coalescence’ processes; that is, they involve the joining together of smaller elements to make a bigger element. Autoconversion refers to the earlier stages of the collision-coalescence process where small droplets collide with other small droplets, which is an effective way of generating small raindrops or ‘drizzle’ (we know all about this in England, eh?). Accretion refers to the later stage, where a large droplet begins falling through the cloud and colliding and coalescing with droplets as it falls. The efficiency with which droplets gain mass through accretion is proportional to their size, which sets up a positive feedback loop and enables rapid growth to the $\sim \mathcal{O}(1\text{mm})$ size raindrops we typically see in nature (Seinfeld

and Pandis, 1998).

The previous ‘warm rain’ processes we just covered are only a part of the story for clouds which are able to reach the freezing level, at which point clouds contain a mixture of supercooled liquid droplets and ice crystals. In this case ice processes start to also play a role in precipitation formation and as the cloud top temperature decreases, ice processes rapidly become the dominant form of precipitation initiation. The role of ice processes in precipitation formation was first explained by Bergeron (1935), who proposed that in cold clouds (i.e. the part of the cloud above the freezing level), ice crystals can grow by diffusion at the expense of supercooled water droplets until either all the drops are gone or all the ice crystals have fallen out of the cloud as rain. The essential physics of this instability comes from the fact that the saturation vapor pressure over ice is always less than over water, with a maximum difference at around -15°C , this allows for ice crystals to grow either competitively or at the expense of water droplets, depending on the ambient vapor pressure (Korolev, 2007). This proposition was based on earlier work by (Wegener, 1911) who studied ice crystal growth and was given additional observational support by Findeisen, 1938, and thus is often referred to as the ‘Wegener-Bergeron-Findeisen’ process. Below around -35°C , the freezing of cloud droplets is almost certain through homogeneous freezing, although the presence of ice nuclei can allow freezing at much higher temperature through heterogeneous freezing. In mixed phase clouds (with both ice crystals and supercooled droplets), ice crystals can also grow by colliding with each other or with supercooled liquid droplets until they reach precipitable sizes. Depending on the environmental temperature, ice crystals may also melt as they fall towards the surface.

1.4.1 Precipitation and the global energy budget

Precipitation plays an important role not only in the Earth’s hydrological cycle, but also in the atmospheric energy budget (Allen and Ingram, 2002; Dagan et al., 2021;

(Jeevanjee and Romps, 2018; Muller and O’Gorman, 2011). To see this, we can consider the vertically-integrated dry static energy budget for an atmospheric column,

$$\frac{\partial s}{\partial t} + \mathbf{u}_h \cdot \nabla s + \omega \frac{\partial s}{\partial p} = Q_R + LP + SHF \quad (1.3)$$

where $s = c_p T + gz$ is the dry static energy, Q_R is radiative heating rate, L is the latent heat of evaporation, P the surface precipitation rate and SHF the surface sensible heat flux. The Q_R term is calculated by differencing the longwave (LW) and shortwave (SW) fluxes at the TOA and surface.

On timescales longer than a month and on the scale of a GCM gridbox, the $\frac{\partial s}{\partial t}$ term can be neglected following a steady-state assumption. The contributions of the dry static energy divergence, latent and sensible heating and atmospheric radiative heating are depicted in Figure 1.8, which shows that meridional changes in zonal-mean precipitation arise primarily due to gradients in dry static energy divergence and sensible heat flux, whereas the atmospheric radiative heating is negative (indicating cooling) and it is relatively constant across latitudes.

Additionally, when integrated over the whole globe - which we denote using, $\langle \dots \rangle$ - the ‘ $\langle \mathbf{u}_h \cdot \nabla s + \omega \frac{\partial s}{\partial p} \rangle$ ’ term is identically zero (but can be retained when considering the ‘regional’ energy budget constraint on precipitation). Hence, in the global mean we have a simple constraint on global mean precipitation:

$$\langle LP \rangle + \langle Q_R \rangle + \langle SHF \rangle \approx 0 \quad (1.4)$$

In the global mean the sensible heat flux term is around 17 Wm^{-2} , and the net radiative heating of the atmosphere is around -98 Wm^{-2} , which means that $\langle LP \rangle \approx 80 \text{ Wm}^{-2}$ or around 3 mm day^{-1} (Randall, 2015).

Equation 1.4 demonstrates the essential role that latent heating via precipitation plays in the atmospheric energy balance. This is particularly true in the free-troposphere, where the atmospheric energy budget is overwhelmingly a balance

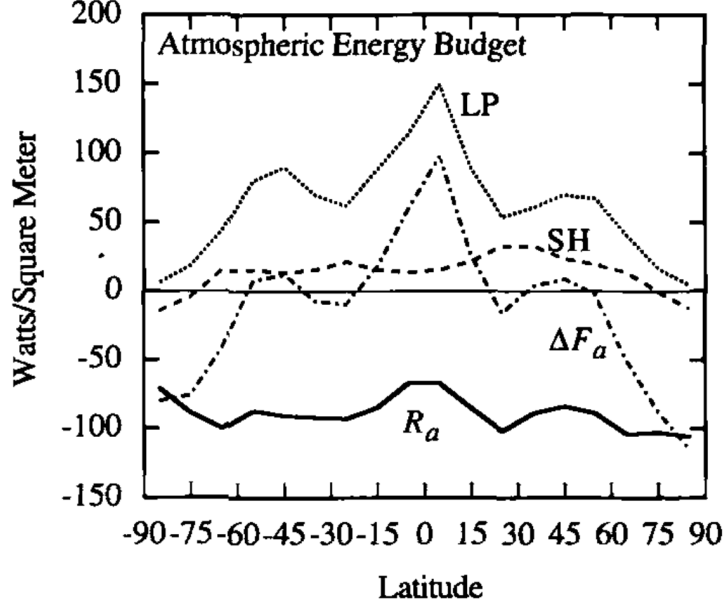


Figure 1.8: Latitudinal distribution of terms in the steady-state atmospheric energy budget. ΔF_a denotes the divergence of dry static energy, and R_a is the same as Q_R in our notation. Adapted from Hartmann (1994).

between condensation heating from precipitation and net radiative cooling (Betts and Ridgway, 1989; Takahashi, 2009). This equation also provides a strong constraint on the response of global-mean precipitation to anthropogenic and natural perturbations. In the context of global-warming, equation 1.4 provides an explanation for why global-mean precipitation only increases at $\sim \mathcal{O}(1)\%K^{-1}$, whereas the saturation vapour pressure of water increases at approximately $7\%K^{-1}$. The reason for this is because changes in radiative cooling are the dominant term on global-mean scales, and it was recently shown by Jeevanjee and Romps (2018) that $\langle Q_R \rangle$ increases only $\sim \mathcal{O}(1)\%K^{-1}$ due to the approximate *invariance* of radiative cooling profiles when expressed in temperature coordinates Jeevanjee and Romps (2018).

Anthropogenic aerosol emissions are known to play a role in determining how human activities influence precipitation through a number of pathways. Firstly, aerosol increases enhance the total atmospheric extinction in a column through scattering processes, as we saw earlier. This reduces the amount of solar radiation

reaching the surface and drives a reduction in surface fluxes, which causes global-mean precipitation to decrease due to energy budget constraints (Myhre et al., 2017; Ramanathan et al., 2001; Samset et al., 2016). Additionally, for the case of absorbing aerosols (such as black carbon, dust or absorbing organic aerosol) these scattering processes are also accompanied by an absorption of incident solar radiation, which heats the atmospheric column and acts as a direct diabatic heating perturbation (Brown et al., 2021; Stier et al., 2007; Williams et al., 2022a). This impacts on the atmospheric radiative heating through increased shortwave absorption, thus driving a decrease in global-mean precipitation.

1.4.2 Regional precipitation: climatology and change

In contrast to global-mean precipitation, which is tightly linked to *radiative* processes, the local rate of precipitation is controlled predominantly controlled by *dynamical* processes. For example, while precipitation increases with increasing Q_R in the global-mean, in the tropics, variations in precipitation and Q_R are often anti-correlated, because deep convection is the primary source of precipitation and also decreases Q_R (Naegele and Randall, 2019). Additionally, the local evaporation rate is strongly tied to the surface wind speed, which further links local precipitation to local atmospheric dynamics.

Figure 1.9a illustrates the observed climatological distribution of precipitation across the globe using data from the Global Precipitation Climatology Project (GPCP). Prominent features include the inter-tropical convergence zone (ITCZ), the extratropical storm tracks and the subtropical dry zones (dominated by shallow clouds). The distribution of precipitation closely follows the distribution of the longwave cloud radiative effect (Figure 1.6b), due to the hegemonic role of deep convection. The ITCZ is also an eye-catching feature in the zonal-mean precipitation distribution, and can be seen in the zonal-mean latent heating term

in Figure 1.8.

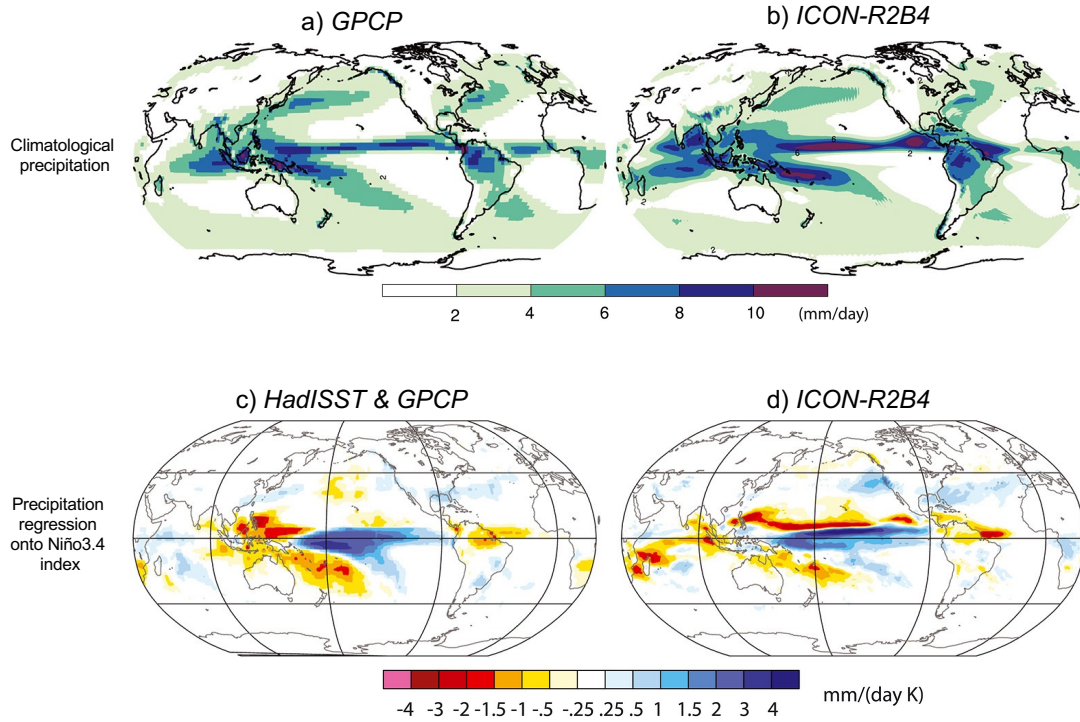


Figure 1.9: Top row: Mean precipitation climatology (1978–2013) from GPCP (panel a) and in a historical AMIP simulation using the ICON model (panel b). Bottom row: Precipitation response to the Niño 3.4 index, derived by regressing the DJF mean precipitation anomalies onto the DJF mean Niño-3.4 index anomalies using HadISST and GPCP (panel c) and the ICON model (panel d). Adapted from Crueger et al. (2018a).

When considering the response of precipitation to anthropogenic perturbations, it is typical to delineate between so-called ‘fast’ and ‘slow’ responses (e.g., Andrews et al., 2010). The ‘fast’ response refers to the changes in precipitation which occur *independently* of changes in surface temperature, whereas the ‘slow’ response refers to the component of the precipitation change which is mediated by surface temperature changes. This ‘temperature-independent’ component is often diagnosed using an atmosphere-only GCM configuration with fixed-SSTs and sea-ice, whereas the slow, ‘temperature-mediated’ component can be diagnosed using either a slab ocean or a fully-coupled atmosphere-ocean GCM. Physically, the ‘fast’ changes are due to changes in the atmospheric radiative heating profiles and associated

changes in clouds, water vapour and circulation which occur on the scale of days to weeks (Stjern et al., 2017). On the other hand the ‘slow’ component refers to the precipitation changes occurring to global and regional changes in temperature, which emerge on decadal timescales due to the large heat capacity of the oceans and the dynamics of ocean heat transport, which means the energy absorbed by the ocean is not necessarily given back in the same place as where it was absorbed (Clement et al., 1996; Randall, 2015).

Some examples of the fast contribution to precipitation changes would be; the direct effect of CO₂, which causes an instantaneous decrease in atmospheric radiative cooling and an initial decrease in precipitation (Andrews et al., 2009) along with a near-instantaneous weakening of the tropical circulation due to meridional and zonal gradients in the forcing from CO₂ (Bony et al., 2013; Merlis, 2015). The direct radiative effects of aerosol also contribute to a fast decrease in global-mean precipitation changes through increased atmospheric absorption (Samset et al., 2016). Some examples of the slow contribution to precipitation changes would be; shifts of the inter-tropical convergence zone (ITCZ) (Ming and Ramaswamy, 2009, 2011; Voigt et al., 2017; Wang, 2004, 2007, 2015; Zhang et al., 2021) or changes in monsoon precipitation (Dong et al., 2016, 2019a; Tian et al., 2018; Xie et al., 2020).

For the case of absorbing aerosol (such as black carbon), the Precipitation Driver Response Model Intercomparison Project (PDRMIP) (Myhre et al., 2017) has shown that the fast contribution dominates the global-mean precipitation response, and that this is unique among the forcings they studied (Samset et al., 2016; Stjern et al., 2017). Furthermore, the inter-model spread in fast precipitation responses to absorbing aerosol are dominated by uncertainty in the so-called ‘rapid adjustments’ of clouds and circulation to the aerosol ¹. Rapid adjustments are particularly complicated

¹We will make the ‘rapid adjustments’ term more precise in Section 2 of the Introduction, for now it is sufficient to note that the ‘fast’ changes in precipitation have contributions from radiative transfer in addition to other mechanisms such as changes in the atmospheric temperature and moisture profiles, which occur even in the absence of surface temperature changes.

for absorbing aerosol due to the fact that through absorption of incoming solar radiation, the aerosol acts as a source of diabatic heating to the atmospheric column (Johnson et al., 2019; Williams et al., 2022a). This localized diabatic heating can have significant impacts on the atmospheric thermodynamics and circulation by influencing moist processes, particularly in the tropics where clouds and circulation are intricately linked through radiative transfer, phase changes and SSTs.

Understanding the full impact of anthropogenic aerosol (and particularly absorbing aerosol) on the regional precipitation distribution is a challenging task due to the many dynamical complexities which drive regional precipitation features such as the monsoons (e.g., Dong et al., 2016) and teleconnections between regional aerosol perturbation and regional climate changes (Fahrenbach and Bollasina, 2023; Wilcox et al., 2019). Additionally, much of the previous work on the question of how regional aerosol perturbations impact on precipitation have used either emissions-driven simulations (e.g., Westervelt et al., 2018; Persad and Caldeira, 2018) which are more realistic but bring with them the uncertainties associated with the conversion of aerosol emissions to aerosol concentrations, or concentrations-driven simulations (e.g., Samset et al., 2016) which require a further region-dependent mapping from aerosol concentrations to absorptivity. Taken together, this has made it difficult to explore the fundamental mechanisms which control the atmospheric response to regional absorbing aerosol and the associated diabatic heating perturbation.

To circumvent this, there has recently been work using aquaplanet (Dagan et al., 2019) and AMIP-style (Dagan et al., 2021) simulations with prescribed aerosol radiative properties to understand the influence of aerosol-radiation interactions on fast precipitation changes - which circumvents the translation from emissions to concentrations to absorption mentioned above. Interestingly, this work has shown that the local precipitation response to absorbing aerosol is positive in the tropics and negative in the extratropics, and has explained these responses using the local atmospheric energy budget. These papers invoke the concept of ‘geostrophic

adjustment” (e.g., Vallis, 2017) to explain why the local precipitation response changes sign between the tropics and extratropics. However, only a few locations have been tested in AMIP-style simulations and thus it remains unclear whether the results found in these idealized configurations hold generally across the globe. Furthermore, Dagan et al. (2021) find significant variations in the local precipitation response to an identical absorbing aerosol perturbation even for a fixed latitude, and the reasoning for this remains unclear.

1.5 General circulation models

1.5.1 Their necessity

Modern climate science relies heavily on complex numerical models of the Earth system known as general circulation models (GCMs). These are essential in pushing forward our understanding of the climate system due to our inability to easily conduct experiments where we perturb the Earth’s climate in a known manner; other than, of course, the ‘experiment’ we are currently conducting by burning fossil fuels. However, in this case we would very much like to know the outcome of *without* seeing this experiment through to its conclusion.

General circulation models come in many flavours, and with differing levels of complexity. The most comprehensive GCMs have separate components for solving the physical equations associated with the atmosphere, ocean, ice, land surface, carbon-cycle and chemistry. The first GCM to effectively combine the atmosphere and the ocean was developed by Manabe and Bryan (1969). On the other hand, much of climate science is conducted with ‘idealized’ models where, for example, the treatment of clouds, radiation and moisture are vastly simplified, and complex processes such as atmospheric chemistry are neglected so as not to obfuscate understanding. A seminal example of this is the ‘Held-Suarez’ configuration (Held and Suarez, 1994), which only utilizes the ‘dynamical core’ of a GCM (i.e. only the

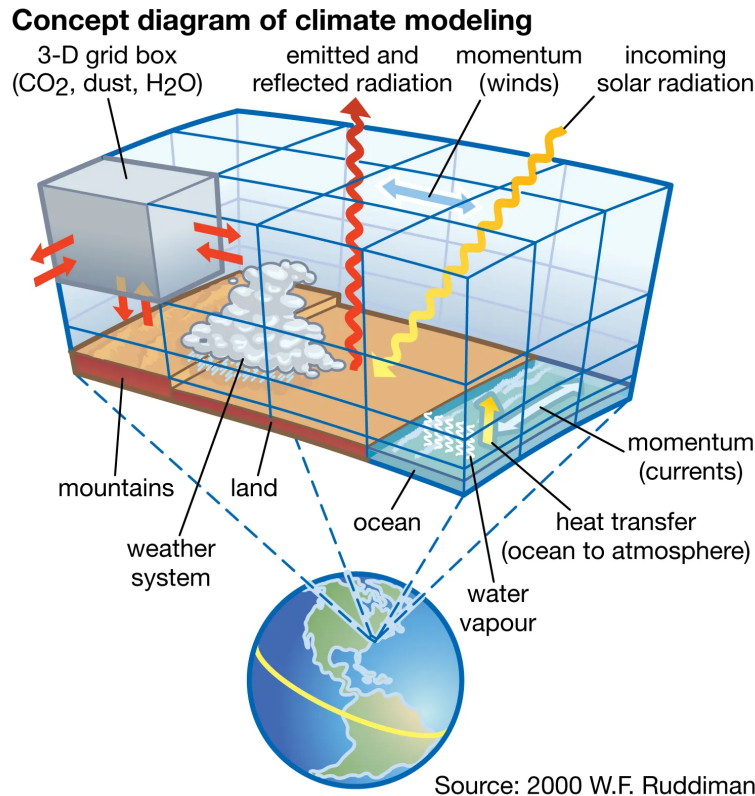


Figure 1.10: Simplified diagram of a climate model. Latest models also include complex atmospheric and ocean chemistry and land vegetation processes. Figure from Ruddiman (2001).

discretized equations of motion for the atmosphere).

GCMs allow us to determine how the basic ‘ingredients’ of Earth’s climate combine to give us the rich diversity of behaviour we see in the real world, and allow us to alter or remove these ingredients in order to gain process-level understanding and to test proposed mechanisms. For example, if the land-ocean contrast is a complicating feature for our understanding then we can simply *remove* all the land; a model set-up known as an ‘aquaplanet’, which we make use of briefly in Chapter 2. Alternatively, experiments can be conducted where we change the forcing and/or boundary conditions felt by the GCM, for example we can change the location of aerosol plumes (Chapters 2 and 3) or the size/location of SST perturbations (Chapter 4).

1.5.2 Their uncertainty

GCMs are based on three-dimensional grids of the Earth, which allow the physics to be applied on each cell, then interact with each other, as schematically shown in Fig. 1.10. However, due to limits on computational power and numerical constraints such as the Courant-Friedrichs-Lewy condition, GCMs are routinely run with a relatively coarse spatio-temporal resolution (grid-spacings of $\sim \mathcal{O}(100)$ km and time-steps of $\sim \mathcal{O}(10)$ min). As a result, GCMs only *explicitly* capture processes which happen on spatio-temporal scales much larger than this, and fail to explicitly represent processes like clouds, turbulence, radiation and microphysics. These are termed ‘sub-grid scale’ processes.

The impact of these sub-grid scale processes on the ‘resolved’ larger scale needs to be represented in order to have any hope of faithfully simulating the climate system. This is done, either partially or completely, by ‘parameterizing’ these processes on a statistical and/or physical basis. Typically, the parameterizations of sub-grid scale processes is a major source of systematic errors in weather and climate models (e.g., Bony and Dufresne (2005)). The part of the model that contains the parameterizations is referred to as the ‘physics package’.

1.5.3 Physical parameterizations in the ICON GCM

In this section we will briefly outline the physical parameterizations which are present in the ICOSahedral Non-hydrostatic (ICON) global circulation model (version id: icon-aes-1.3.00; Giorgetta et al., 2018) which I use in my thesis chapters.

The ICON model has a newly developed dynamical core and has been tuned to give a well-balanced top-of-atmosphere radiation budget and reasonable representation of precipitation compared to other atmosphere-only models and to observations (see Figure 1.9, adapted from Crueger et al., 2018a). All simulations in this thesis were run on a triangular grid (R_2B_4 specification) corresponding to an approximately

uniform horizontal grid spacing of 160km. The model uses a terrain-following vertical sigma-height grid (Leuenberger et al., 2010) with 47 levels between the surface and the model top at 83km. Because the dynamics of clouds, radiation and convection are important in the results of this thesis, we will now provide some details of the convection, cloud cover and cloud microphysics parameterizations.

The model uses a bulk mass-flux convection scheme (Nordeng, 1994; Tiedtke, 1989a), with cloud cover calculated locally using a relative humidity scheme (Sundqvist et al., 1989a). Cloud microphysics is based on the Lohmann and Roeckner scheme (Lohmann and Roeckner, 1996a), however the droplet number is fixed to prescribed profiles over land and ocean (Stevens et al. 2013) and the model does not simulate aerosol-cloud interactions. More details are given below.

Radiation is parameterized using the PSrad radiative transfer scheme (Pincus and Stevens, 2013a), which implements the gas radiative transfer and solvers from the widely used Rapid Radiation Transfer Model (Mlawer et al., 1997). I altered the radiation scheme such that it is called two times at every radiation time-step (both all- and clear-sky), once with aerosol and once without aerosol. Taking the difference between these two ‘double-calls’ gives the total aerosol radiative effect, and the instantaneous aerosol radiative forcing (IRF, defined in Section 1.4) is calculated by differencing the total aerosol radiative effect between a perturbed run and the control (although because I do not include preindustrial aerosol in any of my simulations, the aerosol radiative effect in the control run is zero).

The mass-flux convection scheme in ICON is based on Tiedke (1989) with adaptations by Nordeng (1994). It is formulated in terms of the ‘bulk plume’ equations, which assume that one can calculate height evolution of the net updraft mass flux (M_{up}) by cumulus updrafts through a balance of entrainment (E_{up}) and detrainment (D_{up}) as:

$$\frac{dM_{up}}{dz} = E_{up} - D_{up}. \quad (1.5)$$

Entrainment and detrainment can further be subdivided into contributions from turbulent entrainment/detrainment and, for the case of deep convection, a term representing ‘organized’ entrainment/detrainment. Similar equations can be written for the transport of heat, moisture, cloud water and momentum by both updrafts and downdrafts (though for brevity we only focus on the updrafts here). The cloud base mass fluxes are calculated differently for shallow and deep convection (Möbis and Stevens, 2012). For shallow convection, the mass flux evaluated at cloud base, $M_{up}(Z_{cb})$ is proportional to the moisture convergence within the sub cloud layer. For deep convection, the ICON model follows Nordeng (1994) by determining the mass flux at cloud base using a ‘quasi-equilibrium’ assumption (see Section 1.2.5). It does this by relaxing the vertically integrated buoyancy to zero on a timescale of 2 hours and assuming the convective activity balances destabilization by large-scale processes (Arakawa and Schubert, 1974). Finally, to trigger convection there must be positive moisture convergence at cloud base, specific humidity at least 1% higher in the updrafts than the environment and positive updraft buoyancy with respect to the environment. Detrainment of cloud water is used as a source term for the large-scale cloud scheme. Generation of convective precipitation in the mass-flux scheme is parameterized simply using a height-dependent ‘autoconversion’ coefficient, K , which multiplies the cloud liquid water content.

The scheme for the representation of large-scale, ‘stratiform’ clouds consists of prognostic equations for the vapor, liquid, and ice phase, respectively, a cloud microphysical scheme by Lohmann and Roeckner (1996a), and a diagnostic cloud cover scheme (Sundqvist et al., 1989a). Fractional cloud cover, C , is taken as an empirical function of grid-box mean relative humidity ($\mathcal{R}$):

$$C = 1 - \sqrt{1 - \frac{\mathcal{R} - \mathcal{R}_0}{1 - \mathcal{R}}}. \quad (1.6)$$

Here $\mathcal{R}_0$ is a critical relative humidity which is a function of height, such that $C = 0$ when $\mathcal{R} < \mathcal{R}_0$ at a given height. Condensational growth of cloud droplets occurs if $\mathcal{R} > \mathcal{R}_0$, and cloud droplets are depleted by evaporation when $\mathcal{R} < \mathcal{R}_0$.

The microphysical scheme forms precipitation in warm clouds ($T > 0^\circ\text{C}$) and also in mixed phase clouds ($-35^\circ\text{C} \leq T < 0^\circ\text{C}$) by autoconversion of cloud droplets, Q_{aut} , growth of rain drops by accretion of cloud droplets, Q_{acc} , and growth of snow crystals by accretion of cloud droplets. For temperatures below -35°C cloud liquid water is assumed to freeze homogeneously and instantaneously to cloud ice (Levkov et al., 1992), and an exponential is used to parameterize heterogeneous freezing between $-35^\circ\text{C} \leq T < 0^\circ\text{C}$. The cloud droplet number concentration (N_d) only enters through the Q_{aut} parameterization:

$$Q_{\text{aut}} \sim C N_d^{-a} r_l^b. \quad (1.7)$$

In Equation 1.7, C is the cloud fraction and r_l the grid-box mean mass mixing ratio of cloud liquid water (units of kg kg^{-1}). N_d is prescribed within the atmospheric boundary layer, with a land-ocean contrast of $N_d = 220\text{cm}^{-3}$ over land and $N_d = 80\text{cm}^{-3}$ over oceans; above the boundary layer N_d decreases exponentially to 50cm^{-3} in the upper troposphere over both land and ocean. Due to the invariance of N_d in time, this version of the ICON GCM does not simulation aerosol-cloud interactions.

Aerosol is represented simply in this version of the ICON GCM through use of the MACv2-SP (‘Simple Plumes’) model (Stevens et al., 2017). The MACv2-SP model is essentially an analytical fitting to the second version of the Max Planck Institute Aerosol Climatology (MACv2; Kinne, 2019), whose present-day (2005) distribution is anchored by surface-based observations. This done by fitting a series of two-dimensional Gaussian distributions to the horizontal extinction (denoted ϵ) distributions at 550nm over major source regions. The fitting is done on an annual-mean basis, with an idealized seasonal cycle prescribed based on whether the

plumes are classified as ‘industrial’ or ‘biomass-burning’. The vertical distribution of the aerosol is prescribed analytically with the kernel of Euler’s β -function, and is tuned to the observed vertical distributions over each source region.

To extrapolate the radiative properties of the extinction distribution to other wavelengths than 550nm, the extinction at a given wavelength (ϵ_λ) other than 550nm is related to its 550nm value (ϵ) through:

$$\epsilon_\lambda = \epsilon \exp \left(-\alpha \ln \left[\frac{\lambda \text{ nm}}{550 \text{ nm}} \right] \right), \quad (1.8)$$

where the Ångström parameter is set to $\alpha = 2$ (Seinfeld and Pandis, 1998). As we saw previously, the single-scattering albedo is also an essential parameter determining the radiative impact of aerosol, and we relate its spectral properties to the 550nm values through a cubic dependence on the wavelength. The MACv2-SP model also provides a simple parameterization of the Twomey effect (cf. Figure 1.4), which in theory can be used to provide a link between a prescribed AOD distribution and the associated change in N_d , but as my focus is on aerosol-radiation interactions in this thesis I did not include this simple Twomey parameterization.

In summary, while the ICON model does not represent all of the processes which we have established are acting in nature with respect to clouds, aerosol and precipitation, its physical parameterizations are simple and transparent compared to many state-of-the-art coupled chemistry-climate models. Hence, this GCM is an excellent tool for studying the sensitivity of climate to spatial patterns of forcing in an idealized manner.

1.6 The forcing-feedback framework

The conventional, now widely accepted framework for understanding global climate change includes an external forcing, a response in which the climate system fights the forcing to achieve equilibrium, and feedbacks that either enhance or dampen

the response (e.g., Hansen et al., 1997; Gregory et al., 2004; Sherwood et al., 2015). Such a framework can be used to assess how individual climate drivers impact on quantities such as global-mean temperature change. Assuming linearity, we can write the global-mean energy balance equation as:

$$\Delta N = \Delta F + \lambda \Delta T, \quad (1.9)$$

where ΔN is the net global top-of-atmosphere (TOA) radiation imbalance, ΔF , is the ‘radiative forcing’ and $\lambda \Delta T$ is the product of the ‘feedback parameter’, λ , with the change in global-mean surface air temperature, ΔT . This final term represents the efficiency with which the climate system can damp an energetic imbalance by warming and emitting more radiation to space.

Since the forcing-feedback framework encapsulated in Eq 1.9 became firmly established as a concept in climate science in the 1990s (Hansen et al., 1997; Ramaswamy et al., 2019), much of the literature has focused on understanding exactly *what* the ΔF and λ terms represent, and how to diagnose them in a physically-consistent and useful manner. In the proceeding sections I will provide a brief overview of the historical development of these concepts, ending with the accepted definitions as per the IPCC’s Sixth Assessment Report (AR6; Forster et al., 2021).

1.7 Radiative forcing: concept, definitions and uncertainty

The simplest way to conceptualize radiative forcing is by considering the change in the radiative budget due to a change in forcing agent, with all meteorological fields held fixed, termed the ‘instantaneous’ radiative forcing (IRF, see Fig. 1.11a). However, given that the stratosphere is largely decoupled from the land-troposphere system and thus is able to adjust radiatively on the timescale of weeks to months,

the ‘stratospherically-adjusted’ radiative forcing has traditionally been defined as the instantaneous radiative forcing plus stratospheric temperature adjustments (Myhre et al., 2013a).

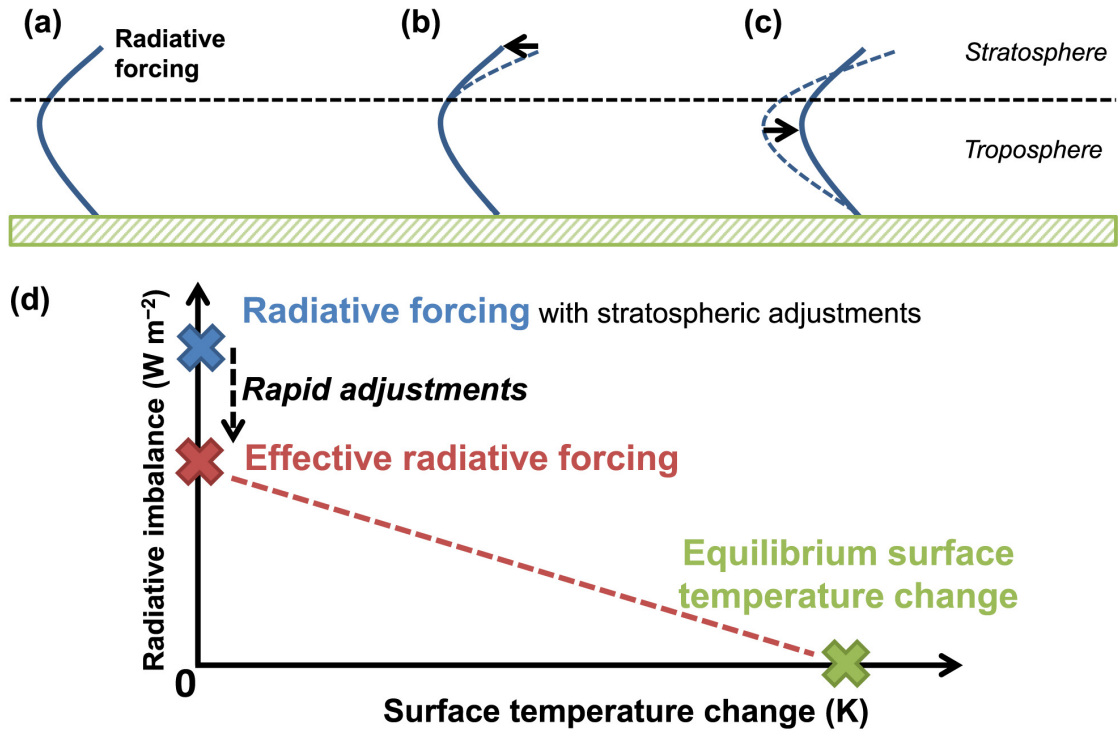


Figure 1.11: Schematic representation of the effects encapsulated by different radiative forcing definitions. Instantaneous radiative forcing (a), stratospheric adjustments (b) and fast adjustments (c) all contribute to the effective radiative forcing, illustrated in (d). Figure from Bellouin et al. (2020).

In practice, this is calculated as the net change in radiative fluxes at the climatological tropopause following the introduction of a forcing agent, allowing stratospheric temperatures to rapidly adjust to their new radiative equilibrium profile but keeping tropospheric state variables (such as temperature, water vapour and cloud-cover) constant (see schematic in Fig. 1.11b). Notably, while this method is computationally inexpensive - requiring only two offline radiative transfer calculations - the issue of rigorously defining a local tropopause adds a degree of ambiguity to its use (Myhre et al., 2013a).

Despite the utility of both the instantaneous and stratospherically-adjusted RF, neither are an accurate indicator of the temperature response for all forcing agents however, and the requirement of forcing-specific ‘efficacy factors’ has limited their usefulness (Smith et al., 2020b). This is because rapid adjustments occur not just in the stratosphere (as discussed above) but also in the troposphere. Rapid tropospheric adjustments can significantly enhance or reduce the initial flux perturbation and thus complicate the link between the initial perturbation and the resultant climate impact. The existence of these rapid, tropospheric adjustments and the problems that arise when attempting to account for them using the traditional framework has motivated the definition of the ‘effective radiative forcing’ (ERF, Sherwood et al., 2015).

As opposed to the traditional, stratospherically-adjusted RF, the effective radiative forcing includes all tropospheric and land surface adjustments which are not mediated by global-mean temperature changes. Through the inclusion of these rapid tropospheric adjustments, the effective radiative forcing has been shown to provide a better indication of the eventual climate impact of a diverse range of forcing agents (Ramaswamy et al., 2019).

The effective radiative forcing, instantaneous radiative forcing plus rapid adjustments, first introduced in AR5 has now become the dominant measure of defining the sensitivity of Earth’s energy balance to radiative perturbations (Smith et al., 2020b). However, whereas the instantaneous radiative forcing and stratospheric temperature adjustments are both reasonably well-constrained by detailed radiative transfer models and physical understanding, the rapid tropospheric adjustments are much more complicated (Forster et al., 2016). The relatively under-constrained and poorly-understood nature of tropospheric adjustments is particularly true for cloud adjustments, which result from the complex interplay between many different processes and are likely to be climate-model dependent.

Furthermore, while including rapid tropospheric adjustments to the effective radiative forcing definition has improved its usefulness as a proxy for equilibrium climate impacts (Richardson et al., 2019b), it somewhat blurs the traditional distinction between rapid climate adjustments (fast responses) and climate feedbacks (Ramaswamy et al., 2019).

Use of the term ‘rapid adjustments’ implies that processes which cause adjustments to the instantaneous radiative forcing occur on much shorter timescales than the response of global mean temperature in a coupled setup. However, there is no a priori timescale which allows us to make this distinction for all adjustment processes, as different components of the climate system have intrinsically different response times (Myhre et al., 2013a; Ramaswamy et al., 2019). Additionally, the IPCC’s Sixth Assessment Report (AR6) removed the timescale definition and instead emphasized the decoupling from surface temperature changes (Forster et al., 2021).

Another method to distinguish between feedbacks and rapid adjustments in the effective radiative forcing framework makes use of the ‘fixed-SST’ approach for calculating effective radiative forcing. In this method, two climate model simulations are conducted, both with specified SSTs and sea-ice concentrations from the present-day climatology (or the model’s own preindustrial climatology), however one run is conducted without the forcing agent. The effective radiative forcing is then calculated as the difference in net top-of-atmosphere (TOA) flux between the two runs. This approach aims to isolate the contribution of instantaneous RF and rapid adjustments by keeping global-mean temperatures constant, as required by the traditional ‘forcing-feedback’ paradigm (Sherwood et al., 2015). Additionally, these runs limit the atmosphere-surface coupling which typically mediates longer-term feedbacks, such as those involving circulation shifts (Ramaswamy et al., 2019; Soden and Chung, 2017).

The efficacy of these ‘fixed-SST’ runs for diagnosing the effective radiative forcing has recently been interrogated by Andrews et al. (2021), who examined the effective radiative forcing in a GCM with fixed land-surface temperatures, in addition to fixed SSTs and sea-ice. They found the impact of land-warming was to reduce the ERF by $\sim 1 \text{ Wm}^{-2}$ in a 4xCO_2 experiment due to its impact on temperatures, water-vapour, clouds and surface albedo, demonstrating the large impact that an interactive land-surface can have on ERF calculations. However, given the relative complexity of prescribing land-surface conditions in a full-complexity GCM, ‘fixed-SST’ runs are still be considered sufficient for many investigations.

1.7.1 Effective radiative forcing from aerosol

Among all forcing agents, the largest source of uncertainty in the effective radiative forcing of climate over the historical period is due to anthropogenic aerosol emissions (Figure 1.12). The assessed value of the total historical anthropogenic ERF is 2.72 Wm^{-2} , which represents a compensation between the positive ERF from rising greenhouse gases such as carbon dioxide and methane and the negative ERF arising from anthropogenic aerosol emissions. In total, greenhouse gases were assessed to have a total ERF of 3.84 Wm^{-2} , whereas aerosol has a total ERF of -1.1 Wm^{-2} . Aerosol-cloud interactions account for around $3/4$ of the total aerosol ERF (ERF_{ACI}), with the rest due to aerosol-radiation interactions (ERF_{ARI}). The uncertainty (90% confidence interval) in ERF_{ACI} is 1.2 Wm^{-2} , and the uncertainty in ERF_{ARI} is around 0.43 Wm^{-2} . In summary, the uncertainty in the total aerosol ERF is about twice as large as the uncertainty on the ERF from well-mixed GHGs.

In line with the discussion earlier, the ERF due to aerosol-radiation interactions is equation to the instantaneous TOA radiation change (IRF_{ARI}) and the subsequent adjustments (Adj_{ARI}) due to the interactions between the diabatic heating/cooling from aerosol and clouds, circulation and humidity in the troposphere. Similarly, the

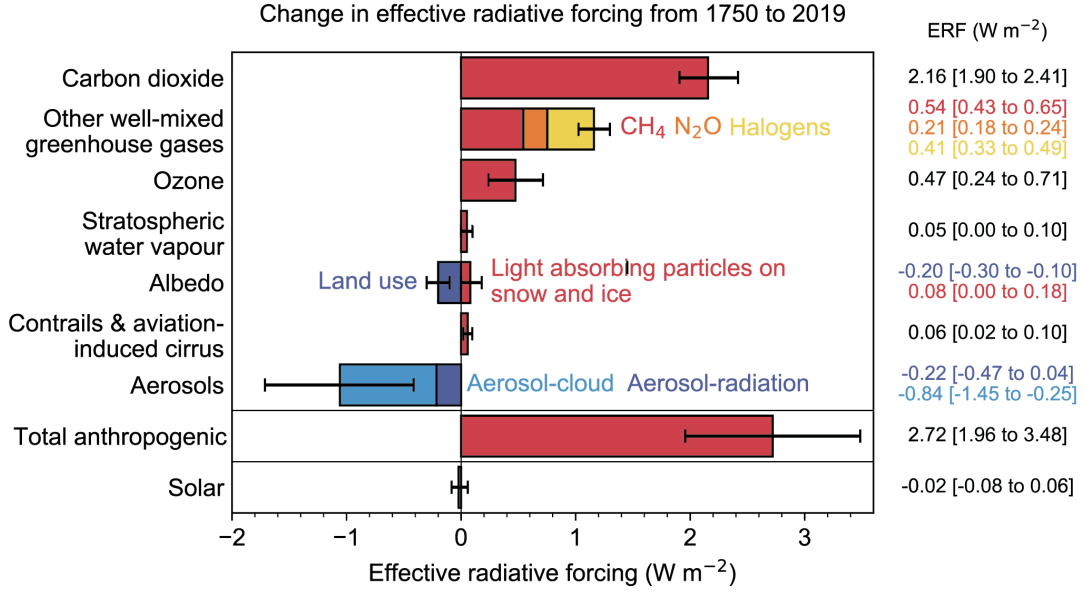


Figure 1.12: Change in effective radiative forcing (ERF) from 1750 to 2019 by contributing forcing agents (carbon dioxide, other well-mixed greenhouse gases (WMGHGs), ozone, stratospheric water vapour, surface albedo, contrails and aviation-induced cirrus, aerosols, anthropogenic total, and solar). Solid bars represent best estimates, and very likely (5–95%) ranges are given by error bars. Non-CO₂ WMGHGs are further broken down into contributions from methane (CH₄), nitrous oxide (N₂O) and halogenated compounds. Surface albedo is broken down into land-use changes and light-absorbing particles on snow and ice. Aerosols are broken down into contributions from aerosol–cloud interactions (ERF_{ACI}) and aerosol–radiation interactions (ERF_{ARI}). Figure and caption taken from Forster et al. (2021).

ERF arising from the interactions of aerosol and clouds is divided into the instantaneous component (IRF_{ACI}) due to changes in cloud droplet/ice crystal number concentrations/sizes, and a component arising from the subsequent adjustments in cloud water content or areal extent (Adj_{ACI}).

The uncertainty in the total aerosol ERF arises for a number of different reasons. In general circulation models the coarse horizontal resolution means that they poorly resolve the small-scale variability of relative humidity and aerosol amount, which together feed into hygroscopic growth and thus determine the scattering potential of the aerosol, which adds uncertainty to model estimates of ERF_{ARI} (Bellouin et al., 2020). It is also difficult to understand the sensitivity of N_d or other cloud quantities to aerosol using satellite retrievals because cloud and aerosol usually

cannot be retrieved in the same column (Gryspeerdt et al., 2015). Furthermore, determining causality of cloud-aerosol relationships from observations can suffer from ‘confounding’ factors such as the humidity swelling of aerosol (Quaas et al., 2010) or the misclassification of clouds as aerosol (Mieslinger et al., 2022).

It has also been shown that poor constraints on natural aerosol emissions in the pre-industrial period account for a large fraction of the uncertainty in aerosol ‘indirect effect’ (or the cloud albedo adjustments due to ACI, in our parlance; Carslaw et al., 2013). However, estimating the anthropogenic contribution to present-day aerosol burdens is challenging from observations (e.g., Watson-Parris et al., 2020), and also difficult using models due to the fact that not all sources and sinks of aerosol are represented in global models (Bellouin et al., 2020). Furthermore, due to non-linearities in some aerosol processes, anthropogenic emissions can affect both the production and removal of natural aerosol (Stier et al., 2006).

As a general point, understanding the relationship between aerosol emissions and the regional concentration of aerosol particles is difficult due to the fact that aerosol emissions can be transmitted thousands of kilometers before being removed from the atmosphere. It follows that the relationship between emissions and concentrations of aerosol depends on the atmospheric circulation state when the aerosol is emitted, the removal processes along its journey (e.g. precipitation), and the interactions between the aerosol and these processes.

Some studies have attempted to constrain aerosol ERF using ‘top-down’ approaches, notably the study by Stevens (2015a). This work used a simple model of aerosol forcing to argue that historical aerosol forcing more negative than -1 Wm^{-2} is inconsistent with the evolution of global temperatures, particularly over the Northern Hemisphere. However, this approach was shown to be highly sensitive to both physical (Booth et al., 2018; Kretzschmar et al., 2017) and statistical

assumptions (Albright et al., 2021; Smith et al., 2021).

One particular assumption made by Stevens (2015a) and later by Smith et al. (2021) and Albright et al. (2021) is that the *global-mean* aerosol radiative forcing is related in some simple way to the *global-mean* aerosol emissions. Such an assumption was motivated by Stevens (2015) through a figure of theirs, reproduced below as Figure 1.13. This assumes a one-to-one relationship, or equivalently that there is no dependence of effective radiative forcing on the *location* of the aerosol burden.

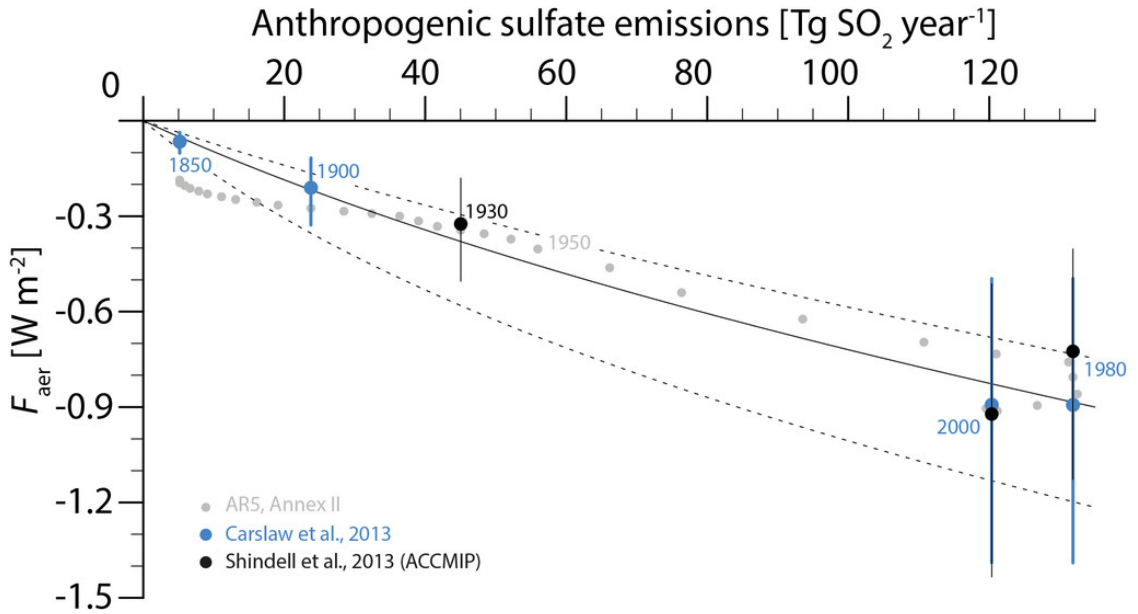


Figure 1.13: Global-mean aerosol forcing estimates versus global-mean anthropogenic sulfate emissions. Taken from Stevens (2015a).

Contrary to the ‘one-to-one’ assumption made by Stevens (2015a) and others, there are a number of reasons to expect a dependence of aerosol ERF on the location of the aerosol. For example, we have seen earlier that the instantaneous radiative forcing from aerosol depends on the effective albedo of the underlying aerosol layer, and hence for absorbing aerosol (with a single-scattering albedo substantially less than one), this means there is a ‘critical surface albedo’ above which their IRF switches from negative to positive (Haywood and Shine, 1995). Furthermore, localized aerosol burdens can influence the atmospheric circulation

(Allen et al., 2012; Dong et al., 2016; Roeckner et al., 2006), with subsequent impacts on cloudiness (Johnson et al., 2004; Kim et al., 2015). It is also known that these adjustments in cloudiness, humidity and circulation are a larger fraction of the ERF for absorbing aerosol than scattering aerosol (Smith et al., 2018a), due to the strong influence of absorbing aerosol-induced diabatic heating on moist processes (Johnson et al., 2019). Aerosol–cloud interactions are also known to depend on both the cloud regime and meteorological conditions (Dagan and Stier, 2020; Gryspeerdt and Stier, 2012). Thus, through the co-location of aerosols with different cloud types, we may also expect a location-dependence through microphysical changes. However, in this thesis our focus is on aerosol-radiation interactions and not their microphysical impacts.

Recent work using emissions perturbations in key geopolitical regions (Diao et al., 2021; Persad and Caldeira, 2018; Westervelt et al., 2020) has also shown that the global-mean ERF is a function of the location of aerosol emissions, which further supports the importance of the geographical location of aerosol in determining its ERF. However, due to differences in aerosol lifetimes and transport between regions, emissions-driven simulations can exhibit order-of-magnitude differences in regional aerosol burden, despite identical emissions (Myhre et al., 2013a; Wilcox et al., 2013). Hence, it is not clear whether the ‘location-dependence’ found in previous studies is because of differences in aerosol transport and lifetimes between regions or whether there is a more fundamental physical mechanism linking aerosol burdens in different regions to different global-mean ERF. Some of these issues can be circumvented using models driven by aerosol concentrations rather than emissions (Myhre et al., 2017; Richardson et al., 2019b). However, such approaches can still generate regional variations in aerosol radiative properties despite identical concentrations due to changes in the mixing state (Brown et al., 2021; Myhre et al., 2013a; Riemer et al., 2019; Schulz et al., 2006; Stier et al., 2006). Furthermore, previous studies have often been targeted at regions of historical emissions (Richardson et al., 2019a), or made use of climatological distributions (Stjern et al., 2017), which makes it

difficult to understand the general mechanisms by which regional aerosol impacts global-mean ERF (or other climate variables).

In summary, the effective radiative forcing from aerosol is the largest source of uncertainty in our knowledge of how anthropogenic activities have impacted global climate over the historical period, and will continue to be a source of intense uncertainty in the coming decades. The bulk of this uncertainty comes from the ‘tropospheric adjustments’ of clouds, humidity and circulation which occur in response to the aerosol perturbation. These adjustments are a particularly large fraction of the ERF for absorbing aerosol, due to the localized diabatic heating from absorbing aerosol of solar radiation which interacts strongly with local and non-local atmospheric processes, particularly in the tropics. There is an emerging body of literature which suggests that there is a strong ‘location-dependence’ of aerosol impacts on their emission region, even for global-mean quantities like global-mean ERF. However, these studies have been conducted using complex models, which involve the transport and deposition of aerosol particles and can obfuscate the essential physics. As such, it is still an open question as to whether the regional aerosol heating *alone* is sufficient to generate a location-dependence of global-mean ERF and other quantities, or whether regional differences in aerosol lifetimes and transport are required. It is this question we will come to in Chapter 2 of the thesis.

1.8 Climate feedbacks

Having now covered the essentials of the ‘forcing’ term in Equation 1.9, we now come to the ‘feedback parameter’, λ . As noted in Section 1.6, the feedback parameter represents the efficiency with which the climate system damps TOA energy imbalances through altering the global-mean temperature. The net global feedback parameter must be negative in order to maintain a stable climate, with a more-negative feedback parameter implying that the Earth needs to warm less to

balance an imposed ΔF .

To first-order, λ can be decomposed into a sum of different terms,

$$\lambda = \sum_x \frac{\partial N}{\partial x} \frac{\partial x}{\partial T}, \quad (1.10)$$

where x refers to a climatic variable which has a direct effect on the TOA energy budget. The components of λ depend on precisely which feedback decomposition is used, but Held and Shell (2012) (among others) have argued for the use of ‘relative humidity (RH)-based feedbacks’. In this framework, λ is the sum of feedbacks associated with uniform vertical warming (Planck feedback), nonuniform vertical warming (lapse rate feedback), and changes in relative humidity, surface albedo, and clouds (Held and Shell, 2012; Soden et al., 2008). The sum of these individual feedbacks (λ_x) must yield a net global feedback parameter which is negative in order to maintain a stable climate, with a negative λ_x termed a ‘stabilizing’ feedback, whereas a positive λ_x is a ‘destabilizing’ feedback.

Using a radiative kernel approach, Zelinka et al. (2020) decomposed the total feedback parameter into the contributions from Planck, lapse rate (LR), relative humidity (RH), albedo and shortwave/longwave cloud feedbacks under an abrupt-4xCO₂ scenario, in both CMIP5 and CMIP6 models. Their results are shown in Figure 1.14. The two most striking features of Figure 1.14 are that: 1) the Planck, lapse rate and relative humidity feedbacks enjoy robust agreement amongst models and 2) that models disagree strongly on the sign of the cloud feedback under warming, with a moderately larger spread in shortwave (SW) cloud feedbacks than longwave (LW) cloud feedbacks. In particular, the increased spread in SW cloud feedbacks in CMIP6 relative to CMIP5 has been implicated in the high climate sensitivity of some CMIP6 models.

Generally, the λ_x values are assumed to be constant over a range of climate states, including past climate states like the Last Glacial Maximum and the Eocene,

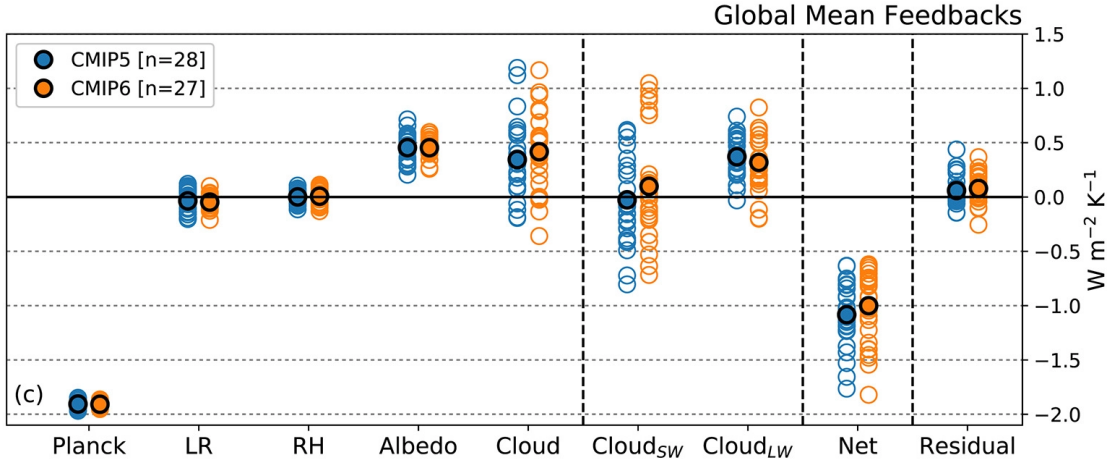


Figure 1.14: Climate feedbacks in CMIP5 and CMIP6 models, calculated using a RH-based radiative kernel decomposition of the abrupt-4xCO₂ scenario. Adapted from Zelinka et al. (2020).

or future climate states we will encounter in the coming century. However, there is now mounting evidence that the climate feedback parameter actually *evolves* over time in response to the evolving patterns of surface warming (Forster et al., 2021), a concept referred to as the ‘SST pattern effect’ (Stevens et al., 2016). In particular, the role of the tropical Pacific sea-surface temperatures has been highlighted, and will be our focus here.

1.8.1 The tropical SST pattern effect

Research over the past decade has identified how changes in tropical SSTs, particularly in the Pacific basin, may exert a strong control over the temporal evolution of climate feedbacks (e.g., Zhou et al., 2017b; Armour, 2017; Andrews and Webb, 2018; Ceppi and Gregory, 2019). In particular, changes in the global TOA radiative energy budget can be induced by spatial changes surface temperature, even without a change in the global-mean (Ceppi and Gregory, 2017; Zhou et al., 2016). Hence, λ depends on the spatial pattern of warming. This ‘SST pattern effect’ (Stevens et al., 2016) can result from either forced changes in the climate system or from internal variability. The SST pattern effect has received considerable attention within the community due to its implications for reconciling values of ECS inferred from the

historical record with values of ECS inferred based on other lines of evidence and simulated by climate models (Andrews et al., 2018; Armour, 2017; Proistosescu and Huybers, 2017).

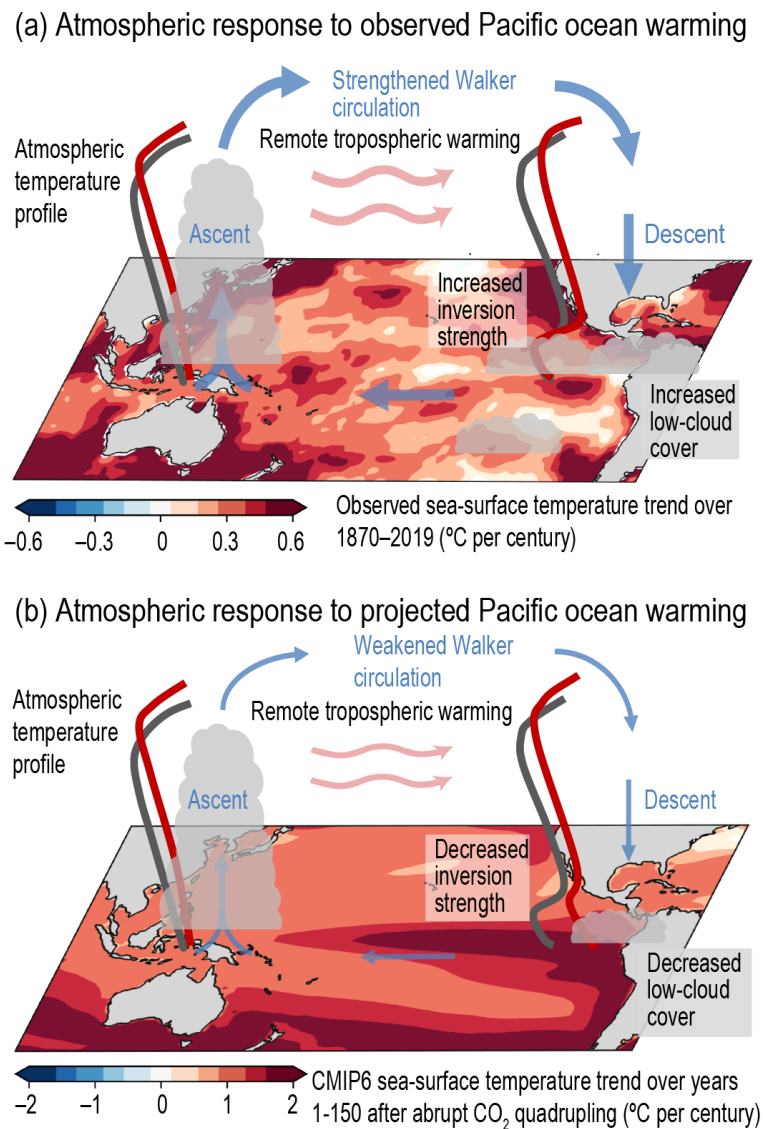


Figure 1.15: Schematic of the physical mechanisms involved in the response to: (a) linear sea surface temperature trend observed over 1870–2019. (b) linear sea-surface temperature trend over 150 years following abrupt $4\times\text{CO}_2$ forcing as projected by CMIP6 ESMs. From Forster et al. (2021)

Both modelling (Andrews and Webb, 2018; Dong et al., 2019b; Zhou et al., 2017b) and observational (Fueglistaler, 2019; Mackie et al., 2021) studies have suggested that the primary mechanism mediating the SST pattern effect is its influence on low clouds in regions of climatological subsidence, and such explanations typically proceed as follows:

- Regions with a high climatological low-cloud amount are generally associated with cold SSTs and a strong inversion (Wood and Bretherton, 2006), thus when SSTs in these regions are warmed the inversion weakens and there is a decrease in low-level cloudiness. This leads to a positive shortwave TOA (SW_{TOA}) anomaly through reduced low-level reflection of incoming solar radiation. The reverse-argument also holds, namely that isolated cooling in these regions strengthens the inversion and increases low-level cloudiness. We term this the ‘local stability-inversion’ mechanism.
- On the other hand, warming in regions of deep convection is communicated vertically throughout the free-troposphere by deep convection (Zhang and Fueglistaler, 2020) and then horizontally across the tropics through the action of gravity waves (Bretherton and Smolarkiewicz, 1989; Charney, 1963; Neelin and Held, 1987; Pierrehumbert, 1995). This causes a remote warming of the free-troposphere over the aforementioned low-cloud regions which again strengthens the inversion and increases the low-cloud amount (Andrews and Webb, 2018). We term this the ‘non-local stability-inversion’ mechanism.

Warming of the free-troposphere has also been shown to affect the TOA flux response through the lapse-rate feedback (Andrews and Webb, 2018; Ceppi and Gregory, 2017), in addition to the cloud feedbacks which we highlight here. However, when a relative humidity-based decomposition (Ceppi and Gregory, 2017; Held and Shell, 2012) is used (which accounts for the partial cancellation between lapse-rate and water-vapour feedbacks; Dong et al. (2019b)), the lapse-rate feedback has been found to be smaller in magnitude than the cloud radiative feedback.

The mechanisms and radiative effects of these changes are illustrated in Figure 1.15, for the cases of observed Pacific ocean warming, which is ‘La Niña-like’ insofar as the zonal SST gradient has strengthened over the 20th century, and projections of Pacific ocean warming from CMIP6 abrupt-4xCO₂ experiments, which is ‘El Niño-like’ in that it demonstrates a weakening of the zonal SST gradient.

1.8.2 SST Green’s Function approaches

To put the SST pattern effect on a more quantitative footing some studies have framed the problem in terms of Green’s functions (Baker et al., 2019; Dong et al., 2019b; Li and Forest, 2014; Zhou et al., 2017b) by assuming there exists some operator, G , which maps the spatial pattern of SST anomalies onto TOA anomalies. In this framework, G can be estimated using ensembles of isolated ‘SST patch’ experiments in an atmosphere only model and then used to reconstruct the TOA response to arbitrary SST patterns. An example of the Green’s Functions generated by this approach is given by Figure 1.16, from Dong et al. (2019b). Figure 1.16a shows that the Green’s Function approach captures the negative TOA flux changes in response to warming in the western tropical Pacific, in addition to the positive TOA response to warming in the eastern tropical Pacific.

Although the Green’s Function approach is an appealing concept, a potential issue is that the Green’s function approach is fundamentally linear (Riley et al., 1999) and requires assuming that the TOA response is linear with respect to the sign and magnitude of the imposed SST anomaly, and also that SST anomalies in different regions combine linearly. Previous studies have made these assumptions (Baker et al., 2019; Barsugli and Sardeshmukh, 2002; Dong et al., 2019b; Li and Forest, 2014; Zhou et al., 2017b), but given the well-appreciated ‘threshold’ behaviour of deep convection (Emanuel, 2007; Johnson and Xie, 2010; Williams and Pierrehumbert, 2017; Xie et al., 2010; Zhang, 1993; Zhang and Fueglistaler, 2020), along with the observed asymmetry in the atmospheric response to positive and

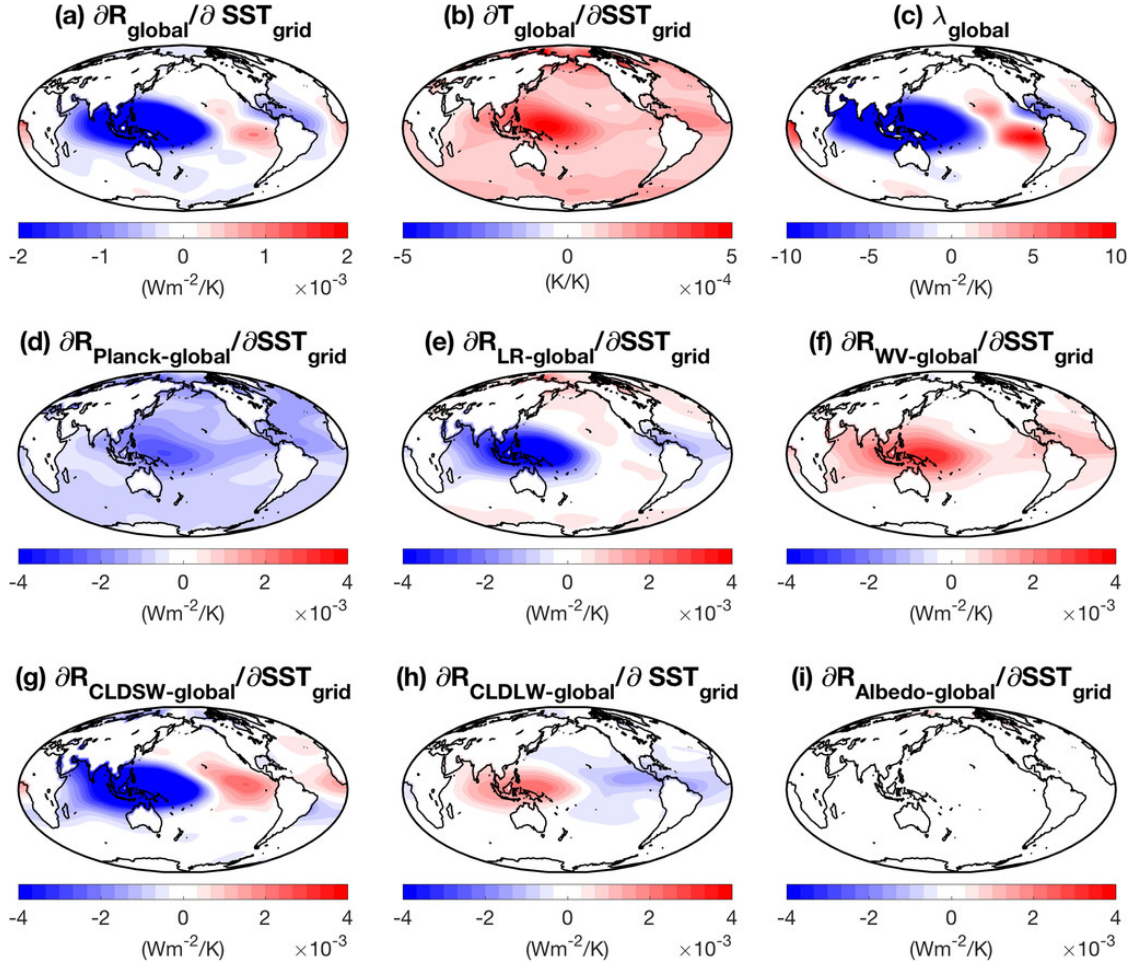


Figure 1.16: Response of annual- and global-mean of (a) net TOA radiation, (b) near-surface air temperature, and (c) feedback per unit SST warming in each grid box. The units are $\text{W m}^{-2} \text{K}^{-1}$, K K^{-1} , and $\text{W m}^{-2} \text{K}^{-1}$, respectively. The global-mean TOA radiation change shown in (a) is decomposed using radiative kernels into contributions due to (d) Planck, (e) lapse-rate, (f) water vapor, (g) cloud SW radiation, (h) cloud LW radiation, and (i) albedo changes. The units for (d)–(i) are $\text{W m}^{-2} \text{K}^{-1}$. Figure from Dong et al. (2019b).

negative ENSO phases (Hoerling et al., 2001, 1997; Johnson and Kosaka, 2016), it seems likely that the linear assumption is not valid.

Recent work has also found that the SST Green’s Functions themselves are sensitive to the sign and magnitude of the SST perturbations imposed during their generation via atmosphere-only experiments (Zhang et al., 2023). Despite this, no studies have systematically investigated under what conditions the response is linear and additive. Furthermore, the conceptual model proposed in Figure 1.15, which

forms the basis for the pattern effect discussions in the latest IPCC report (Forster et al., 2021), does not make any predictions related to the sign or magnitude of the perturbations. Indeed, even simple questions like “What should the TOA response to *cooling* in the western Tropical Pacific (as opposed to warming) look like?” are currently unanswered in the pattern effect literature.

1.9 Structure of the thesis

We have now seen that Earth’s present day climate exhibits rich geographical variations; from the heterogeneous distribution of clouds and radiation to the striking regional features of precipitation. We have also seen that anthropogenic perturbations that *themselves* are rich in geographical diversity. Hence, this thesis argues that to fully understand the influence of anthropogenic perturbations on Earth’s climate, we must convolve the regional with the regional. In other words, we must develop an understanding of how *regional perturbations* influence on global and local climate.

In answering this question, we have chosen to focus down on two particular forms of regional perturbation, namely localized changes in absorbing aerosol and in sea-surface temperatures. This is motivated by the fact that, even in the climatology, distributions of anthropogenic aerosol are highly spatially heterogeneous due to the short lifetimes of atmospheric aerosol. In addition to this, historical trends in aerosol composition have demonstrated an increasing trend in absorbing aerosol, which is projected to increase in importance over the coming decades. Alongside this, the spatial patterns of sea-surface temperature, particularly in the tropics, have been shown to exert a strong influence over the Earth’s TOA radiation balance; a phenomenon termed the ‘SST pattern effect’. The SST pattern effect is particularly noteworthy as it shows that different spatial patterns of temperature change can impact on the global-mean climate, even if those changes are zero in the global-mean.

The aim of this is thus to improve our understanding of how spatial patterns of absorbing aerosol and sea-surface temperatures impact on Earth’s climate on both global and local scales. To do this, we are using various configurations of the ICON GCM (Section 1.5) in order to study these problems in an idealized setting where only the essential aspects of the problem are retained; that is, we aim to make things ‘as simple as possible, but no simpler’. Our focus over the next three chapters will be on three scientific questions:

Q1. Does the global-mean ERF from aerosol-radiation interactions depend on the location of absorbing aerosol? If so, why?

The effective radiative forcing (ERF) of anthropogenic aerosol represents the largest uncertainty in our understanding of how human activity has influenced Earth’s climate over the historical period, and will continue to be a source of uncertainty for decades to come as the spatial pattern and composition of aerosol shift in response to air quality policies. Many studies have attempted to constrain the historical ERF of aerosol in a ‘top-down’ way by combining observations of Earth’s energy budget with a simple model of aerosol forcing which assumes that there is a *one-to-one* correspondence between global-mean aerosol burden and global-mean aerosol ERF (Albright et al., 2021; Smith et al., 2021; Stevens, 2015a). In other words, this simple model assumes there is no ‘location-dependence’ of global-mean aerosol ERF on the location of the aerosol burden.

Contrary to this assumption, recent work has shown that the global-mean ERF of aerosol emissions depends on the location where they are emitted (Persad and Caldeira, 2018; Westervelt et al., 2020). However, these studies have not specifically studied absorbing aerosol, which is known to drive strong adjustments of clouds, humidity and circulation through the absorption of incoming solar radiation and subsequent diabatic heating of the atmosphere. Additionally, these studies have used emissions-driven simulations, which make it difficult to understand whether

this location-dependence arises due to differences in aerosol lifetimes and transport between regions, or whether it is due to more fundamental physical mechanisms linking regional aerosol heating to global-mean aerosol ERF. Furthermore, previous work has often been targeted at regions of historical emissions, or made use of climatological distributions, and thus have only sampled a small amount of the potential co-variability between aerosol heating and regional climate properties.

In Chapter 2, I use a large ensemble of atmosphere-only ICON simulations with identical absorbing aerosol perturbations in different geographical locations to investigate whether a ‘location-dependence’ of global-mean ERF can occur even in the absence of changes in aerosol lifetimes, mixing ratios or transport between regions. To understand these results we focus on the interactions between cloudiness, atmospheric circulation and the diabatic heating resulting from the localized absorbing aerosol.

Q2. How do fast-changes in global and local precipitation depend on the location of absorbing aerosol?

Understanding the impact of anthropogenic aerosol on precipitation is critical for understanding the pattern of climate impacts over the coming decades. The precipitation response to anthropogenic activity is typically separated into ‘fast’ and ‘slow’ components based off of whether SSTs are allowed to adjust to the perturbation (slow) or not (fast). Previous work has shown, for the case of absorbing aerosol, that the fast contribution dominates the global-mean precipitation response (Myhre et al., 2017), and that this is unique among the forcings they studied (Samset et al., 2016; Stjern et al., 2017). It was also shown that the ‘adjustments’ of clouds, humidity and circulation were the dominant contributor to inter-model spread in the precipitation response across models (Stjern et al., 2017).

On global-mean scales, aerosol impacts precipitation in a manner which is well-constrained by the global-mean atmospheric energy budget. Recent studies have also assumed that there is a unique, linear relationship between global-mean precipitation and global-mean absorbing AOD (McCoy et al., 2022; Samset, 2022). However, we will see in Chapter 2 that even on global-mean scales, the perturbation to Earth’s energy budget from a given burden of absorbing aerosol depends strongly on where that aerosol is located. Hence it is important to investigate whether this relationship holds, and if not then to understand what physical mechanisms cause it to falter.

On regional scales local changes in the atmospheric circulation play a dominant role in determining the precipitation change. The dominant role of circulation changes has been highlighted by recent work (Dagan et al., 2019, 2021), who demonstrated a ‘tropical-extratropical’ contrast in the local precipitation response to absorbing aerosol. However, it is not yet clear whether this result holds generally across the globe and there is currently no explanation for what may control *variations* in the local precipitation response between regions, particularly in the tropics. As such, we specifically focus on understanding how the atmospheric circulation (and subsequently cloudiness and humidity) respond to localized absorbing aerosol in a fixed-SST framework, to better understand the impact of absorbing aerosol on precipitation.

Building on the work from Chapter 2, in Chapter 3 we focus on how the geographical location of an idealized absorbing aerosol plume modulates its impact on global and local precipitation. This chapter makes use of the same ensemble of aerosol perturbations as outlined in Chapter 2. We use these experiments to assess whether the ‘tropical-extratropical’ contrast in the local precipitation response to aerosol holds at all locations across the globe, and also assess whether the assumed one-to-one relationship between global-mean AOD perturbation and global-mean precipitation changes hold. In both cases, we form simple physical theories to

explain the results.

Q3. Is the global-mean TOA radiation response to an SST perturbation linear with the respect to the sign and magnitude of the perturbation? What implications does this have for Green’s function approaches?

The SST pattern effect has recently emerged as a prominent example of how spatial patterns of external perturbations can be extremely important for understanding their climatic impact. In particular, understanding how spatial patterns of tropical SST anomalies impacts on Earth’s TOA radiation budget has proved crucial for reconciling disparate climate sensitivity estimates in the literature (Andrews et al., 2018).

Previous work has often attempted to quantify the dependence of global-mean TOA radiation on SST patterns using a Green’s function approach (Dong et al., 2019b; Zhang et al., 2023; Zhou et al., 2017b). However, Green’s functions are *linear* objects, and thus the SST Green’s function approach requires assuming that the TOA radiation response to local SST perturbations is *linear* with respect to the sign and magnitude of the SST perturbation. Furthermore, an assumption of ‘additivity’ is also made, which allows one to calculate a Green’s function by just perturbing individual locations in isolation. The question of whether these assumptions actually *hold*, and what implications this has for the SST Green’s function approach, have not yet been examined.

To answer these questions, in Chapter 4 we turn our attention to the problem of spatial patterns of SST and how they impact on the global-mean TOA radiation budget. The goal in this chapter is two-fold; firstly, to evaluate the assumptions of ‘linearity’ which are essential to prior SST Green’s function studies, and secondly, to explain why deviations from linearity occur by forming a more robust conceptual

model of the SST pattern effect.

The results from individual chapters are summarised in Chapter 5, and potential future work is discussed.

Nigel Tufnel: What we do is, if we need that extra push over the cliff, you know what we do?

Marty DiBergi: Put it up to eleven.

Nigel Tufnel: Eleven. Exactly. One louder.

Marty DiBergi: Why don't you just make ten louder and make ten be the top number and make that a little louder?

Nigel Tufnel: [pause] These go to eleven.

— Spinal Tap

2

Sensitivity of effective radiative forcing to the spatial pattern of absorbing aerosol

The effective radiative forcing (ERF) of anthropogenic aerosol represents a large source of uncertainty in our understanding of how human activity has influenced Earth's climate over the historical period. As discussed in the Introduction, many studies have taken a 'top-down' approach to constrain the aerosol ERF which assumes that there is a one-to-one relationship between global-mean aerosol ERF and global-mean aerosol emissions. This is equivalent to assuming that there is no 'location-dependence' in the link between aerosol emissions and global-mean aerosol ERF.

Recent studies have shown that the global-mean ERF of aerosol emissions depends on the location where they are emitted (Persad and Caldeira, 2018; Westervelt

This chapter is based on the following paper, published in Nature Climate Change:

Williams, A. I. L., Stier, P., Dagan, G. & Watson-Parris, D. "Strong control of effective radiative forcing by the spatial pattern of absorbing aerosol" *Nat. Clim. Chang.* 12, 735–742 (2022). <https://doi.org/10.1038/s41558-022-01415-4>

I designed the study together with the co-authors. I carried out the simulations and analysed the data with input from the co-authors. I wrote and revised the manuscript in response to reviewer comments, with contributions from the co-authors.

et al., 2020). This calls into question the implicit assumptions behind previous ‘top-down’ constraints on the aerosol ERF. However, no robust physical mechanism has been put forward to explain this ‘location-dependence’. Furthermore, these prior studies have used emissions-driven simulations, which raises the question of whether this ‘location-dependence’ requires the modeled translation from emissions to concentrations, or whether it can be achieved in a simpler setting.

Hence, in this Chapter, we aim to answer the question of whether global-mean effective radiative forcing is sensitive to the geographical location of absorbing aerosol, even in the absence of changes in aerosol lifetimes, mixing ratios or transport between regions. To do this, we conduct a suite of idealized experiments in an atmosphere-only GCM, where identical absorbing AOD perturbations are placed systematically across the globe. We analyse the variations in global-mean ERF between these experiments, and provide simple physical explanations to explain the ‘location-dependences’ we find. The results of this chapter help us to better understand the ERF from absorbing aerosol perturbations and the dependence of these effects on both the latitudinal and longitudinal position of the aerosol plume.

2.1 Introduction

Anthropogenic aerosol plays a substantial role in determining the influence of human activities on climate (Lund et al., 2019; Naik et al., 2021; Samset et al., 2019). Due to their short lifetimes, the spatial pattern of anthropogenic aerosol is highly heterogeneous, and this has been shown to impact the climate through changes in atmospheric circulation (Allen et al., 2012; Chemke and Dagan, 2018), tropical precipitation (Chemke and Dagan, 2018; Chung and Soden, 2017; Deser et al., 2020; Voigt et al., 2017) and sea-surface temperature (SST) patterns (Dagan et al., 2020; Deser et al., 2020; Fiedler and Putrasahan, 2021), with implications for the transient climate response (Shindell, 2014).

2. Sensitivity of effective radiative forcing to the spatial pattern of absorbing aerosol

Over the coming decades, emissions of anthropogenic aerosol are expected to change dramatically both in their spatial pattern and composition (Lund et al., 2019), particularly over South-East Asia and Central Africa (Lund et al., 2019; Samset et al., 2019). These trends are further complicated by the fact that policy measures often target sources of scattering and absorbing aerosol differently, such that even with no net change in aerosol optical depth (AOD), there could be climatic impacts from changes in aerosol composition (Samset, 2018). Such a trend is already being seen globally; while the global anthropogenic AOD has been roughly constant since the 1980s, the contribution from absorbing aerosol has almost doubled over this period (Hoesly et al., 2018; Lund et al., 2019). In the future, particularly under weak air-quality policies, these trends towards a more absorbing aerosol distribution will probably continue (Samset et al., 2019).

Given these observations, a key question arises: how does the evolving spatial pattern of anthropogenic aerosol (particularly absorbing aerosol) impact climate? Crucially, how does the global-mean effective radiative forcing (ERF) — defined as the net change in radiative fluxes at the top-of-atmosphere in a fixed-SST experiment following a change in forcing agent, allowing for rapid adjustments (Smith et al., 2020a) — respond to this changing pattern of absorbing aerosol? A robust answer to this question would have substantial ramifications because the impact of the spatial pattern of absorbing aerosol on global-mean ERF is not considered either in the simple energy-balance models (Smith et al., 2018b; Stevens, 2015b) or in integrated assessment models (Nordhaus, 2014) which form an essential part of the IPCC process (Arias et al., 2021).

Although the magnitude of, and mechanisms for, a dependence of ERF on aerosol location are not yet clear, there are reasons why we might expect location-dependence. For instance, the dependence of instantaneous radiative forcing (IRF) on the effective albedo of the below-aerosol layer means that for absorbing aerosol there is a ‘critical surface albedo’ above which their IRF switches from negative to

positive (Bellouin et al., 2020; Haywood and Shine, 1995). Additionally, localized absorbing aerosol (and scattering aerosol) can influence atmospheric circulation (Allen et al., 2012; Dagan et al., 2019; Dong et al., 2014, 2016; Fiedler and Putrasahan, 2021; Roeckner et al., 2006; Wilcox et al., 2019; Zhou et al., 2020), with subsequent impacts on regional precipitation (Chemke and Dagan, 2018; Dong et al., 2014, 2016) and cloudiness (Smith et al., 2018a; Stjern et al., 2017). The component of ERF arising from changes in clouds and circulation is referred to as the ‘rapid adjustments’ (Adj) and is known to be a larger fraction of the ERF for absorbing aerosol than scattering aerosol (Smith et al., 2020b) due to the strong influence of absorbing aerosol-induced diabatic heating on moist processes (Johnson et al., 2019). However, understanding the link between absorbing aerosol, circulation and ERF remains challenging (Smith et al., 2018a). Aerosol–cloud interactions are also known to depend on both the cloud regime and meteorological conditions (Altaratz et al., 2014; Dagan and Stier, 2020; Gryspeerdt and Stier, 2012). Thus, through the co-location of aerosols with different cloud types, we may also expect a location-dependence through microphysical changes (however, this pathway is not examined here).

Recent work using emissions perturbations in key geopolitical regions (Diao et al., 2021; Persad and Caldeira, 2018; Westervelt et al., 2020) has also shown that the global-mean ERF is a function of the location of aerosol emissions. However, due to differences in aerosol lifetimes and transport between regions, emissions-driven simulations can exhibit order-of-magnitude differences in regional aerosol burden, despite identical emissions (Myhre et al., 2013a; Schulz et al., 2006; Wilcox et al., 2013). Hence, it is not clear whether this ‘location-dependence’ is because of differences in aerosol transport and lifetimes between regions or whether there is a more fundamental physical mechanism linking absorbing aerosol in different regions to different global-mean ERF. Furthermore, perturbing emissions of different aerosol species simultaneously (Diao et al., 2021; Persad and Caldeira, 2018) makes it difficult to attribute the changes to absorbing aerosol alone. Some of these issues

can be circumvented using models driven by aerosol concentrations rather than emissions (Myhre et al., 2017; Richardson et al., 2019a). However, such approaches can still generate regional variations in aerosol radiative properties despite identical concentrations due to changes in the mixing state (Brown et al., 2021; Myhre et al., 2013b; Riemer et al., 2019; Schulz et al., 2006; Stier et al., 2006). Furthermore, as these studies have often been targeted at regions of historical emissions (Richardson et al., 2019a), or made use of climatological distributions (Fiedler and Putrasahan, 2021; Stjern et al., 2017), it is still difficult to determine the distinct physical mechanisms by which absorbing aerosol in different locations can impact global-mean ERF.

To tackle this, here we present a large ensemble of simulations with a state-of-the-art, atmosphere-only global climate model forced by idealized AOD perturbations (with fixed magnitude and radiative properties) placed systematically across the globe. This approach allows us to explore the sensitivity of ERF to localized absorbing aerosol, without being biased to specific regions and without the added uncertainties of aerosol transport and deposition. Our results demonstrate a strong unambiguous dependence of global-mean ERF on the spatial pattern of absorbing aerosol, with positive ERF in response to midlatitude absorbing aerosol and a ‘hot spot’ of negative ERF for absorbing aerosol in the tropical Western Pacific. We show that these varying ERF responses can be understood by considering the dynamical response of atmospheric circulation and clouds to aerosol heating in different locations and propose a constraint on the ERF of midlatitude absorbing aerosol due to vorticity conservation. Finally, we highlight the importance of our results for large transient ERF from Indonesian peatland wildfires.

2.2 Methods

2.2.1 absorbing aerosol experiments

An ensemble of 145 fixed-SST simulations were conducted using the ICON (ICOsa-hedral Non-hydrostatic) general circulation model which has a newly developed dynamical core and has been tuned to give a well-balanced top-of-atmosphere radiation budget and reasonable representation of clouds and precipitation compared with observations (Crueger et al., 2018b; Giorgetta et al., 2018). This configuration of the ICON model uses the ECHAM6 physics package (Stevens et al., 2013b), including a bulk mass-flux convection scheme (Nordeng, 1994; Tiedtke, 1989a) and cloud cover calculated locally in each grid box using a relative humidity scheme (Sundqvist et al., 1989a). Cloud microphysics is based on the Lohmann and Roeckner scheme (Lohmann and Roeckner, 1996b); however, the droplet number is fixed to prescribed profiles over land and ocean (Stevens et al., 2013a) and the model does not simulate aerosol–cloud interactions. Radiation is parameterized using the PSrad radiative transfer scheme (Pincus and Stevens, 2013a), which implements the gas radiative transfer and solvers from the widely used Rapid Radiation Transfer Model (Mlawer et al., 1997). We have extended the model so that the total aerosol radiative effect is calculated at every radiation time-step by performing a double-call to the radiation scheme with and without the aerosol perturbation and then subtracting the results. The IRF is then calculated by differencing the total aerosol radiative effect between a perturbed run and the control.

The simulations were run on a triangular grid (R_2B_4 specification) with a 223 km edge length, corresponding to an average grid spacing of approximately 157.8 km on a Cartesian grid. The model uses a terrain-following vertical sigma-height grid (Leuenberger et al., 2010) with 47 levels between the surface and the model top at 83 km. Each simulation, including the control run, is integrated for 20 years with timesteps of 15 min and 90 min for the dynamics and radiation, respectively, after which the first year is discarded as spin-up and analysis is conducted over

2. Sensitivity of effective radiative forcing to the spatial pattern of absorbing aerosols

the remaining 19 years. The simulation length was chosen as a trade-off between computational constraints and the requirement that our runs are long enough to give robust results in the face of internal variability.

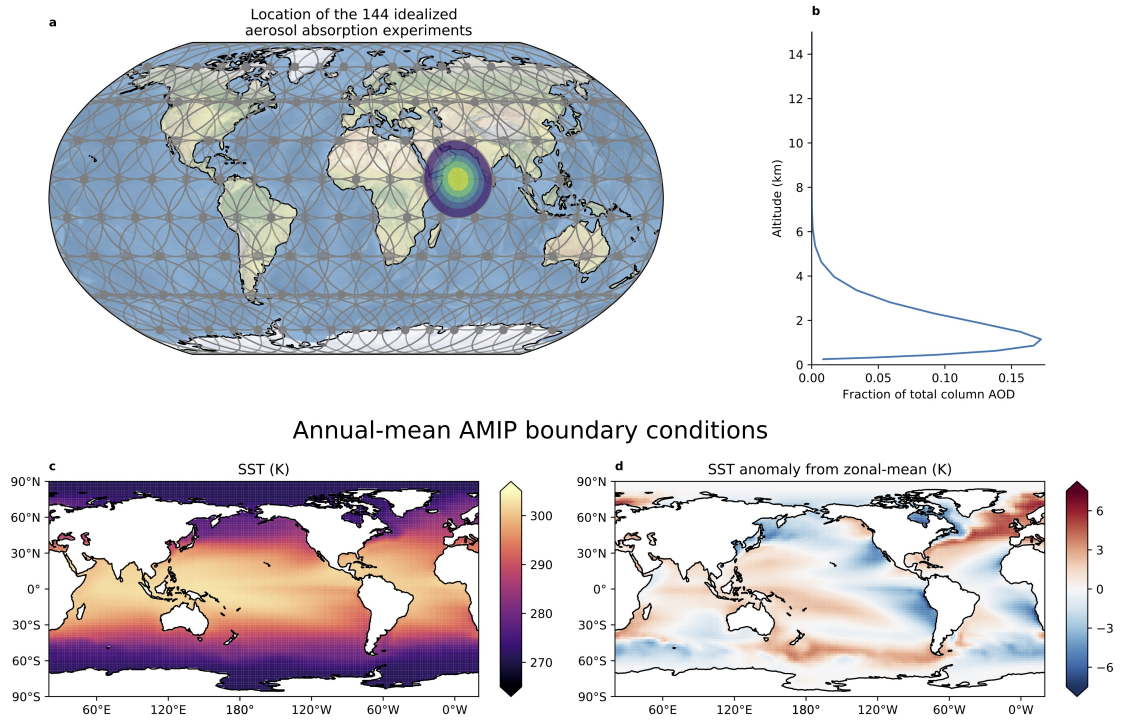


Figure 2.1: **a** Geographic illustration of the 144 idealized aerosol plumes used. Grey ellipses illustrate the extent of the plume in each experiment. Each plume corresponds to a separate experiment and a filled contour illustrating the horizontal structure of the plume AOD is shown for the plume centered over the Indian Ocean. **b** Vertical profile of fractional aerosol optical depth at 533 nm with respect to the column integral. The vertical axis denotes the approximate altitude above the local surface. The vertical profile is constant for all the perturbation experiments we conduct. **c, d** Annual-mean SST field used to drive the simulations, along with the corresponding SST anomaly relative to the zonal-mean temperature.

As a further attempt to reduce variability due to trends in atmospheric composition, each simulation is run with greenhouse gases and ozone fixed at their 1979 levels. Additionally, the simulations are also forced by climatological monthly SSTs and sea-ice concentrations derived from the Atmospheric Model Intercomparison Project (AMIP) (Gates, 1992) boundary conditions over 1979–2016, eliminating SST-driven variability such as the ENSO, which could otherwise bias the results. Figure 2.1c,d illustrates the annual-mean SST field used to force the model, demonstrating neutral

ENSO conditions with a warm pool in the Western Pacific and a cold pool in the Eastern Pacific.

Aside from the control run, the only difference between the remaining 144 simulations is the inclusion of an idealized plume of absorbing aerosol at 144 different geographical locations (Figure 2.1a), parameterized using the Max Planck Institute Aerosol Climatology version 2, Simple Plume (MACv2-SP) model (Stevens et al., 2017). The plume’s horizontal structure is specified to be a 2-dimensional Gaussian function with a standard deviation of 10° . The vertical distribution of aerosol is based on the kernel of Euler’s β -function, with most of the aerosol concentrated in the lower 3 km of the atmosphere and a peak at 1 km (Figure 2.1b). We note that differences in the vertical plume profile may cause different responses of the climate system and general circulation (Ban-Weiss et al., 2012; Persad et al., 2012). However, a vertical distribution similar to ours has been found to be broadly representative of the vertical structure of fine-mode AOD (Kinne et al., 2013). Furthermore, we will show later in sensitivity tests for the Indonesian wildfire perturbation that our results are robust to realistic changes in the vertical aerosol profile. Both the horizontal and vertical structure of the plume are constant in time and the shape of the plume is not influenced by the local meteorology.

In each of the aerosol perturbation experiments, the plume is prescribed with a total column AOD=1 and a single-scattering albedo (SSA) of 0.8, both defined at a wavelength of 550 nm. This corresponds to a strongly absorbing plume of aerosol, generating a large diabatic heating perturbation in the lower atmosphere, even while generating cooling at the surface. Furthermore, to isolate the role of the radiative impact of the plume, we do not include the microphysical effects of the aerosol plume, which are also described in Stevens et al. (2017), so the radiative forcing we calculate is the effective radiative forcing from aerosol-radiation interactions (ERF_{ARI}) in the framework of the IPCC Sixth Assessment Report (Forster et al.,

2021).

2.2.2 Calculation of ERF and adjustments

We calculate the ERF as the difference between the net top-of-atmosphere radiative fluxes between a perturbation experiment and the control (after discarding the first year as spin-up and taking averages over the subsequent 19 years), following the suggestion of Forster et al. (2016). All experiments are conducted with fixed SSTs and sea-ice concentrations.

We then use the IRF, defined as the difference between aerosol radiative effect in the perturbed and control experiments, to calculate the contributions from fast adjustments in clear and cloudy sky as:

$$\text{Adj}_{\text{clear}} = \text{ERF}_{\text{clear}} - \text{IRF}_{\text{clear}} \quad (2.1)$$

and

$$\text{Adj}_{\text{cloudy}} = \text{ERF}_{\text{cloudy}} - \text{IRF}_{\text{cloudy}} \quad (2.2)$$

where $\text{ERF}_{\text{clear}}$ and $\text{ERF}_{\text{cloudy}}$ are defined the same way as the ERF but only using top-of-atmosphere fluxes in clear and cloudy sky, respectively.

2.2.3 Aquaplanet experiments

We run the ICON aquaplanet simulations for 10 years, with the first year discarded as spin-up and analysis conducted on time-averages of the final 9 years. The SST is prescribed in each experiment, with the zonal-mean SST taken as the ‘control’ distribution from Neale and Hoskins (2000a), in addition to a wavenumber-2 SST dipole which spans one half-hemisphere and takes the form,

$$\delta\text{SST} = 3\cos(2\lambda)\cos\left(\frac{\pi}{2}\frac{\phi}{\phi_0}\right), \quad (2.3)$$

for $-\frac{\pi}{4} < \lambda < \frac{3\pi}{4}$, where λ is the longitude, ϕ is the latitude and $\phi_0 = \frac{\pi}{6} = 30^\circ$ is the latitudinal width of the perturbation. The amplitude of the dipole is thus $\pm 3K$. The absorbing aerosol perturbations are prescribed at regular longitude intervals spanning between the peaks of the SST dipole (Fig. 2.8b) and the aerosol properties are identical to those used in the main experiments.

2.2.4 Stream-function calculation

To calculate the zonal-mass stream function for tropical overturning, we first use the ‘windspharm’ library (Dawson, 2016) to calculate the divergent component of the zonal wind u_{div} using a Helmholtz decomposition (Schwendike et al., 2014), and then calculate the zonal-mass stream function at each pressure p and latitude ϕ using:

$$\psi(\phi, p) = \frac{2\pi R_e}{g} \int_{1000\text{hPa}}^p u_{\text{div}}(p', \phi) dp'. \quad (2.4)$$

Here, R_e is the earth’s radius. To plot longitude-pressure cross-sections, we then average ψ between $\pm 20^\circ$ latitude of the equator using area-weighting.

2.2.5 Indonesian plume experiments

We conduct experiments with the realistic climatology of Indonesian biomass-burning from Stevens et al. (2017), which has a simple seasonal cycle peaking in the dry season (Fig. 2.12) and a prescribed single-scattering albedo of 0.87. We then run a series of five 10 year experiments, with the AOD in each experiment scaled by an extra factor of 2 compared with the climatology. The vertical profile of the aerosol is the same in each experiment and only the magnitude is varied. The ‘Climatology_x16’ experiment has a seasonal cycle of AOD which compares favourably with the 2015/2016 AERONET observations and the results of Kiely et al. (2019).

2.3 Results

2.3.1 Dependence of ERF on the location of absorbing aerosol

The global-mean ERF in response to the 144 separate aerosol plume experiments is illustrated in Figure 2.2a. The AOD perturbations are prescribed using two-dimensional Gaussian distributions in the horizontal, with an amplitude of 1 at their centre (Methods). The global-mean AOD perturbation is the same in each experiment and only the location is varied (Fig. 2.1). The single-scattering albedo at 550 nm is set to 0.8 in each experiment, corresponding to strongly absorbing aerosol (Brown et al., 2021). The variations in ERF in Figure 2.2a illustrate that the ERF from a certain global-mean absorbing aerosol perturbation depends strongly on how the absorbing aerosol is spatially distributed. For example, when the absorbing aerosol is concentrated in the extratropics, the ERF is generally positive, whereas when the absorbing aerosol is located over the tropical Western Pacific, there is a strong negative ERF in the global-mean. The two insets below Figure 2.2a show that the spatial pattern of the ERF also varies markedly depending on the plume location, with an extratropical plume generating a positive local increase in ERF, whereas the plume over the tropical Western Pacific generates a global-scale ERF perturbation, with large regions of positive and negative response.

Next, we decompose the ERF into contributions from IRF and rapid adjustments (Adj) and analyse these in clear and cloudy sky. This allows us to disentangle the direct, radiative effect of the absorbing aerosol (the IRF) from its influence on circulation and cloudiness (the adjustments). For most of the plumes, the contribution from IRF is a balance between the changes in clear and cloudy sky, resulting in an overall small contribution from IRF in most cases (Figure 2.2d,e). The cloudy-sky IRF is generally positive because low clouds serve as a high-albedo surface, whereas the clear-sky IRF is negative for most perturbation sites except for over high-albedo surfaces such as the Sahara or the high latitudes.

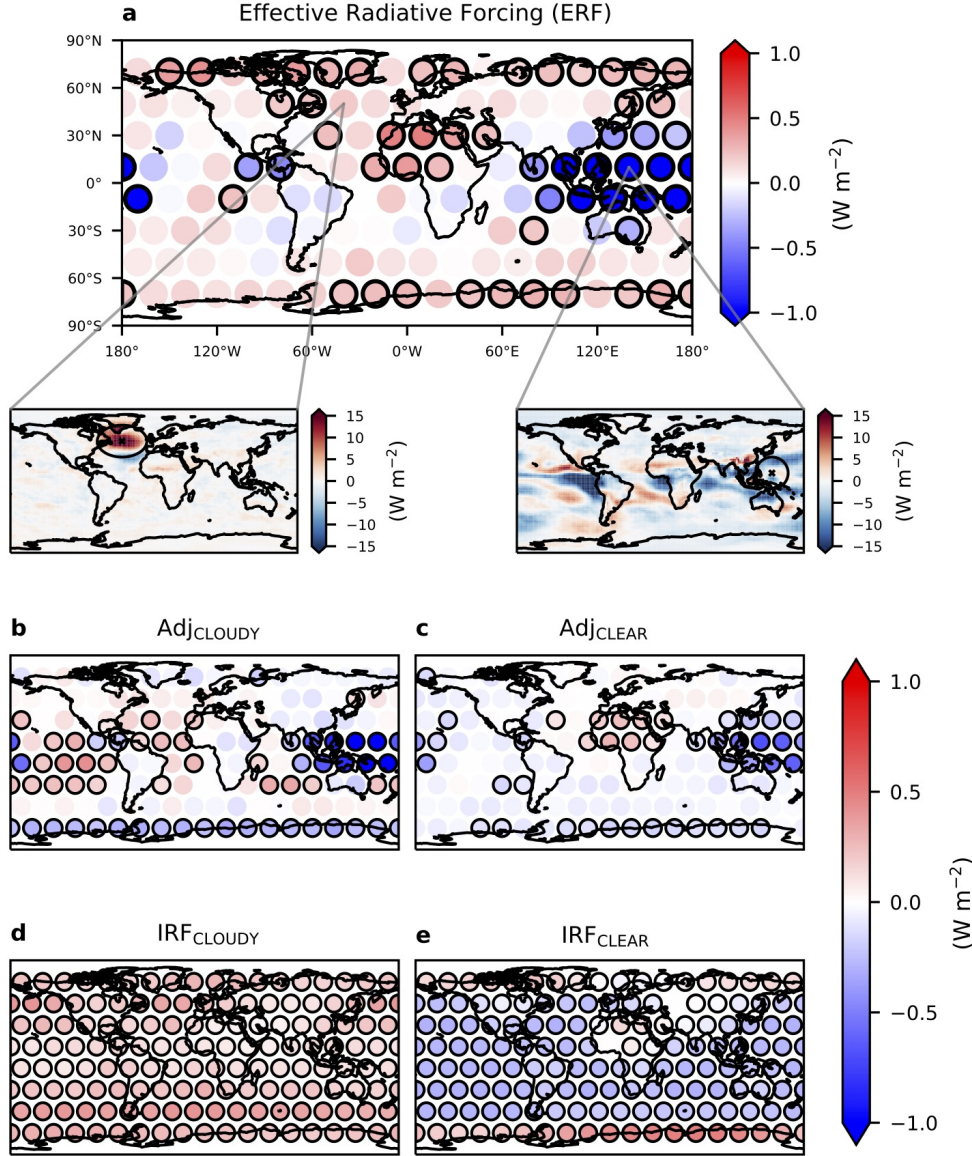


Figure 2.2: **a**, Global- and time-mean effective radiative forcing (ERF) calculated as the net top-of-atmosphere flux difference between simulations with the absorbing aerosol perturbation and the control experiment. Each filled circle denotes the global-mean results from the experiment with the idealized absorbing aerosol plume centred on that location. Enclosing black circles are present where the time-series of global- and annual-mean forcing is statistically significant from zero with $P < 0.05$ according to a one-sided t-test. Two insets at the bottom show the spatial pattern of ERF changes for two illustrative plume experiments. In each of these subplots, the centre of the absorbing aerosol perturbation in that experiment is marked with an x and its size illustrated by an ellipse which contains 95% of the total AOD perturbation in that experiment. Ellipses are used in the insets to represent the distortion introduced by the map projection. **b–e**, Decomposition of the total ERF into contributions from Adj (**b,c**) and IRF (**d,e**) in clear and cloudy sky.

2. Sensitivity of effective radiative forcing to the spatial pattern of absorbing aerosol

Variations in global-mean ERF in response to absorbing aerosol are mainly caused by variations in the global-mean rapid adjustments, as shown in Figure 2.2b,c. The rapid adjustments in clear sky are well-explained by changes in clear-sky outgoing longwave radiation arising from global-mean temperature changes (Fig. 2.3), as land temperatures are free to evolve in our simulations and only the SSTs are fixed.

To understand the variations in clear-sky rapid adjustments, we note that, globally, the outgoing longwave radiation (OLR) can be written as:

$$\text{OLR} = \epsilon \sigma T^4, \quad (2.5)$$

where σT^4 is the surface blackbody emission, and ϵ is an ‘effective emissivity’ defined such that equality holds at a given surface temperature. In our control experiment, with $\text{OLR} \approx 240 \text{ W m}^{-2}$ and $T \approx 280 \text{ K}$, we find $\epsilon \approx 0.8$. Given a change in global-mean surface temperature, one can then relate the clear-sky adjustments to the first-order Taylor expansion about the control state as:

$$\text{Adj}_{\text{clear}} = -\frac{d\text{OLR}}{dT} \delta T \approx -4\epsilon \sigma T^3 \delta T. \quad (2.6)$$

This scaling is found to reproduce most of the variance in clear-sky rapid adjustments in our experiments (Figure 2.3).

The clear-sky adjustments contribute negatively to the total ERF for most locations except the Sahara and the tropical Western Pacific, where they amplify the cloudy-sky contribution. Overall, the cloudy-sky rapid adjustments (Figure 2.2b) account for over 63% of the total variation in ERF in response to absorbing aerosol in different locations and are the key determinant of the ‘hot spot’ of ERF in the tropical Western Pacific. Hence, we mainly focus our attention on the cloudy-sky adjustments.

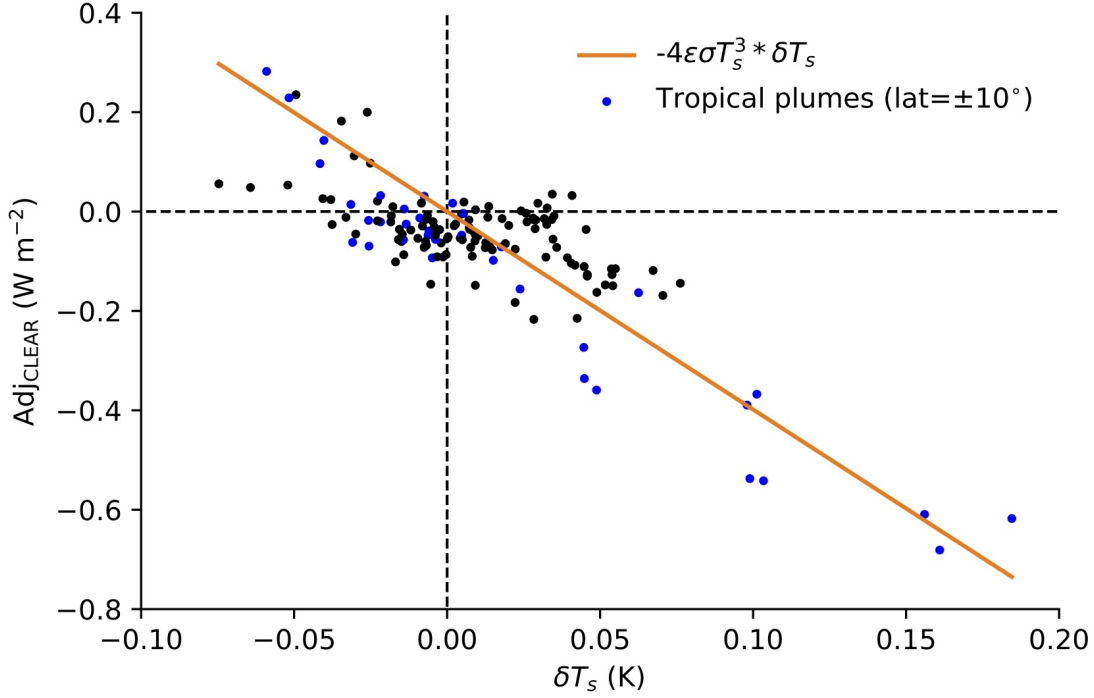


Figure 2.3: Scatter plot of global-mean changes in clear-sky rapid adjustments versus global-mean surface temperature changes for each of the 144 plume experiments. Tropical plumes (i.e., those with their central latitude at $\pm 10^\circ$) are coloured in blue and a parameter-free prediction of the slope (given by Equation 2.6) is drawn in orange.

2.3.2 Circulation changes control the location-dependence of ERF

Cloudy-sky adjustments are strongly anti-correlated with global-mean liquid-water path (LWP) changes across all our experiments (Figure 2.4) because higher LWP is associated with optically thicker clouds which reflect more incoming solar radiation.

But why does absorbing aerosol in some locations cause such an increase in LWP while others cause a decrease? To see why, consider that shortwave absorption from absorbing aerosol introduces a positive temperature tendency to the local atmospheric column, and in the absence of terms that balance this tendency, the temperature in this region would increase without bound. However, in the extratropics (Figure 2.5a), temperature anomalies tend to generate a downstream low-pressure system due to vorticity conservation (Hoskins and Karoly, 1981; Vallis, 2017) which draws down cold air from higher latitudes (Figure 2.6). This cold-air

2. Sensitivity of effective radiative forcing to the spatial pattern of absorbing aerosol

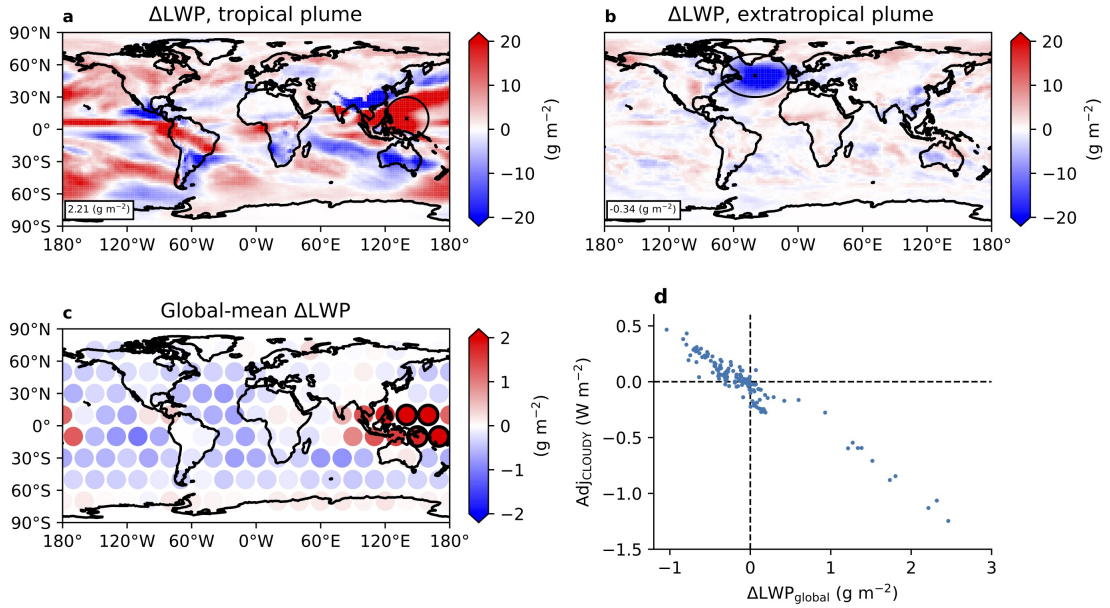


Figure 2.4: **a** Time-mean liquid water path (LWP) changes for the simulation with the idealized aerosol plume located over the Western tropical Pacific and **b** the idealized aerosol plume located over the North Atlantic. The text box shows the value of the global-mean LWP change in each case. **c** Global-mean changes in LWP resulting from idealized aerosol plumes in different locations. **d** shows a scatter plot of the global-mean cloudy-sky adjustments versus the global-mean LWP changes in each of the experiments, demonstrating the strong control of $\text{Adj}_{\text{cloudy}}$ by LWP.

advection in response to absorbing aerosol in the extratropics is a negative feedback that balances the temperature tendency from the absorbing aerosol, generating a finite localized temperature increase.

Other local processes such as radiative cooling or precipitation feedbacks also contribute to balancing the temperature tendency from the absorbing aerosol (Fig. 2.10). This temperature increase causes a decrease in relative humidity, which inhibits cloud formation and decreases LWP in the region of the absorbing aerosol perturbation (Fig. 2.5c). This mechanism explains the broad pattern of positive cloudy-sky adjustments (Fig. 2.2b) and negative LWP changes (Fig. 2.4) throughout the extratropics.

A similar localized decrease in LWP occurs when the absorbing aerosol is in a region of tropical descent (Figure 2.5d–f); however, the mechanism is quite

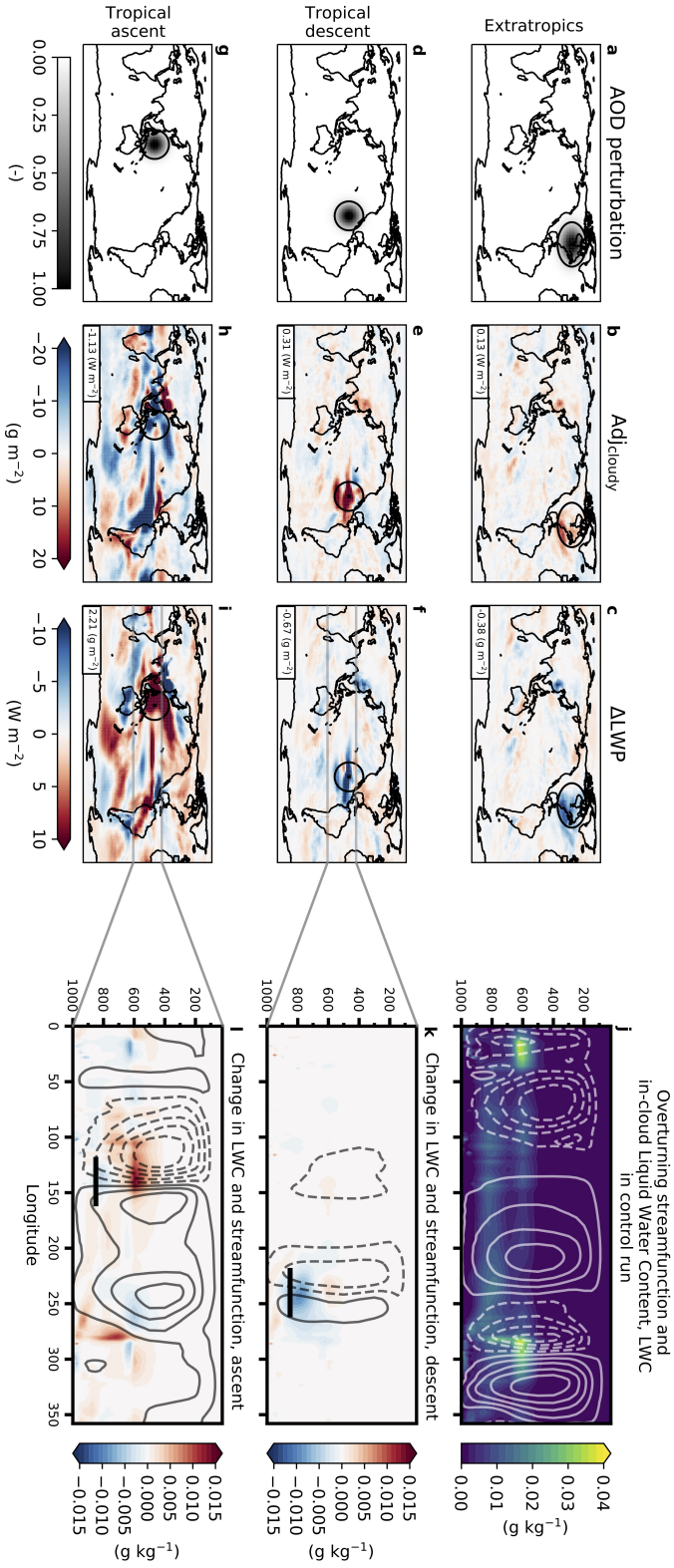


Figure 2.5: a-i, Time-mean AOD perturbation, cloudy-sky adjustments (Adjcloudy) and in-cloud liquid-water path (LWP) changes for idealized absorbing aerosol perturbations in the extratropics (a-c), a tropical descent region (d-f) and a tropical ascent region (g-i). Black ellipses illustrate the horizontal size of the perturbation in each experiment, and the global-mean values for b, c, e, f, h and i are shown in boxes. j, A pressure-longitude cross-section of the in-cloud liquid-water content (colours) and global-mean values for b, c, e, f, h and i are shown in response to absorbing aerosol perturbations in the East and West Pacific, respectively. The contour interval in j-l is $4 \times 10^{-10} \text{ kg s}^{-1}$ and solid contours indicate clockwise overturning, whereas dashed contours indicate counter-clockwise overturning. In k and l, horizontal black lines illustrate the horizontal extent of the AOD perturbation, at a height corresponding roughly to the maximum of the plume AOD profile.

2. Sensitivity of effective radiative forcing to the spatial pattern of absorbing aerosol

distinct. In the tropics, as the Coriolis force is weak, the diabatic heating from absorbing aerosol is balanced mostly by vertical motion, generating a thermally-direct overturning circulation (Bjerknes, 1969; Sobel et al., 2001; Stevens et al., 2017). However, when the absorbing aerosol perturbation is in a region of large-scale descent — in this case, the descending branch of the Pacific Walker cell (Figure 2.5j) — the anomalous overturning from the diabatic heating interacts destructively with the existing large-scale descent and becomes confined to local scales, causing a decrease in low-level liquid-water content (Figure 2.5k).

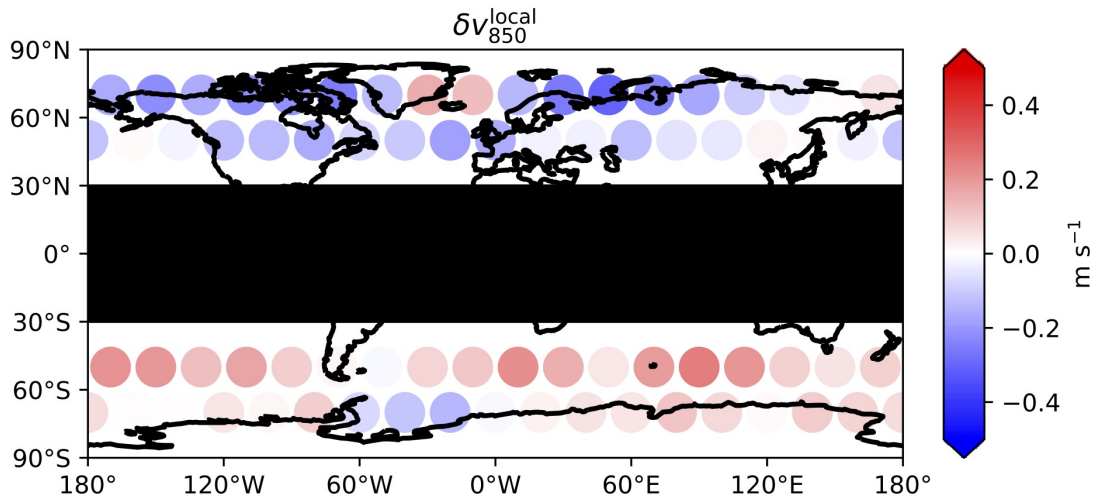


Figure 2.6: Local- and time-mean changes in the meridional velocity at 850hPa for each of the experiments where the absorbing aerosol perturbation was in the midlatitudes. Blue colours indicate anomalous southwards flow (equatorward in the NH) in response to aerosols, and red colours indicate anomalous northwards flow (equatorward in the SH) in response to the aerosol.

Conversely, when the absorbing aerosol perturbation is placed in a region of large-scale tropical ascent (Figure 2.5g), the anomalous vertical motion generated by the diabatic heating projects onto this existing mode of variability, amplifying the Walker circulation and increasing the upper-level divergence. This pattern of low-level convergence and upper-level divergence excites equatorial waves which generate large positive liquid-water content anomalies throughout the tropics (Figure 2.5i,l), and planetary-scale Rossby waves which communicate the perturbation to the

midlatitudes (Hoskins and Karoly, 1981) (Figure 2.5i). A schematic representation of our results is given in Figure 2.7.

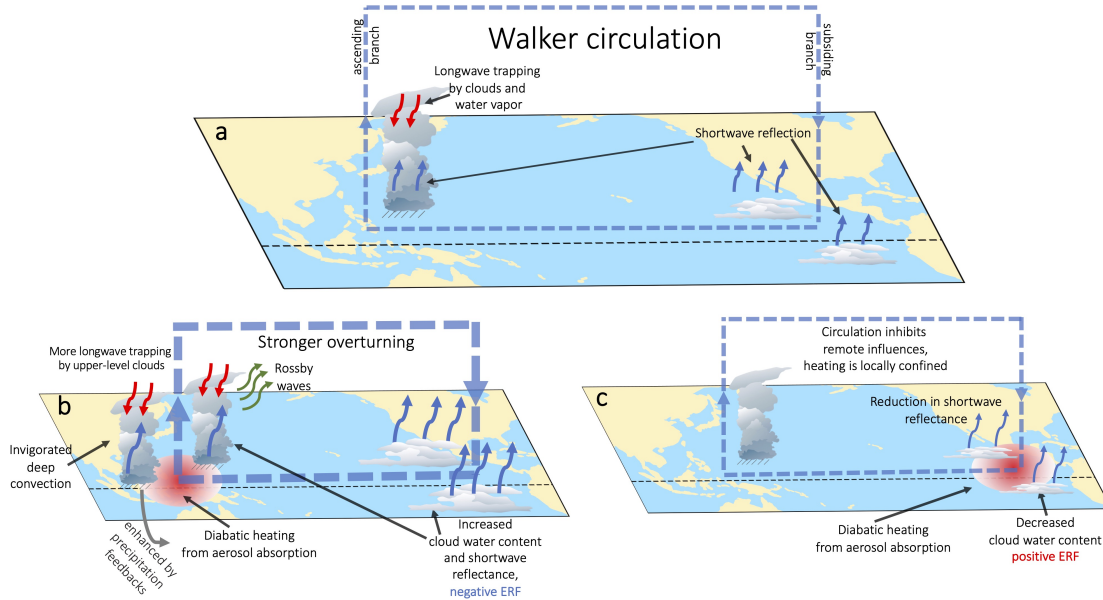


Figure 2.7: **a**, The tropical overturning circulation in the control simulation. **b**, Illustration of how absorbing aerosol in the ascending branch of the overturning circulation leads to a strong increase in shortwave reflectance throughout the tropics. **c**, When absorbing aerosol is placed in the subsiding branch of an overturning cell, the subsidence confines the heating anomaly to local scales, inhibiting cloud formation and causing a local decrease in shortwave reflectance.

To distill the key physical mechanisms, we conducted another set of absorbing aerosol perturbation experiments in a simplified aquaplanet model forced by an SST dipole along the equator (a ‘mock-Walker’ circulation; Methods). This generates an overturning circulation with similar characteristics to the Walker circulation in our control experiment (Figure 2.8). Again, we find that when the absorbing aerosol is located over the warm pool region, there is a strengthening of the circulation and a large increase in LWP which drives a shortwave cloudy-sky adjustment, although this is somewhat offset by increases in upper-level ice, which contributes negatively to the cloudy-sky adjustments. Similar increases in upper-level cloud ice are found in our main experiments, albeit to a lesser extent (Figure 2.9).

2. Sensitivity of effective radiative forcing to the spatial pattern of absorbing aerosols

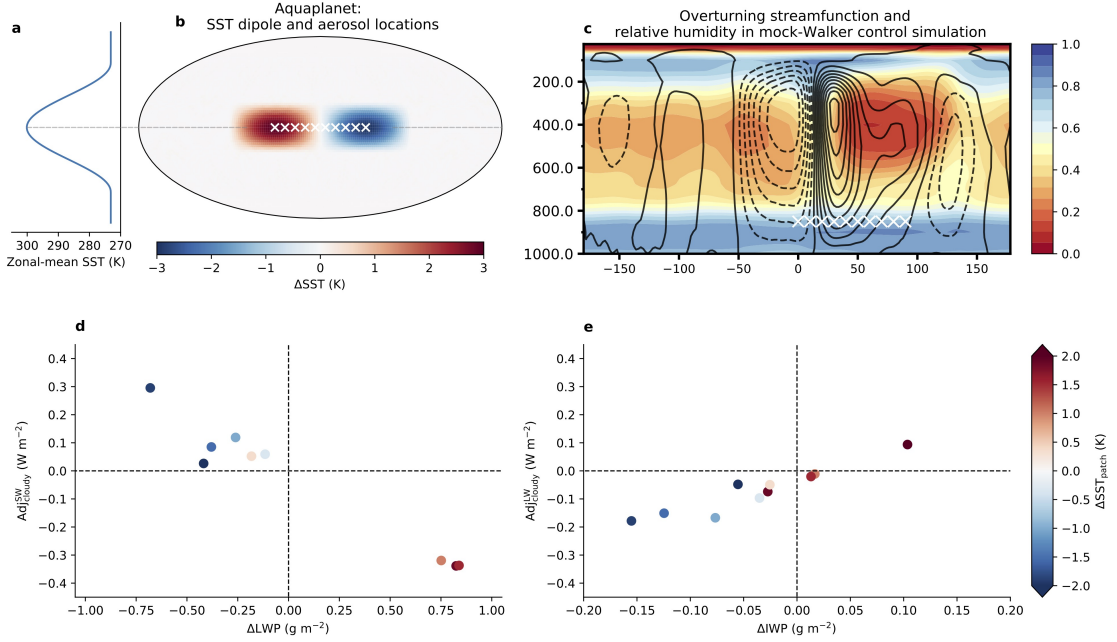


Figure 2.8: **a** The zonal-mean SST profile prescribed in the model. **b** The zonally-anomalous SST dipole imposed along the equator, with white markers illustrating the location of the 10 absorbing aerosol experiments conducted. **c** Pressure-longitude plot of the relative humidity (colours) and zonal mass streamfunction (black contours) averaged between $\pm 20^\circ$ of the equator with area-weighting in the control simulation. The white markers show the approximate longitude and pressure of the AOD maxima in the 10 perturbation experiments. **d** The shortwave component of the global-mean cloudy-sky adjustments in each of the 10 experiments scattered against the global-mean LWP change. **e** The longwave component of the global-mean cloudy-sky adjustments in each of the 10 experiments scattered against the global-mean IWP change. In panels **d** and **e** the colour of the markers illustrates the SST anomaly (from **b**) at the center of the plume in each experiment.

Over land, there is greater convective inhibition due to lower boundary-layer relative humidity (Chen et al., 2020), which inhibits the upward motion associated with diabatic heating in the tropics. This helps explain why similarly large ERFs are not seen in response to absorbing aerosol over South America and the African monsoon regions (which are also associated with large-scale ascent). Additionally, absorbing aerosol causes surface cooling in our experiments when located over land (Dagan et al., 2021). This drives reductions in the surface sensible heat flux and in the diabatic heating perturbation for absorbing aerosol over land (Figure 2.10c), explaining the land–ocean contrast in Figure 2.2b. Adjustments of clouds and precipitation can also impact the diabatic heating perturbation through changes in

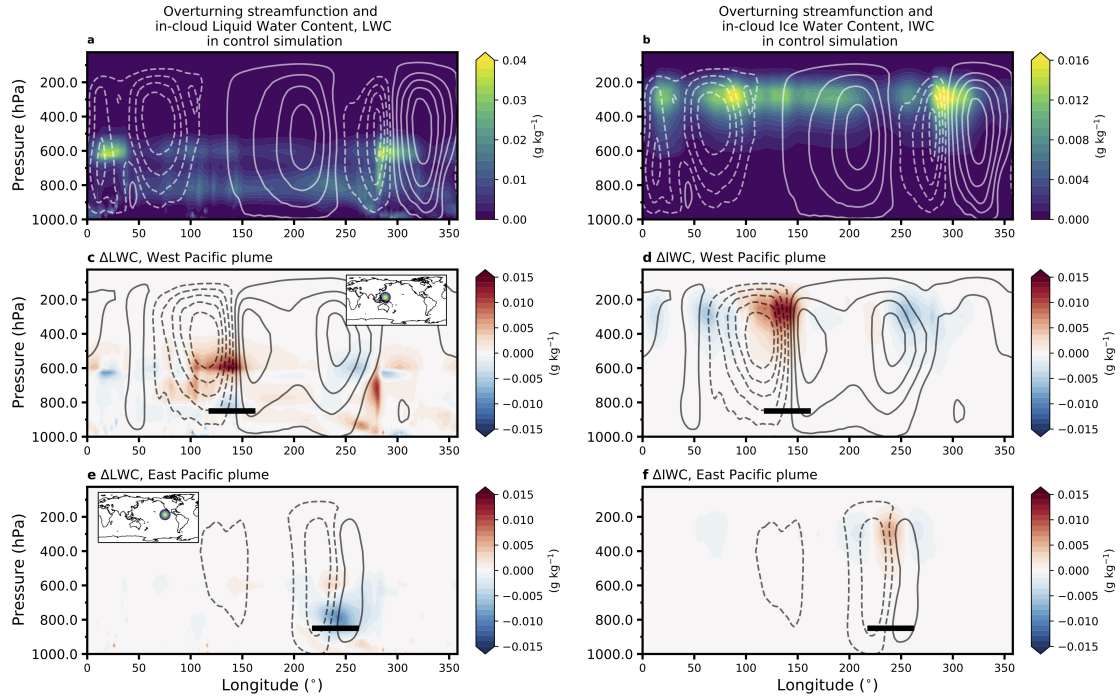


Figure 2.9: Cloud liquid and ice water content in the control simulations (a, b) and changes in response to plumes in the East and West Pacific. Contours indicate the zonal mass stream-function in the control simulation (a, b) and changes in the stream-function are shown in (c, d, e, f). Solid contours indicate clockwise motion. In the lower two rows, horizontal black lines illustrate the 2-sigma (40°) horizontal extent of the plume, at a height corresponding roughly to the maximum of the plume AOD profile.

latent heat release or by reductions in outgoing longwave radiation (Figure 2.10b,f).

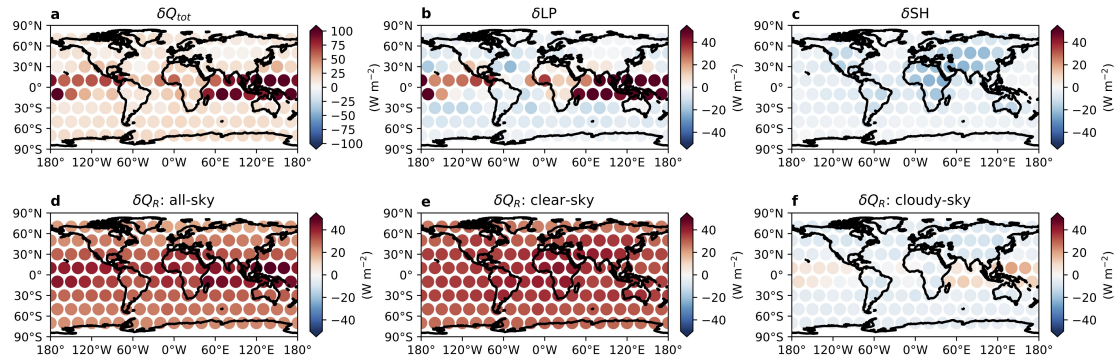


Figure 2.10: a The local diabatic heating anomaly in each of the experiments, and contributions from b precipitation feedbacks, c sensible heat flux changes, d atmospheric radiative cooling. Panels e and f show the decomposition of atmospheric radiative cooling into contributions from clear and cloudy sky, respectively.

To assess how well this explanation captures the simulated variations in cloudy-

2. Sensitivity of effective radiative forcing to the spatial pattern of absorbing aerosols

sky adjustments (and thus, ERF) across the tropics, we quantify it with a simple scaling,

$$\text{Adj}_{\text{cld}} \sim \delta Q_{\text{tot}} \cdot (\nabla \cdot \mathbf{u})_{850-400}^{\text{control}}, \quad (2.7)$$

where δQ_{tot} is the total diabatic heating from the plume (including latent heating) and $(\nabla \cdot \mathbf{u})_{850-400}^{\text{control}}$ is the difference in the horizontal divergence between 850hPa and 400hPa in the control run. Both quantities are averaged within a fixed radius of the centre of the AOD perturbation in that experiment. Similar results are obtained if we replace $(\nabla \cdot \mathbf{u})_{850-400}^{\text{control}}$ with the local vertical velocity at 850hPa.

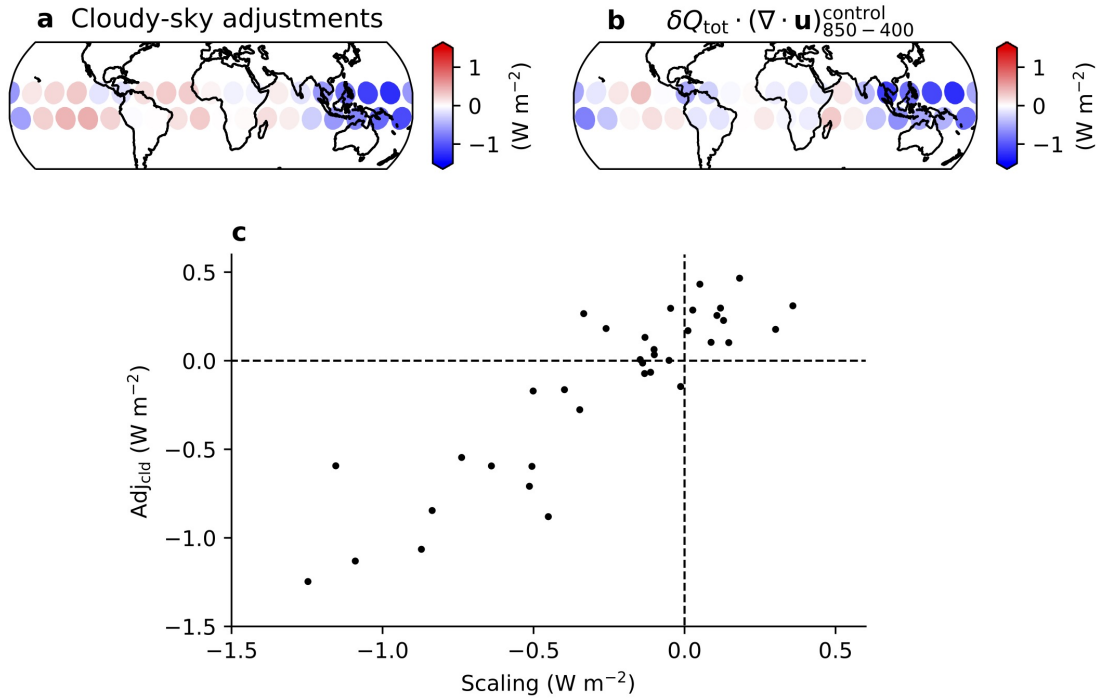


Figure 2.11: **a,b**, Map of the global-mean cloud-sky rapid adjustments from tropical absorbing aerosol experiments (**a**) and the corresponding estimate when using a simple circulation-based scaling that relates the total diabatic heating from the absorbing aerosols and the local vertical shear of the horizontal divergence in the control run to the cloud-sky adjustments (**b**). **c**, Scatterplot of the results in **a** and **b**. In **b** and **c**, the scaling was evaluated in S.I. units ($\text{W m}^{-2} \text{ s}^{-1}$) and then, for plotting purposes, each scaling estimate was multiplied by the same dimensional constant such that the results match the magnitude of the cloud-sky adjustments in **a** (this does not affect the correlation).

Our scaling captures the variation in simulated cloud-sky adjustments in

response to absorbing aerosol perturbations across the tropics (Figure 2.11) and demonstrates that this can be understood quantitatively in terms of simple tropical dynamics, as illustrated in Figure 2.7. Notably, the scaling also successfully predicts the ‘hot spot’ of negative ERF over the tropical Western Pacific.

2.3.3 Implications for Indonesian biomass-burning events

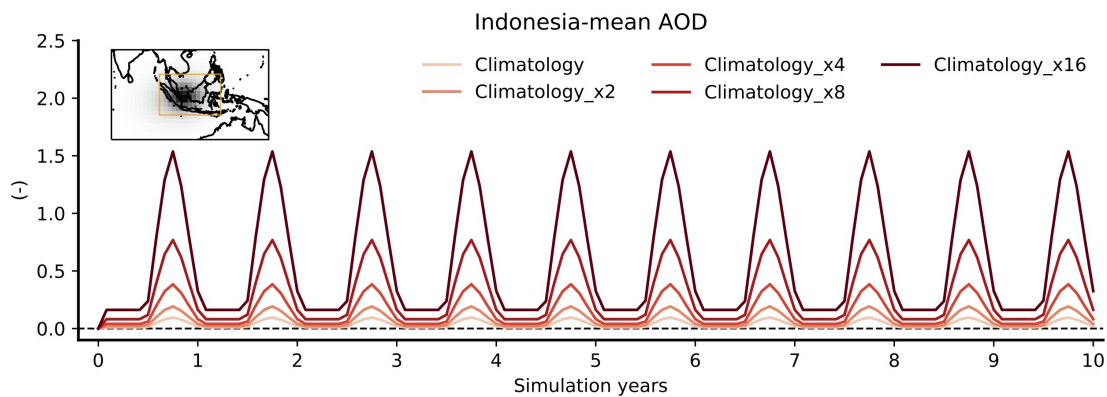


Figure 2.12: Time-series of Indonesia-mean AOD in our experiments with the realistic climatology of Indonesian biomass-burning, with a seasonal cycle which peaks in the dry-season (Methods). The orange box in the inset shows the averaging area and encompasses the same area as the inset in Fig. 2.13 of the main article. For each experiment we double the AOD relative to the climatological distribution.

The strong ‘hot spot’ of ERF over the tropical Western Pacific region raises the question: could this have real world implications? One potential candidate is Indonesian wildfire events. Wildfires have been recorded in this region since the late Pleistocene (Anshari et al., 2001); however, the worst known fires have occurred since the 1980s due to changes in land use and population density (Field et al., 2009). Large peat deposits make Indonesian wildfires uniquely impactful since peat is highly combustibile and difficult to extinguish, and peat fires have repeatedly led to high AOD ‘haze’ events (Kiely et al., 2019). Time series of ground-based observations of mid-visible AOD are plotted in Fig. 2.13a using data from five Aerosol Robotic Network (AERONET) sites located in Indonesia. There is substantial biomass-burning in the dry seasons in most years, particularly in

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2015 (associated with a strong El Niño) and 2019 (without an El Niño, but with a positive Indian Ocean dipole).

To investigate whether these events could generate large ERF as suggested by our idealized experiments, we conduct experiments using a realistic AOD climatology of Indonesian biomass-burning (Methods), with a seasonal cycle that peaks in the dry season (September-October-November, SON) (Figure 2.12). We perform five experiments in which we successively double the climatological Indonesian plume AOD until the peak dry-season AOD is comparable to the 2015/2016 observations (Kiely et al., 2019), corresponding to 16 times the climatological AOD. Despite the use of a more realistic AOD distribution and a seasonal cycle, Fig. 2.13b shows that the spatial pattern and magnitude of the resultant ERF are remarkably similar to our idealized simulations over the ‘hot spot’ (for example, Figure 2.5h).

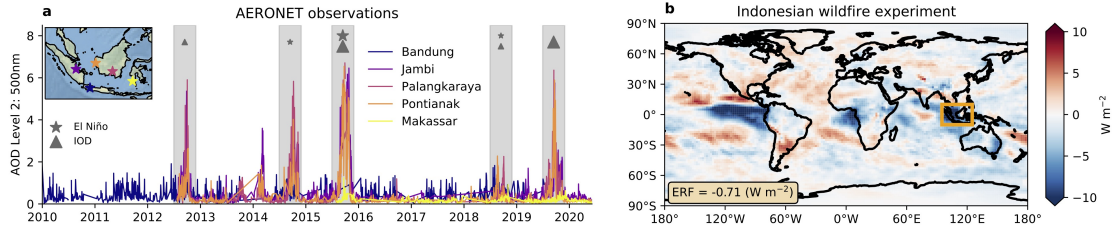


Figure 2.13: **a**, Time series of aerosol optical depth at 500 nm from five AERONET sites in Indonesia. Grey shading denotes dry seasons with substantial biomass-burning. Years with an El Niño event are shown with a star and years with an Indian Ocean Dipole (IOD) event are shown with a triangle, the size of the marker illustrating the magnitude of the event. **b**, Map of the time-mean ERF from our ‘Climatology_x16’ experiment (Methods). The orange box shows the region corresponding to the inset in **a** and illustrates the approximate area of the AOD perturbation in this experiment.

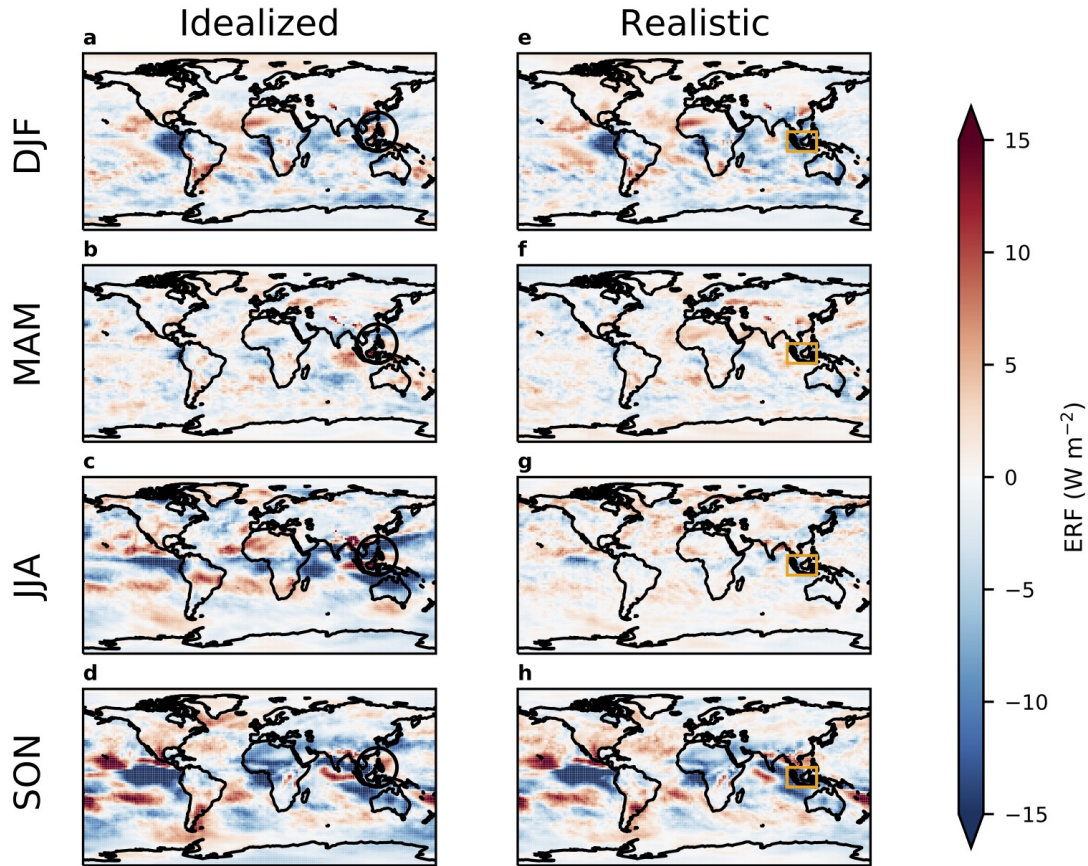


Figure 2.14: Seasonal-mean ERF response for the idealized (a,b,c,d) and realistic (e,f,g,h) absorbing aerosol perturbations over Indonesia.

The magnitude of the ERF response varies seasonally in response to seasonal shifts in tropical circulation, as it also does in the idealized experiments (Fig. 2.14). However, the lack of AOD outside of the dry season (SON) in the realistic experiment means that there is little ERF response in March-April-May (MAM), in contrast to the idealized experiment.

Firstly, to evaluate the extent to which the vertical location of the aerosol varies over this region during the dry season (SON), we examine the vertical profile of CALIOP cloud-free aerosol extinction at mid-visible wavelengths, averaged over the Indonesian region (Figure 2.15). Figure 2.15 demonstrates that the aerosol extinction associated with Indonesian wildfires is confined almost entirely to below the lowest 4km of the atmosphere, which is broadly in line with the results we

show in the main experiments. However, as illustrated by the SON 2015 results (orange line, associated with a strong wildfire season), there is considerable inter-annual variability in the vertical profile, and so we have conducted three additional simulations to test the impact of this variability.

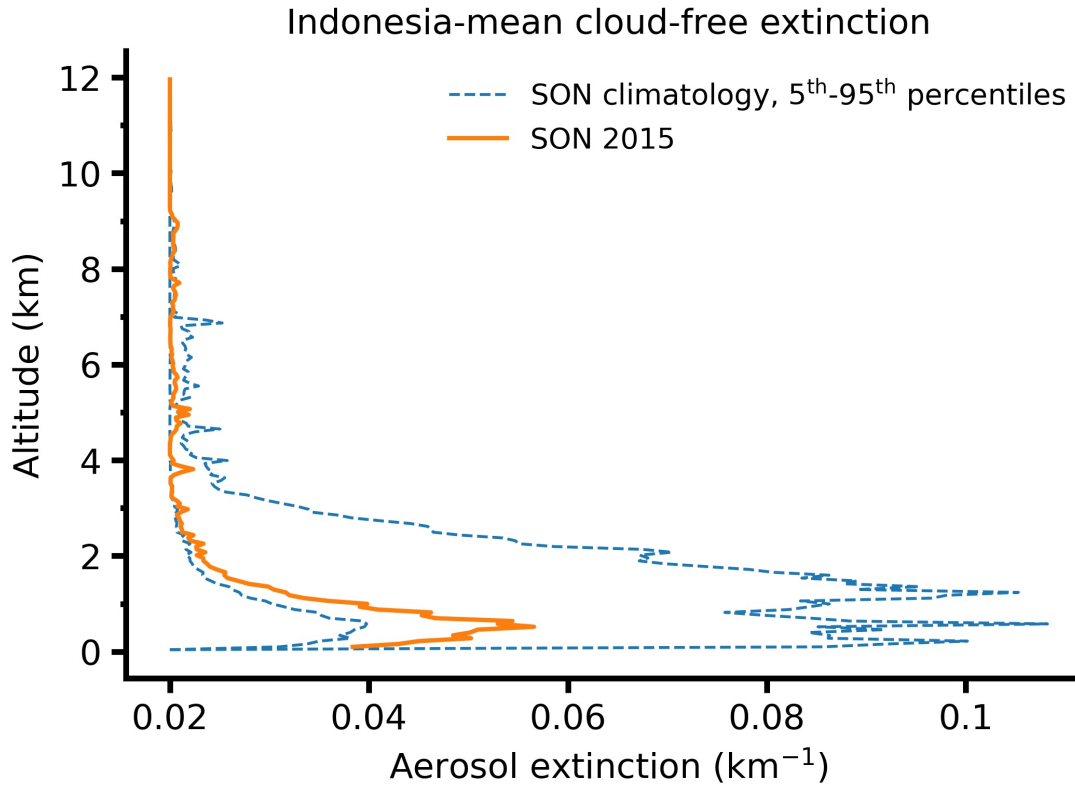


Figure 2.15: Vertical profile of cloud-free, nighttime aerosol extinction at mid-visible wavelengths taken from CALIOP data from 2009-2017. Before plotting we have averaged the gridded data over the Indonesian region and have subsampled the data to the dry season months (SON) before calculating the 5th and 95th percentiles (blue dashed lines).

Figure 2.16 shows the vertical aerosol profile in the baseline realistic Indonesian plume experiment, along with the corresponding change in LWP and ERF (top row). The next three rows show the same results, but for three different vertical aerosol profiles. The additional vertical profiles are taken from the three other ‘biomass-burning’ plumes in the default MACv2-SP model (Stevens et al., 2017). The MACv2-SP data is anchored by surface-based observations with the aid of a full-complexity aerosol-climate model (specifically to determine the vertical profiles).

Overall, Figure 2.16 shows that the magnitude and spatial pattern of the annual-mean ERF response are also robust to realistic variations in the vertical distribution of the total column AOD.

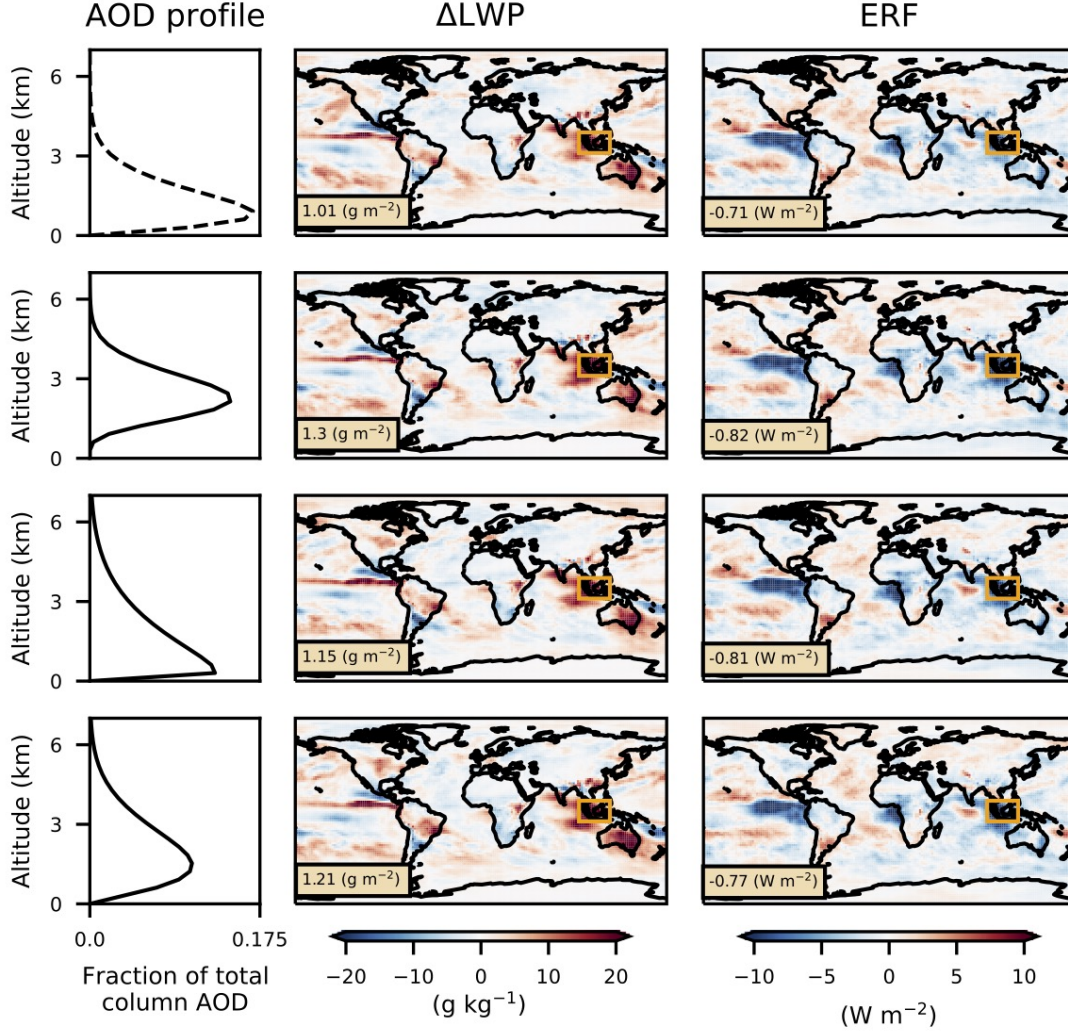


Figure 2.16: Top-row shows the vertical aerosol profile used in the ‘Climatology_x16’ Indonesian wildfire experiment (dashed line, top-left), along with the change in LWP and the ERF in the middle and right columns. Each row below this corresponds to the same ‘Climatology_x16’ experiment, except we have varied the vertical aerosol profile to sample realistic variability in the vertical structure of the perturbation. This figure shows that our main results regarding the Indonesia wildfire experiment are not sensitive to realistic variations in the vertical profile.

We also find that the global-mean ERF scales linearly with the AOD burden in each of the experiments (Figure 2.17). This provides evidence that the reason

similarly large ERF has not been noted in previous studies which use climatological AOD distributions (Stjern et al., 2017) or emissions from a single year (Johnson et al., 2019) is not because we have misunderstood the physics, but rather because these large biomass-burning events are under-represented in climatological distributions and have strong inter-annual variability.

Tropical SST variability such as the El Nino Southern Oscillation (ENSO) also plays a role by modulating the pattern of tropical overturning and generating anomalous subsidence over this region, which would inhibit the remote influence on LWP and ERF we see in our experiments (Fig. 2.7). This helps to explain why previous studies on the ERF of the 1997 Indonesian wildfire (Podgorny et al., 2003) (also associated with a strong El Nino) have found a positive ERF, localized to small scales. Our proposed mechanism predicts such a response (Figure 2.7).

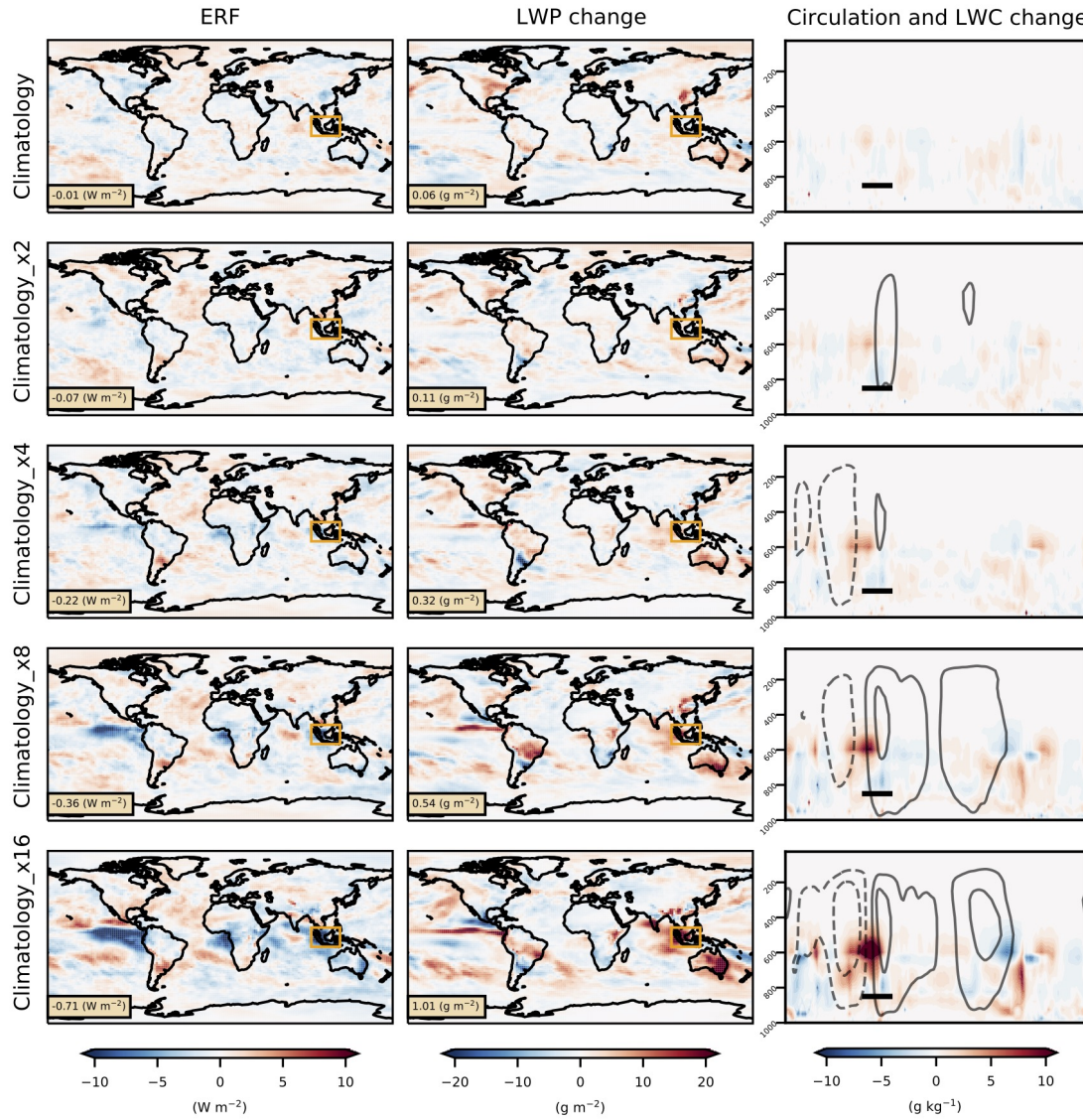


Figure 2.17: Annual-mean Effective Radiative Forcing (ERF), in-cloud Liquid Water Path (LWP), Liquid Water Content (LWC) and zonal-mass stream function changes in response to the five Indonesian biomass burning experiments compared to the control experiment with no AOD. The left column shows the ERF in each of the successive doubling experiments outlined in Fig. 2.12, with the global-mean ERF presented in the corner of each panel. The middle column shows the LWP changes in each of the experiments, illustrating again that the ERF and the LWP are highly anticorrelated. The rightmost panel shows pressure-longitude cross-sections of the in-cloud liquid water content (LWC, colours) and zonal mass streamfunction (contours) in the control simulation, averaged 20° on either side of the equator in each experiment - same plotting conventions as in Fig. 2.5 of the main article. In the right column, horizontal black lines illustrate the horizontal extent of the AOD perturbation, at a height corresponding roughly to the maximum of the plume AOD profile (see Methods).

2.4 Discussion and Conclusions

In summary, our knowledge of how the ERF due to absorbing aerosol depends on its spatial pattern is still uncertain because of uncertainty in the translation of absorbing aerosol emissions to regional absorbing aerosol concentrations, and the subsequent interactions of heterogeneous absorbing aerosol heating on atmospheric circulation and clouds. To address this, we have conducted a series of experiments with a state-of-the-art climate model in which identical absorbing AOD perturbations are placed across the globe. We have shown that the ERF of absorbing aerosol depends strongly on its spatial pattern, with even the sign of the forcing varying markedly between different perturbation sites. In the extratropics, we find that meridional advection of high-latitude air confines the heating anomaly to local scales, driving a positive ERF through reductions in LWP local to the plume. On the other hand, in the tropics the ERF is determined by interactions between the diabatic heating and the structure of the tropical overturning circulation. In regions of large-scale descent, the diabatic heating anomaly is confined to local scales, generating a similar response to the extratropical case. However, in regions of large-scale ascent, the absorbing aerosol excites substantial wave activity which generates positive LWP anomalies across the globe and a strong negative ERF.

Of particular interest is the ‘hot spot’ of negative ERF we identify in response to absorbing aerosol over the tropical Western Pacific, which suggests that there could be large transient ERF from seasonal biomass-burning over this region. We explored this question using a realistic climatology of Indonesian biomass-burning, constrained by AERONET observations during the 2015/2016 wildfire event, and showed that significant global-mean ERF of $\sim \mathcal{O}(-1\text{Wm}^{-2})$ is possible via our proposed mechanism. Our results demonstrate that Indonesian biomass-burning could represent a large source of negative ERF on interannual timescales, particularly during neutral ENSO or La Niña years (Gaveau et al., 2014).

Although non-local circulation adjustments are included in the definition of ERF (Forster et al., 2021; Smith et al., 2020a) and are known to play some role in the ERF of well-mixed greenhouse gases (Forster et al., 2021; Merlis, 2015), their importance for the global-mean aerosol ERF is highly uncertain (Johnson et al., 2019; Richardson et al., 2019a). The inherent uncertainty in linking circulation changes to global-mean ERF is amplified for aerosol due to the additional challenge of understanding how spatially heterogeneous forcings impact circulation (Chemke and Dagan, 2018; Diao et al., 2021; Dong et al., 2014, 2016; Johnson et al., 2019; Kang et al., 2021; Zhou et al., 2020). By taking an idealized approach to this problem, we have shown that non-local circulation adjustments can play a first-order role in determining the ERF of absorbing aerosol, even on global-mean scales. This contrasts with a recent study by Johnson et al. (2019) which showed that suppressing circulation changes with nudging did not impact estimates of global-mean ERF (Johnson et al., 2019). However, this experiment used emissions from a single year (2014) which did not have substantial biomass-burning over the tropical Western Pacific. Overall, this suggests that caution is needed when using nudged experiments to determine the ERF from regional aerosol, particularly for absorbing aerosol. However, quantitative comparisons between our study and that of Johnson et al. (2019) are complicated by the fact that the atmospheric response to regional aerosol can be non-linear (Herbert et al., 2021b), and this should be investigated in future work.

One feature of our approach is the use of fixed SSTs (as is required to calculate ERF), which prevents the ocean from responding to the imposed aerosol forcing. Dynamical ocean feedbacks can play a crucial role in determining the overall climate response to an imposed perturbation (Chemke and Dagan, 2018; Kang et al., 2021; Myhre et al., 2017; Voigt et al., 2017) and sea-surface temperatures can play a role in preconditioning the atmospheric response to aerosol changes (Dong et al., 2019a; Podgorny et al., 2003). For the case of transient aerosol perturbations (such as wildfires on seasonal timescales), one might expect ocean dynamics/thermodynamics

to play a smaller role in the real world than is found in previous studies that used long integrations (Kang et al., 2021; Westervelt et al., 2020). However, this is still uncertain and future work should approach this question using a hierarchy of idealized experiments as illustrated here.

Although the responses we identify are robust across a hierarchy of model configurations and interpretable in terms of robust atmospheric dynamics, we have only used one model and the uncertainties associated with parameterized convection suggest that it would be fruitful to perform similar idealized experiments in other global climate models, or in global storm-resolving models (Stevens et al., 2020). Our work could also be extended by conducting analogous experiments with scattering aerosol. Both would help to build confidence in the physical mechanisms driving this location-dependence and provide a unique community resource for investigating inter-model differences in aerosol ERF, which are particularly large for absorbing aerosol (Stjern et al., 2017; Thornhill et al., 2021), potentially allowing this location-dependence to be parameterized in simpler models.

Finally, our experiments focus on the ERF from aerosol–radiation interactions and neglect the indirect impacts of aerosol on cloud microphysical processes. However, given the uncertainty surrounding the microphysical interactions between clouds and absorbing aerosol, we consider this to be a reasonable limitation of our study, especially as the uncertainty in ERF from aerosol–radiation interactions contributes considerably to the total uncertainty in aerosol ERF (Forster et al., 2021).

It was always raining in Llamedos. Rain was the country's main export. It had rain mines.

— Terry Pratchett

3

Dependence of fast changes in global and local precipitation on the geographical location of absorbing aerosol

As shown in Chapter 2, the global-mean ERF from absorbing aerosol depends strongly on the geographical location of the aerosol plume, contrary to the assumptions made by previous ‘top-down’ studies of aerosol ERF. We found that the marked ‘location-dependence’ of global-mean ERF on the location of absorbing aerosol could be understood by thinking about how the diabatic heating from aerosol absorption was balanced in different atmospheric circulation regimes.

Following from this, a natural question is whether similar mechanisms also act to create a ‘location-dependence’ of precipitation changes. For example, do changes in global-mean precipitation also vary markedly between these experiments? This

This chapter is based on the following paper, which has been accepted at Journal of Climate:

Williams, A. I. L., Watson-Parris, D., Dagan, G. & Stier, P. “Dependence of fast changes in global and local precipitation on the geographical location of absorbing aerosol”. <https://doi.org/10.22541/au.167364749.93845737/v1>

I designed the study together with the co-authors. I carried out the simulations and analysed the data with input from the co-authors. I wrote the manuscript with contributions from the co-authors.

is particularly relevant as a similar ‘one-to-one’ relationship between changes in global-mean absorbing AOD and global-mean precipitation has also been assumed by recent studies (McCoy et al., 2022; Samset, 2022).

In addition to changes in global-mean precipitation, aerosol is also known to generate strong *local* changes in precipitation. On regional scales, the role of circulation changes is likely even larger than for global-mean scales, as discussed in the Introduction. However, despite this added complexity when considering local precipitation changes, recent work (Dagan et al., 2019, 2021) has argued that there should exist a ‘tropical-extratropical’ contrast in the local precipitation response to absorbing aerosol. However, the existence of these ‘contrasting local responses’ has not been tested systematically across the globe, nor has there been a robust physical mechanism provided for what controls variations in the local precipitation response in the tropics.

Hence, building on the work from Chapter 2, in Chapter 3 we aim to test whether the geographical location of absorbing aerosol has any control on the response of global and local precipitation. To do this we make use of the suite of idealized absorbing aerosol experiments outlined in Chapter 2. Similarly, our analysis focuses on how the atmospheric circulation (and thus cloudiness and precipitation) responds to the localized diabatic heating from absorbing aerosol, and how this varies between regions. A crucial aspect of this Chapter is to test whether the ‘one-to-one’ relationship between global-mean absorbing AOD and global-mean precipitation holds across our experiments, and if not to identify basic physical mechanisms for why it fails.

3.1 Introduction

Precipitation is a crucial component of the hydrological cycle and many aspects of human society have evolved in response to the dynamics of regional precipitation (Kathayat et al., 2022). Latent-heat release through precipitation is an essential part

of the atmospheric energy budget (Myhre et al., 2017; Pendergrass and Hartmann, 2012; Pendergrass and Hartmann, 2014) and is balanced by changes in atmospheric radiative cooling and sensible heat fluxes on global-scales (Allen and Ingram, 2002; Myhre et al., 2018). Furthermore, modelling work has suggested a link between the global-mean precipitation response and the equilibrium climate sensitivity (Watanabe et al., 2018). Thus understanding how human activity is altering the characteristics of precipitation is important from both a scientific and a societal perspective.

Anthropogenic aerosol emissions are known to influence precipitation through a number of pathways. Firstly, increase of aerosol enhances the total atmospheric extinction in a column through scattering and absorption. This reduces the amount of solar radiation reaching the surface and drives a reduction in surface fluxes, which causes global-mean precipitation to decrease due to energy budget constraints (Myhre et al., 2017; Ramanathan et al., 2001; Samset et al., 2016). Additionally, for absorbing aerosol (such as black carbon, dust or absorbing organic aerosol) the absorption of incident solar radiation heats the atmospheric column and acts as a direct diabatic heating perturbation (Brown et al., 2021; Stier et al., 2007; Williams et al., 2022a). This diabatic heating perturbation generally causes a decrease in global-mean precipitation due to energetic constraints (Pendergrass and Hartmann, 2012). Anthropogenic aerosol can also influence precipitation by influencing the microphysical properties of clouds (Albrecht, 1989; Dagan et al., 2017; Herbert et al., 2021a; Levin and Brenguier, 2009), although this pathway is subject to considerable uncertainty (Seifert et al., 2015; Stevens and Feingold, 2009) and thus in our current work we focus purely on the precipitation changes arising from aerosol-radiation interactions (ARI).

The response of precipitation to anthropogenic aerosol is commonly decomposed into ‘fast’ and ‘slow’ contributions (Samset et al., 2016; Voigt et al., 2017; Zhang et al., 2021). Fast responses, also referred to as rapid adjustments, occur in

the first hours or weeks following the perturbation and are independent of sea-surface temperature (SST) changes (and so are efficiently studied using fixed-SST simulations). Slow responses occur in the years and decades following the perturbation and are mediated by the SST changes. Some examples of the slow contribution to precipitation changes would be; shifts of the inter-tropical convergence zone (ITCZ) (Ming and Ramaswamy, 2009, 2011; Voigt et al., 2017; Wang, 2004, 2007, 2015; Zhang et al., 2021) or changes in monsoon precipitation (Dong et al., 2016, 2019a; Tian et al., 2018; Xie et al., 2020).

For the case of absorbing aerosol (such as black carbon), the Precipitation Driver Response Model Intercomparison Project (PDRMIP) (Myhre et al., 2017) has shown that the fast response dominates the global-mean precipitation response, and that this is unique among the forcings they studied (Samset et al., 2016; Stjern et al., 2017). Furthermore, the inter-model spread in fast precipitation responses to absorbing aerosol are dominated by aerosol-driven rapid adjustments in clouds and circulation. Rapid adjustments are particularly complicated for absorbing aerosol due to the fact that through absorption of incoming solar radiation, the aerosol acts as a source of diabatic heating to the atmospheric column (Johnson et al., 2019; Williams et al., 2022a). This localized diabatic heating can have significant impacts on the atmospheric thermodynamics and circulation by influencing moist processes, particularly in the tropics where clouds and circulation are intricately linked through radiative transfer, phase changes and SSTs.

In summary, the response of precipitation to absorbing aerosol perturbations is still uncertain and a significant portion of this uncertainty can be attributed to difficulties in understanding the ‘fast’ response of clouds and circulation to the localized diabatic heating which results from absorbing aerosol of solar radiation. Much of the previous work on this question has used either emissions-driven simulations (e.g., Persad and Caldeira, 2018; Westervelt et al., 2018) which are more realistic but bring with them the uncertainties associated with the conversion

of aerosol emissions to aerosol concentrations, or concentration-driven simulations (e.g., Samset et al., 2016) which require a further region-dependent mapping from aerosol concentrations to absorptivity. Taken together, this has made it difficult to explore the fundamental mechanisms which control the atmospheric response to regional aerosol heating.

Furthermore, previous studies have often focused on perturbations in a small number of regions. This is a sensible trade-off given the added complexity of emission-driven simulations (Persad and Caldeira, 2018) or the desire to have multiple models to compare between (Myhre et al., 2017), but does make it harder to understand the generality of these results, particularly because absorbing aerosol has a highly inhomogeneous spatial distribution (Kinne et al., 2013).

Recently, there has been work using aquaplanet (Dagan et al., 2019) and AMIP-style (Dagan et al., 2021) simulations with prescribed aerosol radiative properties to understand the influence of aerosol-radiation interactions on precipitation - which circumvents the translation from emissions to concentrations to absorption mentioned above. Interestingly, this work has shown that the local precipitation response to absorbing aerosol is positive in the tropics and negative in the extratropics, and has explained these responses using the local atmospheric energy budget. However, although these results are qualitatively similar between the aquaplanet simulations and AMIP-style simulations with land, only a few locations have been tested in AMIP-style simulations and thus it is unclear how the location of the aerosol perturbation mediates these changes.

To tackle these questions, we have used a large set of atmosphere-only simulations forced by identical absorbing aerosol perturbations across the whole globe to explore the sensitivity of the global- and local- precipitation response to absorbing aerosol. Our idealized experimental setup also allows us to better understand the physical mechanisms involved compared to previous work (Persad and Caldeira, 2018; Samset

et al., 2016) and thus we are able to explicitly link the precipitation responses to the fundamental physics governing the atmospheric circulation response to diabatic heating in different regions.

3.2 Methods

3.2.1 Experiment design

The experiments conducted in this study adopted the same design of (Williams et al., 2022a) (Chapter 2 of the thesis) to investigate the sensitivity of effective radiative forcing to changes in the spatial pattern of absorbing aerosol.

An ensemble of 145 fixed-SST simulations were conducted using the ICON (ICOsahedral Non-hydrostatic) general circulation model (icon-aes-1.3.00) which has a newly developed dynamical core and has been tuned to give a well-balanced top-of-atmosphere radiation budget and reasonable representation of precipitation compared to other atmosphere-only models and to observations (Crueger et al., 2018a; Giorgetta et al., 2018). The model uses a bulk mass-flux convection scheme (Nordeng, 1994; Tiedtke, 1989a), with cloud cover calculated locally using a relative humidity scheme (Sundqvist et al., 1989a). Cloud microphysics is based on the Lohmann and Roeckner scheme (Lohmann and Roeckner, 1996a), however the droplet number is fixed to prescribed profiles over land and ocean (Stevens et al., 2013a) and the model does not simulate aerosol-cloud interactions. Radiation is parameterized using the PSrad radiative transfer scheme (Pincus and Stevens, 2013a). The radiation scheme is called two times (both all- and clear-sky) at every radiation time-step, once with aerosol and once without aerosol. Taking the difference between these two ‘double-calls’ gives the total aerosol radiative effect, and the instantaneous aerosol radiative forcing is calculated by differencing the total aerosol radiative effect between a perturbed run and the control.

The simulations were run on a triangular grid (R_2B_4 specification) corresponding to an approximately uniform horizontal grid spacing of 160km. The model uses a terrain-following vertical sigma-height grid (Leuenberger et al., 2010) with 47 levels between the surface and the model top at 83km. Each simulation, including the control run, is integrated for 20 years, after which the first year is discarded as spin-up and analysis is conducted over the remaining 19 years. The simulation length was chosen as a trade-off between computational constraints and the requirement that our runs are long enough to give robust results in the face of internal variability.

As we are focusing on the ‘fast’ response of precipitation to anthropogenic absorbing aerosol, we run the simulations with a prescribed monthly climatology of sea-surface temperatures and sea-ice concentrations derived from time-averaged Atmospheric Model Intercomparison Project (Gates, 1992) (AMIP) boundary conditions over the 1979-2016 period.

Aside from the control run, the only difference between the remaining 144 simulations is the inclusion of an idealized plume of absorbing aerosol at 144 different geographical locations (Fig 2.1 a), parametrized using the Max Planck Institute Aerosol Climatology version 2, Simple Plume (MACv2-SP) model (Stevens et al., 2017). The extinction is prescribed at 550nm and then extrapolated to other wavelengths in the shortwave spectrum using an Ångström parameter of 2; no aerosol extinction is prescribed in the longwave. The plume’s horizontal structure is specified to be a 2-dimensional Gaussian function with a standard deviation of 10° degrees. The vertical distribution of aerosol is based on the kernel of Euler’s β -function (see details in Stevens et al., 2017), with most of the aerosol concentrated in the lower 3km of the atmosphere and a peak at ≈ 1 km. We note that differences in the vertical plume profile may cause different responses of the climate system and general circulation (Ban-Weiss et al., 2012; Persad et al., 2012), however Williams et al., 2022a showed that their results were robust to realistic variations in the shape and height of the aerosol perturbation. Both the horizontal and vertical structure

of the plume are constant in time and the shape of the plume is not influenced by the local meteorology.

In each of the aerosol perturbation experiments the plume is prescribed with a total column AOD=1 and a single-scattering albedo (SSA) of 0.8, both defined at a wavelength of 550nm. This corresponds to a strongly absorbing plume of aerosol, generating a large diabatic heating perturbation in the lower atmosphere, even while generating cooling at the surface (primarily over land in our experiments due to the use of fixed-SSTs). Furthermore, to isolate the role of the radiative impact of the plume we do not include the microphysical effects of the aerosol plume which are also described in Stevens et al. (2017).

3.2.2 Atmospheric energy budget decomposition

From an energetic perspective, on global-scales and at equilibrium, the net energy input to the atmosphere (shortwave absorption of incoming solar radiation and latent heat released by precipitation) must be balanced by atmospheric radiative cooling and surface sensible heat fluxes (Dagan et al., 2021; Muller and O’Gorman, 2011; Pendergrass and Hartmann, 2014; Zhang et al., 2021). Mathematically we can write this as,

$$LP + Q_{\text{SW}} = Q_{\text{LW}} + SH, \quad (3.1)$$

where L is the latent heat of condensation of water vapour, P is the precipitation rate, Q_{SW} is the shortwave absorption, Q_{LW} is the longwave radiative cooling and SH is the upwards sensible heat flux at the surface. Taking a difference between our control simulation and simulations with an absorbing aerosol perturbation and rearranging yields a well-known constraint on the global-mean precipitation change associated with an anthropogenic perturbation (e.g., Allen and Ingram, 2002; Myhre et al., 2018),

$$L\delta P = \delta Q_{\text{LW}} - \delta Q_{\text{SW}} + \delta SH. \quad (3.2)$$

We use Eq. 3.2 to understand the fast changes (i.e., under fixed-SSTs) in global-mean precipitation in each of our 144 absorbing aerosol experiments. We calculate the Q_{LW} and Q_{SW} terms in Eq. 3.2 as follows:

$$Q_{\text{LW}} = LW_{\text{TOA}}^{\uparrow} + (LW_{\text{SFC}}^{\downarrow} - LW_{\text{SFC}}^{\uparrow}) \quad (3.3)$$

and

$$Q_{\text{SW}} = (SW_{\text{TOA}}^{\downarrow} - SW_{\text{TOA}}^{\uparrow}) + (SW_{\text{SFC}}^{\uparrow} - SW_{\text{SFC}}^{\downarrow}), \quad (3.4)$$

with all fluxes defined as positive upwards. We further decompose the δQ_{SW} into a component associated with the *instantaneous* change in radiative fluxes associated with the inclusion of the aerosol, $\delta Q_{\text{SW}}^{\text{inst}}$, and a component which represents the change in Q_{SW} which results from rapid-adjustments, $\delta Q_{\text{SW}}^{\text{adj}}$, such as changes in clouds, atmospheric water vapor or surface albedo.

All of the terms on the right-hand side of Eq. 3.2 are initially calculated in energy units (Wm^{-2}), before being transformed into ‘precipitation units’ (mm day^{-1}) for ease of presentation.

3.3 Sensitivity of global-mean precipitation to localised absorbing aerosol perturbations

In Fig. 3.1 we have plotted the change in global-mean precipitation associated with each of the 144 absorbing aerosol experiments conducted (a), along with a decomposition (following Eq. 2) into contributions from longwave cooling (b), surface sensible heat fluxes (c) and shortwave absorption (d). Panels (e) and (f) decompose the changes in shortwave absorption into changes arising from instantaneous changes and rapid-adjustments, respectively. Furthermore, to aid interpretation, changes in

shortwave absorption have been multiplied by -1 in (d), (e) and (f) (to account for the negative sign in Eq. 3.2) so that the total precipitation change in (a) can be reconstructed by summing (b) + (c) + (d).

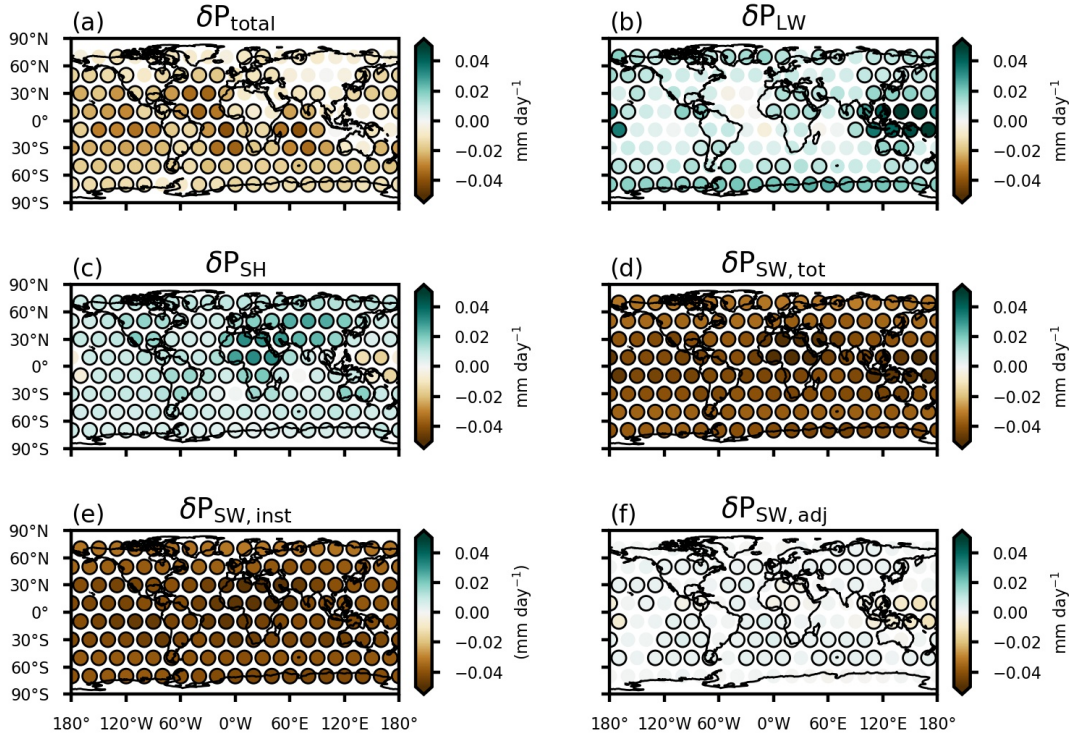


Figure 3.1: Global- and time-mean changes in precipitation (a) and contributions from changes in longwave cooling (b), surface sensible heat flux (c) and shortwave absorption (d) according to Eq. (2). Panels e and f show a further decomposition of the shortwave absorption term into contributions from instantaneous radiative fluxes and rapid adjustments, respectively. Each colored circle represents the results from the simulation imposing an aerosol perturbation in that location. Black enclosing circles are present where the time-series of global- and annual-mean changes are statistically different from zero at the 99% level according to a 1-sided t-test.

There is a remarkable amount of variation in the global-mean precipitation change in response to the absorbing aerosol plumes in different locations (Fig. 1a). This is particularly striking as the aerosol perturbation in each of these experiments is identical and only the location is varied. All aerosol locations result in a decrease in global-mean precipitation, but the response is weaker over parts of Northern Hemisphere land, and the Sahara. Notably, the response of global-mean

precipitation to absorbing aerosol over the Western tropical Pacific is also negligible.

3.3.1 Contribution from shortwave absorption

To first-order, the decrease in global-mean precipitation is set by the additional shortwave absorption of incoming solar radiation (δP_{SW} , Fig. 3.1d) which results from the inclusion of the absorbing aerosol plumes. Most of this change is due to the contribution from instantaneous radiative fluxes, as shown in Fig. 3.1e. This contribution drops off slightly towards higher latitudes due to reductions in incoming solar radiation (Fig. 3.2). Rapid-adjustments also contribute towards changes in shortwave absorption, but this is a much smaller effect (Fig. 3.1f), and occurs primarily in response to absorbing aerosol over the Western tropical Pacific, which is associated with a strong convective response to diabatic heating as demonstrated by Williams et al., 2022a. Overall, the contribution of δP_{SW} to global-mean precipitation changes is rather agnostic to the longitudinal position of the absorbing aerosol plume.

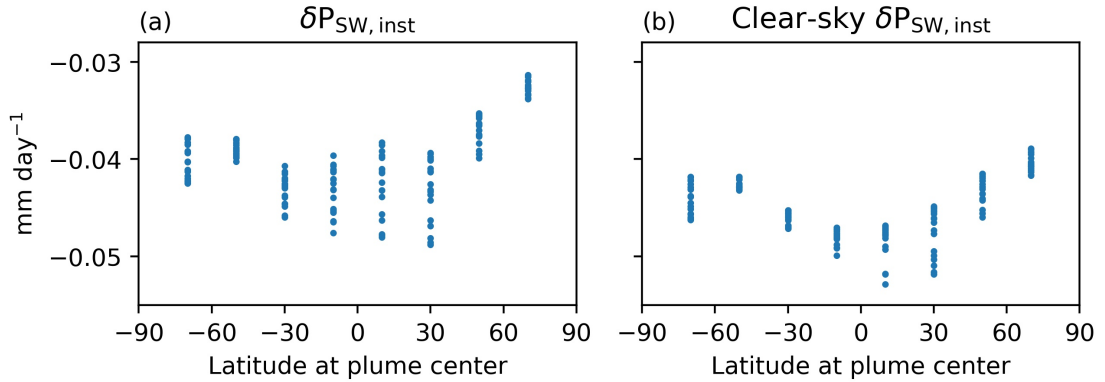


Figure 3.2: Global- and time-mean changes in precipitation resulting from instantaneous changes in shortwave absorption (SW_{inst} , panel a), and the contribution from clear-sky (b). This plot shows that the contribution of shortwave absorption drops off towards the poles because of the decrease in incoming solar radiation.

3.3.2 Contribution from sensible heat fluxes

The sensitivity of fast changes in global-mean precipitation to variations in the geographical position of absorbing aerosol arises predominantly through changes in SH flux and longwave cooling and (Fig. 3.1c and b, respectively). Firstly, we focus on the SH contribution. The SH flux contributes positively to the global-mean precipitation change for almost all locations, indicating a reduction in the global-mean upwards surface sensible heat flux. This positive contribution offsets some of the negative global-mean precipitation changes associated with increased shortwave absorption (particularly over land). The reduction in global-mean upwards sensible heat flux (and thus positive contribution to global-mean precipitation changes) occurs because lofted aerosol tends to generate local surface cooling due to the reduction in incoming solar radiation from absorption and scattering processes (e.g., Dagan et al., 2020; Muller and O’Gorman, 2011; Samset et al., 2016; Williams et al., 2022a). The particularly strong reduction in SH fluxes in response to absorbing aerosol over the Sahara and the Eurasian continent is associated with stronger local surface cooling over this region in response to the plume, and contributes to a muted response of global-mean precipitation in response to absorbing aerosol over these regions.

The exception to this occurs in response to absorbing aerosol over the Western tropical Pacific, where the SH changes contribute negatively to global-mean precipitation, which indicates an increase in global-mean upwards surface sensible heat flux in response to absorbing aerosol in these regions. This can be understood by noting that absorbing aerosol over this region is associated with a strong convective response, an increase in upper-level divergence and a strengthening of the tropical Walker Circulation (as shown previously in Chapter 2; Williams et al., 2022a). This strengthening of the Walker Circulation is associated with an intensification of low-level winds, which increase the upwards surface sensible heat flux. It should be noted that black carbon has been found to also increase the lower tropospheric stability and thus reduce surface SH fluxes (Myhre et al., 2018), but in our experiments the

dominant mechanism appears to be an intensification of the large-scale tropical circulation and low-level winds.

3.3.3 Contribution from longwave cooling

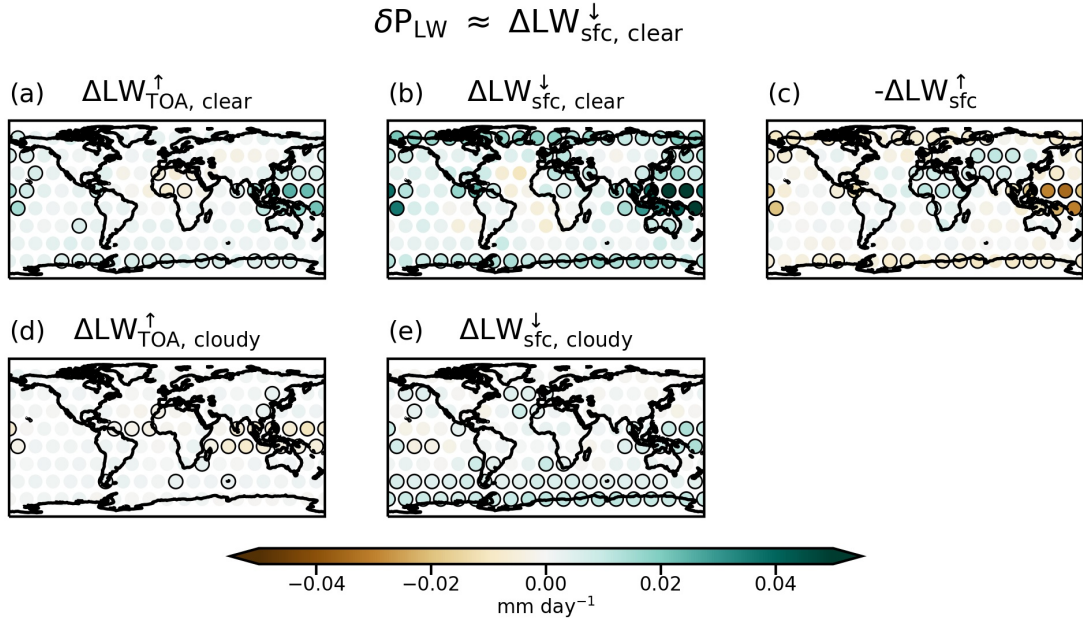


Figure 3.3: Decomposition of δP_{LW} into contributions from changes in outgoing LW radiation at top-of-atmosphere in clear (a) and cloudy (d) sky, downward LW emission at the surface in clear (b) and cloudy (e) sky, and upward LW emission at the surface (c). Each colored circle represents the results from the simulation imposing an aerosol perturbation in that location.

In addition to sensible heat flux, changes in global-mean atmospheric longwave cooling are also a large contributor to the ‘location dependence’ of fast changes in global-mean precipitation (δP_{LW} , Fig. 3.1b). Generally, δP_{LW} contributes positively to changes in global-mean precipitation, and there is a notable ‘hot spot’ of positive δP_{LW} in response to absorbing aerosol over the Western tropical Pacific (Fig. 3.1b). In our experiments, changes in global-mean longwave cooling are dominated by changes in clear-sky rather than cloudy-sky (Fig. 3.3). To a good approximation, the global-mean changes in longwave cooling are largely controlled by changes in downwelling longwave emission from the atmosphere to the surface (compare Fig.

3.1b and 3.2b). We also note that there is a negative contribution from cloudy-sky outgoing LW radiation, which is consistent with the strong convective response over this region generating more anvil clouds which reduce the outgoing LW radiation locally.

To gain some intuition for what controls these changes in $LW_{\text{sfc, clear}}^{\downarrow}$, we assume that they arise due to changes in lower-tropospheric temperature and associated changes in blackbody emission. If we assume that the emission from the lower-troposphere can be approximated as σT_{850}^4 , then the predicted change in $LW_{\text{sfc, clear}}^{\downarrow}$ following a change in global-mean 850hPa temperature (ΔT_{850}) is equal to $4\sigma T_{850}^3 \Delta T_{850}$. In Fig. 3.4 we show that this estimate captures most of the variation in global-mean clear-sky, downwelling LW emission across our plume experiments, and in particular the large values of $\Delta LW_{\text{sfc, clear}}^{\downarrow}$ which occur in response to absorbing aerosol over the Western tropical Pacific. The results in Fig. 3.4 are not sensitive to the precise vertical level we choose. We note that our ‘temperature-only’ scaling of $LW_{\text{sfc, clear}}^{\downarrow}$ generally underestimates the true response, and this is likely because it does not include the effect of increases in water vapor in response to the aerosol plume, which would also increase $\Delta LW_{\text{sfc, clear}}^{\downarrow}$.

We have now shown that changes in clear-sky downwelling LW emission are the dominant contributor to δP_{LW} (Fig. 3.3), and generally contribute to an increase in global-mean precipitation, with a ‘hot spot’ of positive global-mean δP_{LW} in response to absorbing aerosol over the Western tropical Pacific (Fig. 3.1b). We have also shown that these changes in downwelling LW emission are highly correlated with changes in tropospheric temperatures (Fig. 3.4). So, the question now becomes: why does absorbing aerosol generally increase tropospheric temperatures, and cause such a large increase when placed in the Western tropical Pacific?

As we discussed earlier, absorbing aerosol warms the atmosphere locally through absorbing incoming solar radiation. This explains why the changes in Fig. 3.3b

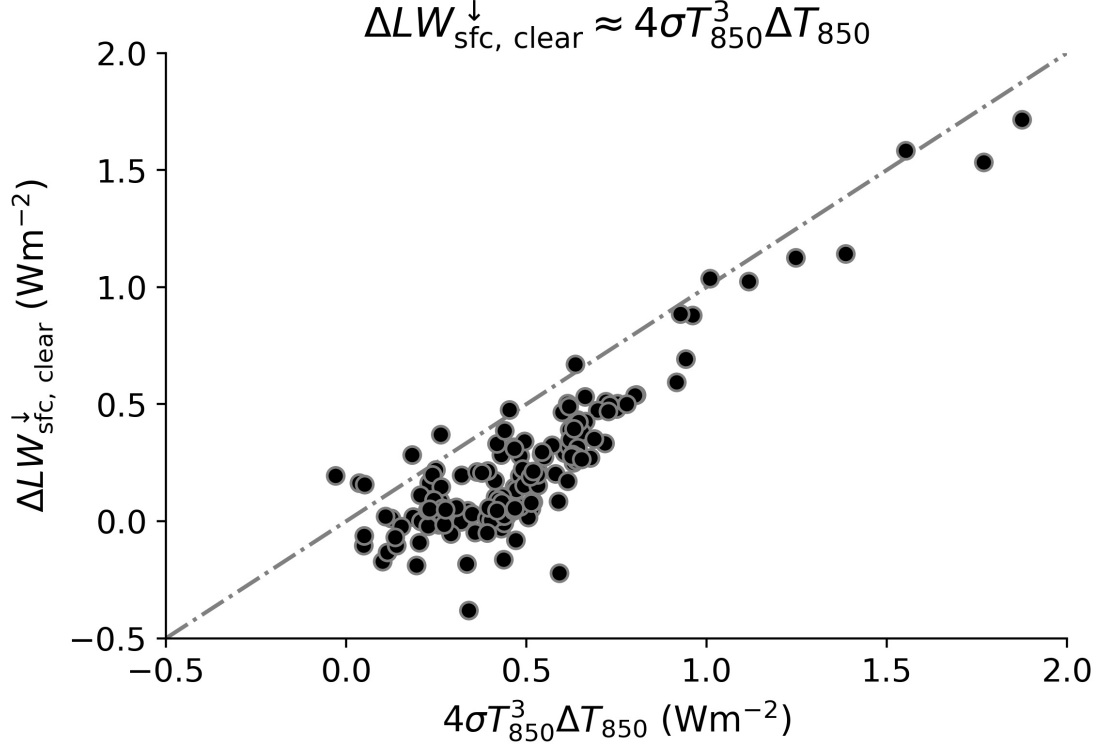


Figure 3.4: Scatter plot of global-mean changes in clear-sky, downwelling LW emission from the atmosphere to the surface versus an estimate of these changes which accounts only for altered blackbody emission from the 850hPa layer.

are generally positive, as the absorption warms the atmosphere aloft and causes an increase in downwelling LW emission. With regards to the ‘hot spot’ of strong δP_{LW} over the Western tropical Pacific, this arises primarily because the Western tropical Pacific is a region of intense and persistent deep convection throughout the annual cycle. One of the cornerstones of our understanding of tropical dynamics, the ‘convective quasi-equilibrium’ hypothesis, states that deep convection acts to maintain a moist adiabatic temperature profile which is pegged to the temperature of the lower-troposphere (e.g., Emanuel, 2007). Hence, when absorbing aerosol is placed in the Western tropical Pacific, it is able to warm the whole atmospheric column due to the action of deep convection. This warming of the local atmospheric column is then transmitted horizontally across the whole of the tropical free-troposphere through the action of gravity waves (Charney, 1963). As a result, absorbing aerosol in the Western tropical Pacific is able to cause strong, non-local warming of the

whole tropical free-troposphere. This mechanism appears to be less efficient in other regions of the tropics which are typically associated with deep convection (e.g., the Amazon). This is potentially because the interactive land surface enables surface cooling from the aerosol to take hold and reduce the subcloud buoyancy (we will discuss this point in more detail in the section on local responses in the tropics).

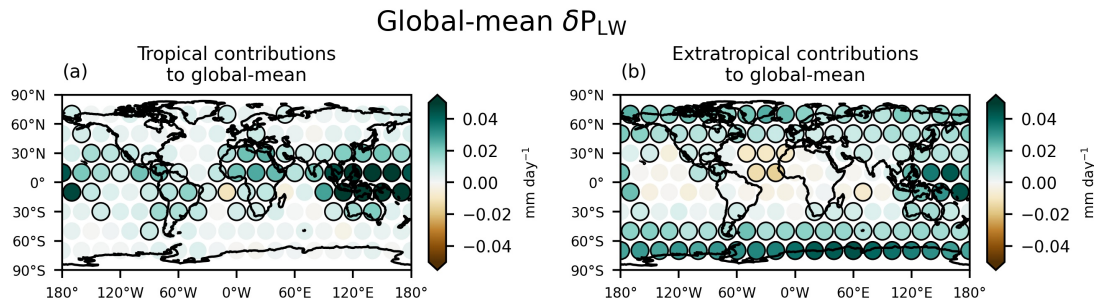


Figure 3.5: Contributions of tropical changes ($\pm 30^\circ$ latitude) (a) and extratropical changes (everywhere else) (b) to the global-mean changes in δP_{LW} , the sum of panels (a) and (b) is equal to Fig. 1b. Each colored circle represents the results from the simulation imposing an aerosol perturbation in that location.

While tropical warming contributes to increased δP_{LW} in response to absorbing aerosol over the Western tropical Pacific, the ‘hot spot’ of δP_{LW} over the Western tropical Pacific (Fig. 3.1b) also has an additional, smaller contribution from the extratropics (Fig. 3.5). This raises the question: how can an absorbing aerosol perturbation in the tropics influence the extratropics? A potential explanation for this is that absorbing aerosol plumes over this region are associated with a strong convective response and an increase in upper-level divergence (Williams et al., 2022a). Previous work (e.g., Sardeshmukh and Hoskins, 1988) has shown that tropical upper-level divergence is associated with teleconnections to the midlatitudes because the upper-level divergence can act as a vorticity source, and the resultant Rossby waves then propagate along great-circle paths into the midlatitudes. Figure 3.6b shows the response of 200hPa meridional wind to absorbing aerosol in the Western tropical Pacific, and a stationary Rossby wave can be seen to propagate from the tropics into the extratropics. In contrast, absorbing aerosol in the East

Pacific is not associated with such a teleconnection (Fig. 3.6d). The plume in Fig. 3.6b was chosen because it most clearly demonstrates the Rossby wave train, but wave trains exist for most of the Western tropical Pacific plumes.

To better quantify this proposed teleconnection mechanism, in Fig. 3.6a,c we plot the Rossby wave source (S) calculated following Sardeshmukh and Hoskins (1988) as,

$$S = -\mathbf{u}_\chi \cdot \nabla(\zeta + f) - (\zeta + f)\nabla \cdot \mathbf{u}_\chi, \quad (3.5)$$

where $\mathbf{u}_\chi$ is the divergent component of the horizontal wind, ζ is the relative vorticity and f is the planetary vorticity, all terms are taken at 200hPa. In simple terms, Eq. 3.5 has two contributions. The first term represents vorticity advection by the divergent wind, meanwhile the second term represents the generation of vorticity by ‘vortex stretching’. This ‘vortex stretching’ term arises as a consequence of angular momentum conservation. For example, upper-level divergence (like that generated by deep convection) means $\nabla \cdot \mathbf{u}_\chi > 0$, which causes the fluid column to elongate and contract. This contraction causes a spin-up (increase in vorticity) in the same way as an ice-skater spins up when they ‘contract’.

Panels a and c of Fig. 3.6 show the change in 200hPa Rossby wave source (ΔS_{200}) for absorbing aerosol in the Western and Eastern tropical Pacific, respectively. In order to focus on the tropical wave source (which is more relevant for our mechanism) we have tapered the ΔS_{200} fields to zero between 30-40° of the equator. Panel a demonstrates that the Western Pacific aerosol generates a significant Rossby wave source to the west of the perturbation, with a positive ΔS_{200} south of the equator and a negative ΔS_{200} North of the equator. This region of strong ΔS_{200} also corresponds to the region of initiation of the stationary Rossby wave in Fig. 3.6b, which begins in the Indian Ocean, westward of the absorbing aerosol itself. In contrast, the ΔS_{200} perturbation is negligible for the aerosol perturbation in the East Pacific.

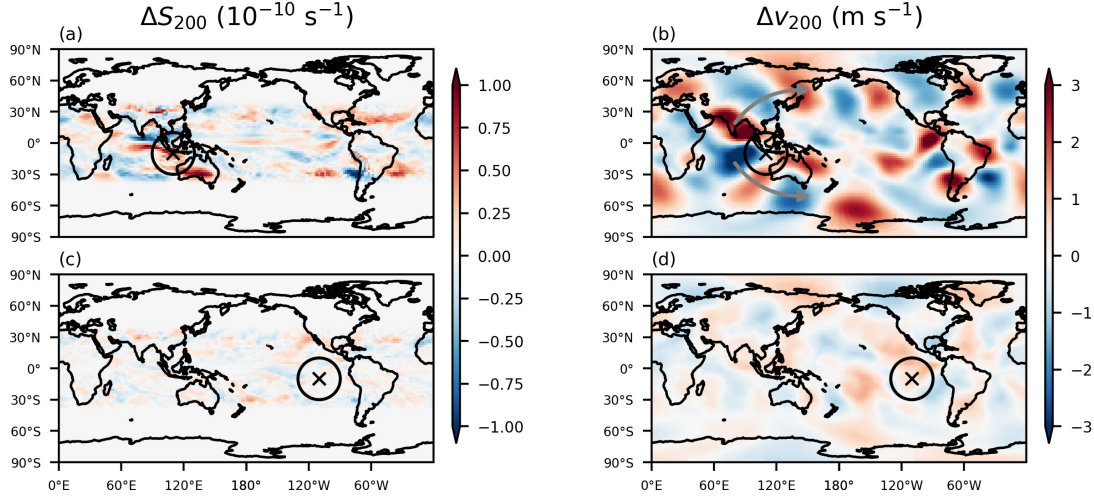


Figure 3.6: (a,c) the change in 200hPa Rossby wave source (ΔS_{200}) for absorbing aerosol in the Western tropical Pacific (a) and the Eastern tropical Pacific (c). To focus on the tropical component of the Rossby wave source, the fields have been linearly tapered to zero between 30-40° of the equator. (b,d) the change in 200hPa meridional wind (Δv_{200}) for absorbing aerosol in the Western tropical Pacific (b) and the Eastern tropical Pacific (d). A stationary Rossby wave train (highlighted by the grey arrows) which emanates from the Indian Ocean into the midlatitudes in both hemispheres can be seen in (b), whereas no such change occurs in response to the absorbing aerosol in the Eastern tropical Pacific (d).

Having propagated across the Pacific basin, the Rossby wave-train becomes less pronounced (Fig. 3.6b) and this is likely due to the twin influences of orography and interference with additional Rossby waves which are initiated in response to the Rossby wave source hotspots in the tropical Atlantic and Eastern Pacific (Fig. 3.6a). It is also worth noting that while upper-level divergence in the tropics can act as a Rossby wave source, whether or not the Rossby waves can propagate into the midlatitudes depends sensitively on the background zonal winds, which can act as atmospheric ‘waveguides’ which determine the turning latitude of the waves (Hoskins and Ambrizzi, 1993). As a result the ability of the Rossby waves to propagate into the midlatitudes depends not only on the upper-level divergence but also the local state of the upper-level zonal winds. This, along with a reduced convective response over land in response to the aerosol (which we will discuss further in the section on local responses in the tropics), helps explain why we do not see strong Rossby wave responses in other parts of the tropics.

To summarize this section, we have shown that global-mean changes in surface sensible heat fluxes and longwave atmospheric cooling are a strong determinant of the ‘location dependence’ of fast, global-mean precipitation changes in response to absorbing aerosol in different locations. The sensible heat flux changes are most important over land and over regions where the surface cooling from aerosol is most pronounced (namely the Sahara and Eurasia). The longwave cooling term also contributes to this location-dependence, and these changes are controlled to first-order by downwelling emission from the atmosphere to the surface. We have argued that these changes can be explained largely by changes in lower-tropospheric temperatures, which explains the general positive contribution of this term to the global-mean precipitation changes as absorbing aerosol tends to warm the atmosphere aloft. We also noted a ‘hot spot’ of δP_{LW} in response to absorbing aerosol over the Western tropical Pacific, which are associated with global-scale changes in temperature and downwelling LW emission. To explain this, we argue that the Western tropical Pacific is unique in two regards. Firstly, heating in this region is able to be communicated across the entire tropical free-troposphere due to persistent deep convection, which transports energy in the vertical, and the action of gravity waves which efficiently homogenize temperatures in the tropical free-troposphere. Secondly, heating in the Western tropical Pacific also has the potential to influence the extratropics through generating upper-level divergence and exciting a Rossby wave train. The precise mechanisms by which these Rossby wave trains influence the extratropics are difficult to disentangle in our experiments, in part because we only archived monthly-mean data. However, one plausible explanation could be that the Rossby waves generate an increase in the poleward eddy heat flux (Chemke and Polvani, 2020), thus transferring more heat from the tropics to the extratropics. Another possible explanation could involve the fact that Rossby waves generally induce an alternating pattern anomalous poleward and equatorward flow (as can be seen in Fig. 3.6b). If the regions of anomalous poleward flow coincide with land regions (which have interactive surface temperatures) but the regions

anomalous equatorward flow coincide most with ocean regions (with fixed SSTs) then this could also generate an increase in extratropical temperatures. Future work is needed to fully disentangle the contributions of these potential mechanisms.

3.4 Sensitivity of local precipitation changes to localised absorbing aerosol perturbations

Having investigated the mechanisms governing the sensitivity of global-mean precipitation changes to variations in the geographical location of absorbing aerosol, we now move on to examining the sensitivity of *local* precipitation changes. In Fig. 3.7 we show the ‘local’ changes in precipitation (δP_{local}) in each of the 144 simulations, we define δP_{local} as the fast changes in precipitation averaged over an area containing the 95% of the aerosol perturbation. Our results are not sensitive to this precise definition of δP_{local} .

Our results in Fig. 3.7 demonstrate that the local precipitation response to the radiative effects of absorbing aerosol is generally positive in the deep tropics and switches to negative in the extratropics, in agreement with previous work using both idealized aquaplanet simulations (Dagan et al., 2019) and more realistic AMIP-style simulations (Dagan et al., 2021).

An intuitive way to understand these variations in the local response of precipitation to absorbing aerosol is to consider the temperature tendency equation for the region covered by the aerosol plume, written as,

$$\frac{\partial T}{\partial t} = \frac{\Delta Q_{\text{aero}}}{c_p} + \text{balance terms} \quad (3.6)$$

where ΔQ_{aero} represents the (positive) temperature tendency driven by the shortwave absorption from the aerosol plume and c_p is the heat capacity of dry air. In the absence of terms which balance this heating tendency, the temperature in

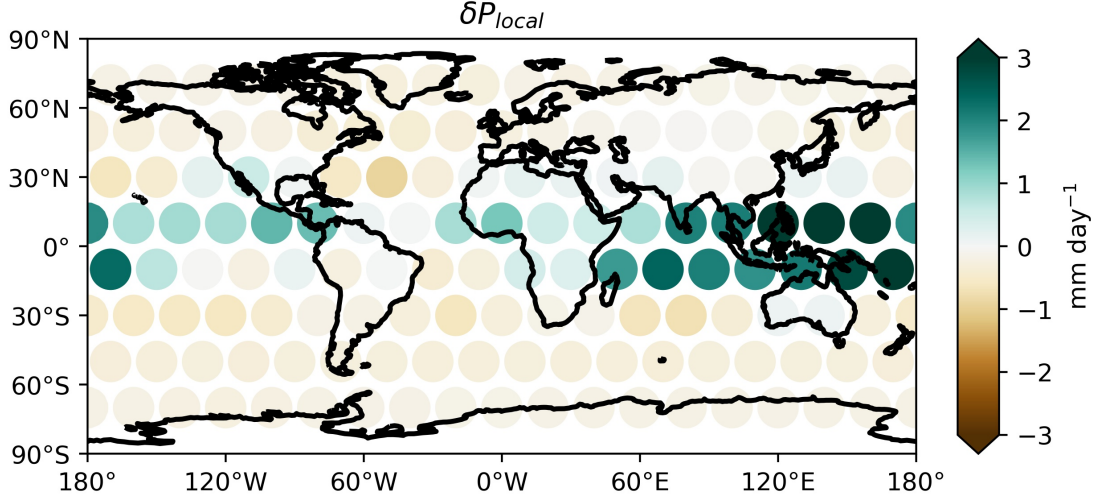


Figure 3.7: Map of the local changes in precipitation for each of the 144 aerosol perturbation experiments. Each colored circle represents the results from the simulation imposing an aerosol perturbation in that location. Because of the long integration times (20 years per experiment) and the definition of ‘local’, all of the changes here are significant and so in order to keep the plot legible we have chosen not to add black circles.

this region would increase without bound. The Planck response helps to balance this shortwave heating tendency, in addition to changes in precipitation and surface fluxes (Williams et al., 2022a). Here though we will focus on the role that changes in the atmospheric circulation play in helping to balance this imposed shortwave heating. In the extratropics, linear quasi-geostrophic theory (Hoskins and Karoly, 1981; Sardeshmukh and Hoskins, 1988; Vallis, 2017; Williams et al., 2022a) predicts that diabatic heating (such as through absorbing aerosol) will tend to result in the formation of a downstream low pressure region and a cyclonic circulation (in the Northern Hemisphere). This is illustrated in Fig. 3.8, with the low-pressure system located over the Hudson Bay to the North-East of the aerosol plume. The downstream low helps to balance the heating from the absorbing aerosol by drawing down cold air from higher latitudes. Additionally, this meridional advection of high-latitude air explains the local decrease in precipitation for these extratropical perturbations because the air drawn down by the low pressure system is associated with a lower moisture content and a low relative humidity, both of which inhibit precipitation formation. This decrease in precipitation then also helps to balance

the local aerosol heating through a reduction in latent heat release.

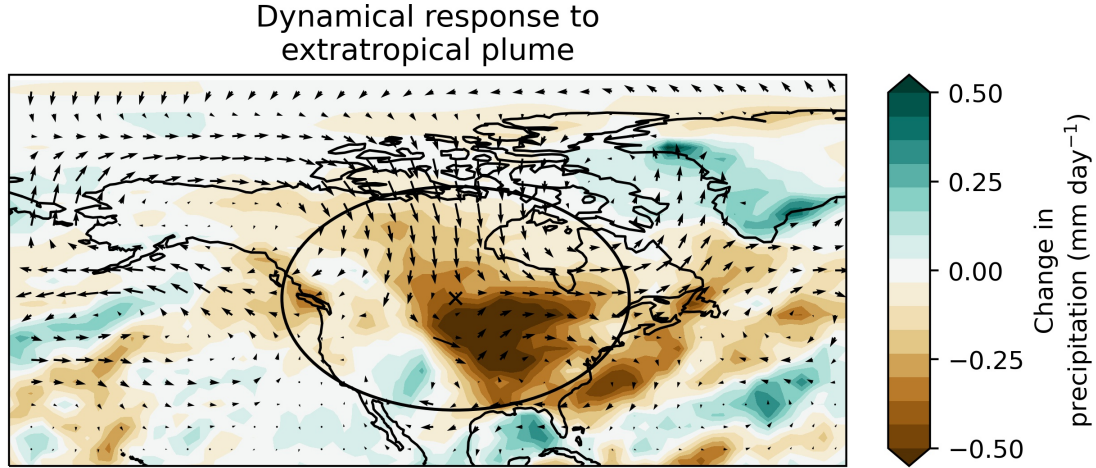


Figure 3.8: Map showing the dynamical response of the midlatitude atmospheric circulation to aerosol heating. The black ellipse denotes the area encompassing 95% of the total absorbing aerosol perturbation, the background contours show the change in precipitation compared to the control run and the arrows show the corresponding changes in 850hPa winds.

In summary, the negative precipitation response to absorbing aerosol in the extratropics can be understood in terms of the atmospheric circulation changes which are required in order to balance the local aerosol heating. This framing can also be used to explain the positive δP_{local} response in the tropics, albeit by a very different set of mechanisms. In the tropics, the weak Coriolis force means that strong gradients of pressure and temperature cannot persist, which is the basis of so-called ‘weak-temperature-gradient’ (WTG) theories of the tropical atmosphere (e.g., Pierrehumbert, 1995; Sobel et al., 2001; Williams et al., 2022b; Zhang and Fueglistaler, 2020). WTG dictates that diabatic heating anomalies in the tropics (e.g., arising from absorbing aerosol) are balanced by adiabatic cooling from increased ascent. This increase in ascent drives an anomalous overturning circulation in the vicinity of the heating, which is associated with low-level moisture convergence and an increase in precipitation (Dagan et al., 2019). Thus, WTG theory qualitatively explains the positive δP_{local} response in the tropics.

So far, we have used a simple framework to explain the contrasting δP_{local} responses between the tropics and the extratropics in terms of the dynamical response of the atmospheric circulation to localized aerosol heating. However, while this framework nicely encapsulates the differing response between the tropics and the extratropics, our results also demonstrate that the longitudinal position of the aerosol is important in determining its local precipitation response, particularly in the tropics. For example, Fig. 3.7 shows that δP_{local} is much stronger in the Western tropical Pacific than in the Eastern tropical Pacific and the tropical Atlantic. To understand these varied responses to the longitudinal position of tropical absorbing aerosol, in Fig. 3.9a we have plotted δP_{local} versus the local change in mid-tropospheric vertical velocity (ω_{500}) for each of the 144 simulations. Figure 3.9a demonstrates clearly that stronger local precipitation changes are associated with increased local ascent, which drives a stronger overturning circulation and a larger low-level moisture convergence near the heating region.

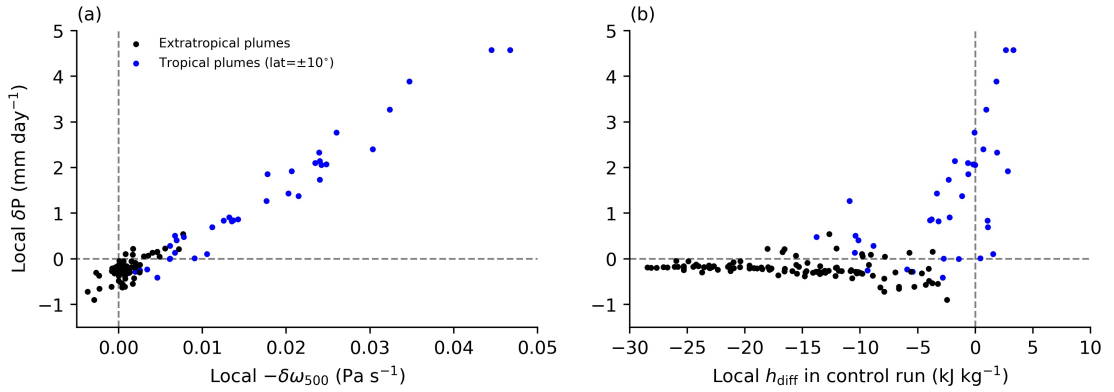


Figure 3.9: (a) Scatter plot of local changes in precipitation versus local changes in mid-tropospheric vertical velocities (ω_{500}), where the $\delta\omega_{500}$ values have been multiplied by -1 and so positive values on the x-axis indicate increased ascent over the heating region. (b) Scatter plot of local changes in precipitation versus the local h_{diff} over the same region, but in the control run. The h_{diff} coordinate is defined in the main text and acts as a measures of the regions of the tropical atmosphere which are unstable to deep convection.

While this exercise is a useful sanity check, it does not in-and-of-itself provide us with any predictive power. Indeed, if asked the question, “Which regions do we *a priori* expect to have a strong local precipitation response to absorbing aerosol?”

we would currently be unable to provide an answer. To tackle the questionings of our imagined critic, we follow the work of previous studies (Williams et al., 2022b; Williams and Pierrehumbert, 2017; Williams et al., 2009) and introduce a simple measure of sub-cloud buoyancy, h_{diff} , which we define as,

$$h_{\text{diff}} = h_0 - h_{500}^*, \quad (3.7)$$

where h_0 is the near-surface moist static energy and h_{500}^* is the saturated moist static energy of the free troposphere, again taken at 500hPa (e.g., Emanuel, 2007; Emanuel et al., 1994b; Singh et al., 2017; Williams et al., 2022b; Williams et al., 2009)¹. The atmosphere is stable against deep convection when h_{diff} is negative, and is unstable to deep convection when h_{diff} is significantly positive. In Fig. 3.9b we again plot δP_{local} but this time as a function of the local values of h_{diff} over the same region in the control simulation, which shows that the smallest values of δP_{local} occur when the absorbing aerosol is located in regions which are stable to deep convection ($h_{\text{diff}} < 0$) and the largest values occur when the aerosol is placed in unstable regions ($h_{\text{diff}} > 0$). However, the transition between regions associated with a large δP_{local} and a small δP_{local} does not simply occur at the transition between positive and negative values of h_{diff} , and in fact occurs when h_{diff} is slightly negative. This is because in regions which are weakly stable to deep convection in the control run, the additional heating from the absorbing aerosol causes them to transition to a deeply convective regime with $h_{\text{diff}} > 0$ and thus a larger increase in precipitation.

Although it is hard to predict the exact change in h_{diff} which arises due to the absorbing aerosol, we can gain some insight by assuming that the absorption alters h_{diff} primarily by altering h_0 through changes in low-level temperature (δT_0).

¹Similar results are found if we scatter the local change in precipitation against the baseline ω_{500} , but the causality is somewhat confusing here because while ω_{500} is correlated with sub-cloud buoyancy, it is in fact a result and not a cause of the convection. We also show later on that h_{diff} lets us have more analytical insight into the behaviour.

Noting that $h_0 = c_p T_0 + L_v q_0 + g z_0$, we can write this as,

$$\delta h_{\text{diff}} = \delta(h_0 - h_{500}^*) \approx \delta h_0 = \delta(c_p T_0 + L_v q_0), \quad (3.8)$$

where q_0 is the low-level specific humidity, and we have taken the changes at a fixed height. Assuming constant relative humidity (RH), we can also write the $L\delta q_0$ term as,

$$L_v \delta q_0 = L_v \text{RH}_0 \delta(e_s(T_0)) \approx L_v \text{RH}_0 \left(\frac{de_s(T_0)}{dT} \delta T_0 \right) = L_v \left(\frac{L_v \text{RH}_0 e_s(T_0)}{R_v T_0^2} \delta T_0 \right), \quad (3.9)$$

where $e_s(T_0)$ is the saturation vapour pressure and we have linearized the saturation vapour pressure around the baseline temperature T_0 . Combining this with Eq. 3.8 we arrive at a simple prediction for δh_{diff} in response to a change in boundary-layer temperature,

$$\delta h_{\text{diff}} \approx \left(c_p + \frac{\text{RH}_0 L_v^2}{R_v T_0^2} e_s(T_0) \right) \delta T_0. \quad (3.10)$$

This equation allows us to gain insight into the change in h_{diff} (and thus local precipitation) which may arise due to changes in the boundary-layer temperature distribution (δT_0).

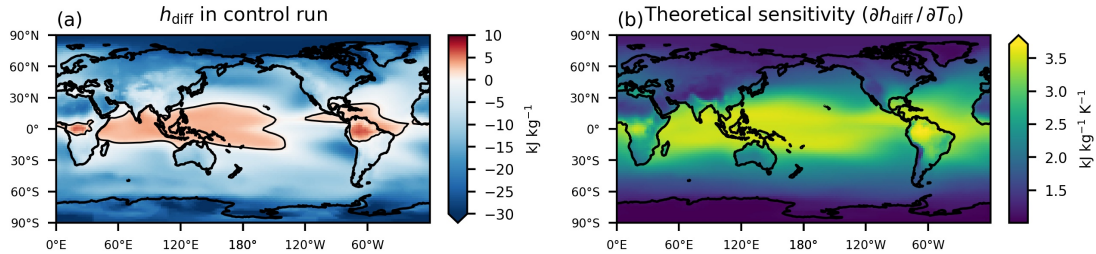


Figure 3.10: Map showing the distribution of time-mean h_{diff} in the control run (panel a), with the zero contour drawn in black. Also shown in panel b is the theoretical sensitivity of h_{diff} to warming calculated using Eq. 3.10 and using the time-mean air temperature and relative humidity at the lowest model level from the control simulation.

In Fig. 3.10 we plot the baseline distribution of h_{diff} in our control run (panel a), along with the sensitivity of h_{diff} to warming (e.g. the change which would be associated with a uniform 1K increase in low-level temperature) (panel b). The control simulation has positive values of h_{diff} in regions of climatological ascent, as mentioned earlier, with more negative values indicating a strong inversion (e.g., over the Peruvian coast). Note that in our linear framework, the decrease in h_{diff} associated with boundary layer cooling is just the negative of what we have plotted in Fig. 3.10b.

The theoretical change in h_{diff} associated purely with changes in boundary-layer temperature is highest over regions with the highest climatological temperature and humidity (e.g., over the Western tropical Pacific or Central and Southern America). The magnitude of this change in h_{diff} with boundary-layer temperature changes is comparable to the magnitude of h_{diff} in the control run over the central Pacific, which again demonstrates that the additional ‘kick’ from the warming effect of absorbing aerosol is able to push regions which were negatively buoyant ($h_{\text{diff}} < 0$) in the control run into regions which are positively buoyant ($h_{\text{diff}} > 0$) following the perturbation.

Previously we also saw a local decrease in precipitation in response to absorbing aerosol over South America (Fig. 3.7), and using the h_{diff} framework we can understand this in terms of the surface and boundary-layer cooling by the aerosol when placed over land (Fig. 3.11). While this region of South America is unstable to deep convection in the control run (i.e., it has $h_{\text{diff}} > 0$ in Fig. 3.10a), the large moisture content of this region also makes it highly susceptible to decreases in h_{diff} from the boundary-layer cooling (which is strong over land). This means that for sufficient boundary-layer cooling (driven by aerosol scattering and absorption) h_{diff} will eventually become negative, suppressing deep convection and causing a decrease in precipitation locally.

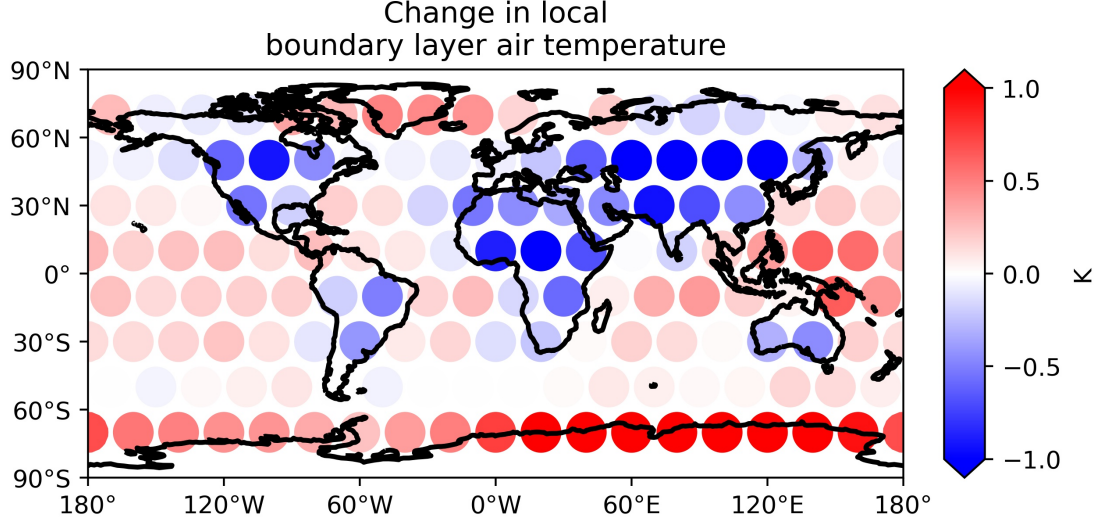


Figure 3.11: Local- and time-mean changes in the average temperature boundary layer (defined as the temperature of the lowest model level) for each of the 144 plume experiments. “Local” is defined as an average over the region containing 95% of the AOD perturbation. Each colored circle represents the results from the simulation imposing an aerosol perturbation in that location. As before, black circles are not added as all changes are significant.

While this framework allows for some basic insights into the response of the tropical atmosphere to absorbing aerosol, it does not account for changes in relative humidity or changes in circulation and moisture advection which would also affect the value of h_{diff} in addition to just the changes which comes from increased low-level temperature. This, along with a more thorough investigation of the response over land and the sensitivity to different land surface properties would be a fruitful avenue for future work.

3.4.1 Relationship with the local atmospheric energy budget

Previous work (Dagan et al., 2019, 2021) has explained the contrasting local precipitation responses to tropical and extratropical absorbing aerosol plumes using a version of the atmospheric energy budget which also accounts for the ability of the atmosphere to converge/diverge dry static energy on local scales, even though this term vanishes on global scales and thus does not appear in Eq. 3.2. In

these studies, they explain that the contrasting response of local precipitation (a local increase in response to tropical aerosol, and a local decrease in response to extratropical aerosol) arises due to the contrasting ability of the atmosphere to diverge excess dry static energy in the two regions. They argue that aerosol heating in the tropics is able to efficiently diverge dry static energy by generating large-scale, thermally driven overturning circulations. Because of the strong increase in dry static energy with height, even if the magnitude of upper-level divergence is equal and opposite in magnitude to the low-level divergence, such circulations represent a net energy export from the perturbed region. However, as we have shown, this explanation assumes that all regions are susceptible to deep, thermally-driven overturning circulations, whereas in the real world zonal-asymmetries (such as the Walker circulation) inhibit the formation of such circulations in regions such as the South-Eastern part of the tropical Pacific (Fig. 3.7). Hence, it is important to consider the longitudinal location of the absorbing aerosol when interpreting local precipitation changes, and particularly the aerosol location relative to the existing structure of tropical overturning.

The work of Dagan et al., 2019 also explains the local decrease in response to extratropical aerosol (reproduced in our experiments, Fig. 3.7) by arguing that the ability of the extratropical atmosphere to diverge dry static energy is weaker in the tropics due to the presence of the Coriolis force. They emphasize the process of geostrophic adjustment (Vallis, 2017), which is argued to ‘confine’ the aerosol heating to local scales. However, while geostrophic adjustment is a ubiquitous aspect of rotating fluids, it is only able to explain why the atmosphere remains in a state of geostrophic *balance*, and it is not immediately clear that this should result in a localization of the heating anomaly and decrease in precipitation. Here, we have used the argument put forward by (Williams et al., 2022a) to link the decrease in precipitation in response to extratropical aerosol to the propensity of the atmosphere to balance heating by horizontal advection when presented with

strong temperature gradients (as exist in the extratropics).

In summary, our experiments build upon the results of previous work (Dagan et al., 2019, 2021) done with the same ICON model, and we provide an alternative, dynamical perspective to explain these results which is complimentary to the local energetic perspective.

3.5 Conclusions

Understanding the response of precipitation to anthropogenic changes in atmospheric composition is essential in order to know how to adapt to the changes which will occur over the coming decades. However, understanding the response of precipitation to changes in absorbing aerosol presents a unique challenge due to the highly inhomogeneous spatial pattern of anthropogenic aerosol (which contrasts starkly with well-mixed greenhouse gases such as CO₂), and the complex tropospheric adjustments which occur in response to the diabatic heating from absorbing aerosol. In addition, precipitation changes are commonly separated into ‘fast’ and ‘slow’ components, differentiated by the timescales involved and whether or not they are mediated by SST changes. Previous work has demonstrated that a significant portion of the inter-model spread of precipitation changes in response to black carbon perturbations arises in the fast response (Samset et al., 2016), due to the rapid tropospheric adjustments of clouds and circulation to aerosol heating. However, the complexity of the models used for these studies (where aerosol burden and/or radiative properties are emergent properties rather than prescribed), makes it difficult to understand the mechanisms by which localized absorbing aerosol causes these rapid tropospheric adjustments.

To study the mechanisms by which localized absorbing aerosol in different geographical locations impacts on these rapid-adjustments and henceforth to precipitation, we have taken an idealized modelling approach. We have conducted

an ensemble of 144 atmosphere-only runs with a state-of-the-art GCM, forced by identical perturbations of aerosol optical depth and radiative properties placed systematically across the whole globe. This idealized approach allows us to explore the basic physics controlling the regional sensitivity of fast precipitation changes to absorbing aerosol.

On global-mean scales, we have found that absorbing aerosol tends to cause a decrease in precipitation in agreement with previous studies (Pendergrass and Hartmann, 2012; Samset et al., 2016) and physical reasoning (Pendergrass and Hartmann, 2014). However, despite prescribing identical aerosol burdens and radiative properties in each of our experiments, the global-mean precipitation decrease varies by over an order of magnitude between experiments. We explained this by using a global-mean energetic perspective and have linked the spread in global-mean precipitation response to; latitudinal variations in incoming solar radiation (Fig. 3.2), larger surface cooling from the aerosol over land than ocean (Fig. 3.1c), and large-scale temperature and circulation responses to absorbing aerosol over the Western tropical Pacific (Fig. 3.1b and Fig. 3.3).

The most surprising result from this part of our study is that the global-mean response to absorbing aerosol over the Western tropical Pacific is negligible (Fig. 3.1a), due mostly to the increased longwave cooling of the atmosphere which occurs in response to the larger global-mean temperature changes resulting from planetary waves propagating to the extratropics. We find that the muted global-mean precipitation response is due to increases in longwave cooling, which offset the increased shortwave absorption from the absorbing aerosol. This result has parallels with the earlier work of (Barsugli and Sardeshmukh, 2002), where they found that SST perturbations in the Indo-Pacific led to weaker changes in global-mean precipitation compared to perturbations in the Central Pacific. We hypothesize this is due to the enhanced ability of SST anomalies in this region to generate global mean temperature changes and drive compensating longwave cooling (Barsugli

et al., 2006). The first-order importance of circulation adjustments in response to absorbing aerosol over the Western tropical Pacific is similar to recent work focusing on the sensitivity of global mean effective radiative forcing (ERF) to absorbing aerosol (Williams et al., 2022a), who showed the circulation adjustments cause this region to be a ‘hot spot’ of ERF from absorbing aerosol.

We have also investigated the changes in ‘local’ precipitation, which generally confirm the findings of previous work which have shown a strong contrast in the local response to absorbing aerosol between the tropics and the extratropics and a weaker response over land than ocean (Dagan et al., 2020, 2021). However, these studies have not systematically explored the sensitivity of these results to the longitudinal position of the aerosol, and our results in Fig. 3.7 show that the longitudinal position of the aerosol has a large effect on its local precipitation response. In the tropics, we have explained this by introducing a variable (h_{diff}) which uses the boundary-layer temperature and moisture fields to measure the instability of the atmosphere to deep convection. This shows that in regions which are already unstable to deep convection, a boundary-layer heating perturbation (such as that provided by absorbing aerosol over the fixed-SST oceans) can cause a strong enhancement of local precipitation. Similar results also hold for marginally stable regions such as the Central Pacific, and the decrease in boundary layer temperatures over tropical land also helps to explain the weaker precipitation response in these cases.

We have also provided a dynamical explanation for the local decreases in precipitation in the midlatitudes (Fig. 3.7) by considering the linear quasi-geostrophic response of the atmosphere to diabatic heating. This simple argument, first put forward by Hoskins and Karoly (1981), explains that the tendency of the midlatitude atmosphere is to balance localized heating by horizontal advection of cooler air from the poles via the generation of a downstream low pressure system. This draws down cool air with a low moisture content to the region of the aerosol perturbation,

suppressing the precipitation locally over this region. Future work could examine the sensitivity of this result to the vertical location of absorbing aerosol, as this highly local response may break down as the diabatic heating perturbation moves upwards and generates a stronger potential vorticity anomaly and potentially a more wave-like response (Lin et al., 2018; Shen and Ming, 2018). Repeating our analysis for specific seasons would also be an interesting avenue for future work, particularly when thinking about the local precipitation response.

Finally, as noted throughout the paper, we have only focused on the ‘fast’ precipitation response to absorbing aerosol, and have not included the influence of the aerosol on the SSTs directly (through altering the surface energy balance) or through shifting the large-scale patterns of wind stress or cross-equatorial energy balances (possibly inducing ITCZ shifts (Voigt et al., 2017)). In particular, using interactive SSTs may weaken the response of precipitation to absorbing aerosol over the Western tropical Pacific, as the SSTs cool in response to the aerosol plume. This would be a fruitful avenue for future work and it would be useful to include similarly idealized absorbing aerosol perturbations in a hierarchy of slab-ocean and coupled simulations, perhaps focused on only a few regions. Based on our current results, it is sensible to imagine that the local precipitation responses would be weaker if we allowed the aerosol to locally cool the underlying SSTs (for example, by using a slab ocean) because we see weaker responses of local precipitation to aerosol when the aerosol is located over land (which has an interactive temperature). However, it is not clear what role land-atmosphere interactions play in setting this response and thus more targeted experiments are needed to understand these dynamics and link them to future changes in the hydrological cycle.

My picture of the tropics is that of a circus tent (the temperature in the tropical upper troposphere) held up by poles (deep tropical convection) of different heights (the SST determining how high the pole is).

— Isaac Held, in a 2011 blog post

4

A new conceptual model for tropical SST pattern effect and its non-linearities

In Chapters 2 and 3, we have focused on understanding how the geographical location of absorbing aerosol modulates its climate response. We have focused on understanding how the global-mean ERF_{ARI} changes as a function of the location of a fixed AOD burden (Chapter 2), and how precipitation responds both on global- and local- scales to the location of these AOD burdens (Chapter 3). The reason for focusing on spatial patterns of absorbing aerosol was motivated largely by the fact that, due to their short lifetimes, the pattern of atmospheric aerosol loading is highly heterogeneous. However, there are other aspects of Earth’s climate which also exhibit rich spatial variations. In this Chapter, we focus on understanding the climate response to spatial patterns of *sea surface temperature (SST) anomalies*.

This chapter is based on a paper published in Geophysical Research Letters:

Williams, A. I. L., Jeevanjee, N. & Bloch-Johnson, J.: “Circus tents, convective thresholds and the non-linear climate response to tropical SSTs” *Geophysical Research Letters* 12, 735–742 (2023). <https://doi.org/10.1038/s41558-022-01415-4>

I designed the study together with the co-authors. I carried out the simulations and analysed the data with input from the co-authors. I wrote and revised the manuscript in response to reviewer comments, with contributions from the co-authors.

As highlighted in the Introduction, the spatial pattern of tropical SST anomalies has recently been established as a key player in Earth’s energy balance due to the so-called ‘SST pattern effect’ (Stevens et al., 2016). The SST pattern effect encapsulates the notion that anomalies in Earth’s top-of-atmosphere (TOA) energy balance depend on the *spatial pattern* of SST anomalies, due to the impact of altered SST patterns on low clouds in the tropics. To quantify the SST pattern effect, recent work has adopted a Green’s Function approach (Dong et al., 2019b; Zhang et al., 2023; Zhou et al., 2017b), which assumes that the TOA response to spatial patterns of SST anomalies can be *linearly* reconstructed from a set of isolated SST perturbation experiments in an atmosphere-only GCM.

This assumption of ‘linearity’ is quite counter-intuitive given the well-established ‘threshold’ behaviour of deep convection (Emanuel, 2007; Williams et al., 2009) and the observed asymmetry between the climate response to extremes of the ENSO cycle (Hoerling et al., 1997). As such, our goal in this chapter is to evaluate the linear assumptions made by previous studies, and to explain why non-linearities arise from the perspective of basic tropical dynamics.

We begin by reviewing the theoretical basis for the ‘threshold’ behaviour of deep convection, and then use atmosphere-only model experiments to show that the TOA response to isolated SST anomalies is in fact non-linear. To explain these findings we then introduce a simple conceptual picture of the tropical atmosphere’s response to SST perturbations, which we term the ‘circus tent’ model, which builds on previous work in the tropical dynamics community. We show that in regions of deep convection the change in TOA flux associated with a positive SST anomaly can be explained simply by the change in low-level moist static energy and examine to what extent these changes are predictable purely from the SST changes. Finally, we briefly investigate the implications of this model for the ‘additivity’ of tropical SST perturbations, and discuss the implications of this result for Green’s function

approaches.

4.1 Theory

To link the dynamics of deep convection to the free-tropospheric temperature profile, we will frequently make use of the moist static energy in this chapter, defined as:

$$h = c_p T + L_v q + g z, \quad (4.1)$$

where c_p is the heat capacity of dry air, T is temperature, g is gravitational acceleration, z is height, L_v is the latent heat of vaporization of water, and q is the water vapour specific humidity. The moist static energy is useful as it is approximately conserved under moist, adiabatic motion.

A central pillar of our understanding of the tropical atmosphere is the assumption of convective quasi-equilibrium (Betts, 1982; Emanuel, 2007; Raymond, 1995), which holds that deep convection relaxes the free-tropospheric temperature profile to a moist adiabat set by the properties of the subcloud layer. As a moist adiabat is associated with constant saturation moist static energy, this is another way of saying that in regions of deep convection the saturation moist static energy (that is, Eq. 1 but with q replaced by its value at saturation, q^*) of the free-troposphere, h_{FT}^* , should approximately equal the moist static energy of the subcloud layer, h_0 . Note also that because ascending parcels mix with relatively dry air from the environment through entrainment (Singh and O’Gorman, 2013) regions of convection tend to be associated with a h_0 slightly in excess of h_{FT}^* (e.g., Figure 4.5a), but this does not strongly affect our results.

As mentioned before, gravity waves are efficient at communicating this local temperature profile imposed by convection (or equivalently, h^* profile) across the tropics. This means that deep convection in one region can communicate high values

of h_{FT}^* across the tropical free-troposphere, establishing a ‘convective threshold’ which inhibits the formation of deep convection if the subcloud h_0 is not sufficient in these regions to overcome the imposed h_{FT}^* .

Using this observation, we follow previous work (e.g., Williams and Pierrehumbert, 2017) in defining a ‘convective instability index’, $h_0 - h_{500}^*$, (taking the 500hPa level to be representative of the free-troposphere). In regions unstable to deep convection, we expect this quantity to be positive, whereas in regions stable against deep convection (e.g., subsiding regions) we expect this to be negative and to act as a measure of the inversion strength (similar to Wood and Bretherton, 2006, but also accounting for moisture differences as in Koshiro et al., 2022).

A further simplifying assumption which is frequently made is that h_{500}^* is uniform across the free-troposphere (the ‘weak temperature gradient’ assumption, WTG), which allows one to replace h_{500}^* with its tropical average value (establishing a single value for the convective threshold). However, as we will show later, this assumption is misleading in the context of the pattern effect, where warming-induced changes in h_{500}^* are comparable to the spatial variations in baseline h_{500}^* arising from non-zero zonal temperature gradients (Bao et al., 2022; Bao and Stevens, 2021; Fueglistaler et al., 2009). This means that there are actually a spectrum of values for the convective threshold, depending on the local value of h_{500}^* , and to fully understand the pattern effect we cannot assume a single value for h_{500}^* .

4.2 Methods

We performed a series of atmosphere-only simulations using the ICON (ICOsahedral Non-hydrostatic) general circulation model. The model is run on a triangular grid, at R₂B₄ specification, corresponding to an approximately uniform grid-spacing of 160km on a Cartesian grid. The model uses a terrain-following vertical sigma-height grid with 47 levels between the surface and the model top at 83km. Radiation

is parameterized using PSrad scheme (Pincus and Stevens, 2013b), and other parameterizations include a bulk mass-flux convection scheme (Tiedtke, 1989b), a relative-humidity based cloud cover scheme (Sundqvist et al., 1989b) and a single-moment microphysics scheme (Baldauf et al., 2011). In our experiments all greenhouse gases and ozone are fixed at their 1979 levels (Williams et al., 2022a).

The control simulation was run for 20 years with a prescribed monthly climatology of SSTs and sea-ice concentrations derived from the Atmospheric Model Intercomparison Project (AMIP) (Neale and Hoskins, 2000b) boundary conditions over 1979–2016. Then additional simulations were conducted for 10 years each with an additional ‘cosine patch’ SST perturbation (following Barsugli and Sardeshmukh, 2002) covering different locations throughout the tropics (perturbation centers correspond to the centers of each of the circles in Figure 4.2 and an example SST perturbation is shown in Figure 4.1). For all simulations, we discard the first year as spin-up and conduct analysis using time-averages over the rest of the simulation period.

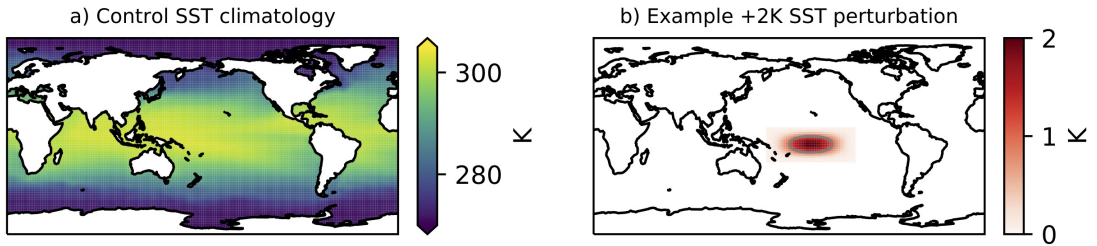


Figure 4.1: a) Annual-mean climatological SST in the control run. b) Representative example of a +2K Δ SST perturbation in the Central Pacific, with the half-amplitude (+1K) contour shown in grey.

The subcloud moist static energy (h_0) was calculated using values at the lowest model level, and the free-tropospheric saturation moist static energy was approximated by its value at 500hPa (h_{500}^*). Our conclusions are insensitive to the precise choice of levels, and similar results were obtained when either calculating the subcloud moist static energy as the average moist static energy over the lowest

1km (lowest 5 model levels) or calculating the saturation free-tropospheric moist static energy as a bulk average over 700hPa-300hPa. Also note that because h_0 is defined at a given height level, warming-induced changes in h_0 are only due to changes in temperature and humidity at that level.

We also define two averaging operators which make the presentation clearer. An overbar $\overline{(\cdot)}$ indicates an average over tropical latitudes ($\pm 30^\circ$), and angle brackets $\langle(\cdot)\rangle$ indicates a *local* ‘patch-average’ over the region covered by the SST patch perturbation in that experiment (i.e., over ocean grid-points with $|\Delta\text{SST}| > 0$). For example, $\langle\Delta h_0\rangle$ is the change in low-level moist static energy in a particular patch experiment, averaged over all *ocean* grid-points where $|\Delta\text{SST}| > 0$ (see Figure 4.1b for an example). Similarly, $\langle\Delta h_{500}^*\rangle$ is the change in saturation free-tropospheric moist static energy at 500hPa, averaged over the underlying grid-points which are ocean and have $|\Delta\text{SST}| > 0$. All spatial averages include area-weighting.

4.3 Results

4.3.1 Non-linear TOA response to tropical SSTs

In Figure 4.2 we plot the globally-averaged change in net TOA radiation (ΔR_{TOA}) for each of the tropical SST perturbation experiments, relative to the control. In agreement with previous work Figure 4.2a,c shows that SST warming in the Western tropical Pacific (a region of climatological deep convection), is associated with strong, negative changes in the global-mean R_{TOA} flux via the ‘non-local stability-inversion’ mechanism (Dong et al., 2019c; Zhou et al., 2017a). The results are similar if we focus on the SW (Figure 4.3). Additionally, in subsiding regions the ΔR_{TOA} is slightly positive for the +2K and +4K patches, indicating the ‘local stability-inversion’ mechanism is at work. However, if we compare these results to the cooling patch experiments in Figure 4.21b,d we can see there is a marked asymmetry in the ΔR_{TOA} response between the warming and cooling experiments (note that we have normalized all experiments by the magnitude and sign of the

SST perturbations, so if the system was linear all of the panels in Figure 4.2 would be identical). Notably, whereas warming in convective regions generates a strongly negative ΔR_{TOA} , cooling in convective regions has a comparatively small influence on the R_{TOA} (Figure 4.2b). Additionally, while the ΔR_{TOA} increases approximately linearly with SST warming (i.e. the values are similar in Figure 4.3a,c), for SST cooling in convective regions the R_{TOA} does not change linearly (i.e. the negative values over the West Pacific are smaller in Figure 4.2d than Figure 4.2c). The spatial patterns of ΔR_{TOA} are presented in Figure 4.7 for a representative patch in the West Pacific and confirm these findings.

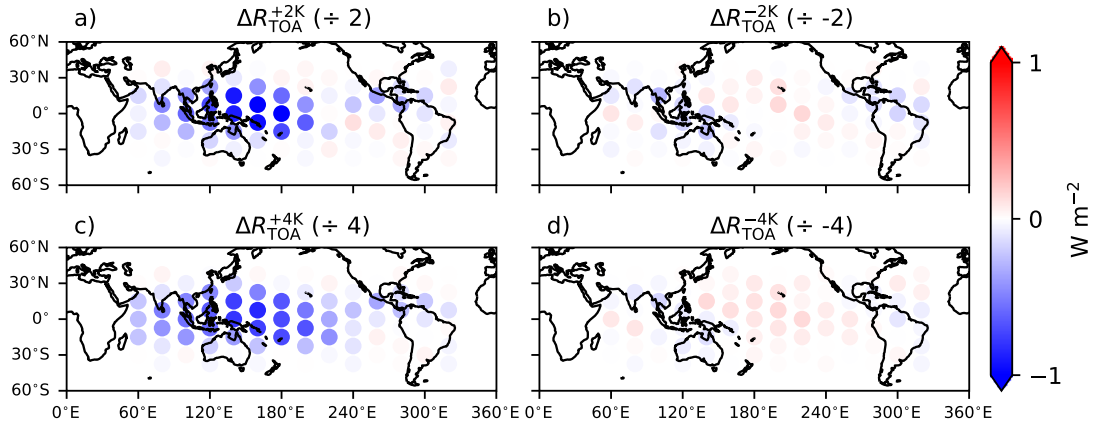


Figure 4.2: Change in global-mean TOA flux (ΔR_{TOA}) associated with each of the SST perturbation experiments. a) Changes for +2K patches, divided by 2. b) Changes for -2K patches, divided by -2. Panels c,d) as in a,b) but divided by ± 4 .

Alongside this ‘asymmetry’ in the ΔR_{TOA} between warming and cooling patches, there is also another, more subtle, non-linearity that appears in our experiments. To spot this, compare the ΔR_{TOA} responses in the Central Pacific between the +2K experiments in Figure 4.2a and +4K experiments in Figure 4.2c. Although the $\Delta R_{\text{TOA}}^{+4\text{K}}$ response is generally around twice the $\Delta R_{\text{TOA}}^{+2\text{K}}$ response (note that the values in Figure 4.2a,c are divided by 2 to make them comparable to the +2K experiments), in the weakly stable regions of the Central Pacific there are numerous experiments where the ΔR_{TOA} either changes sign or becomes much more negative at +4K compared to +2K. We term this the ‘magnitude-dependence’ of the SST

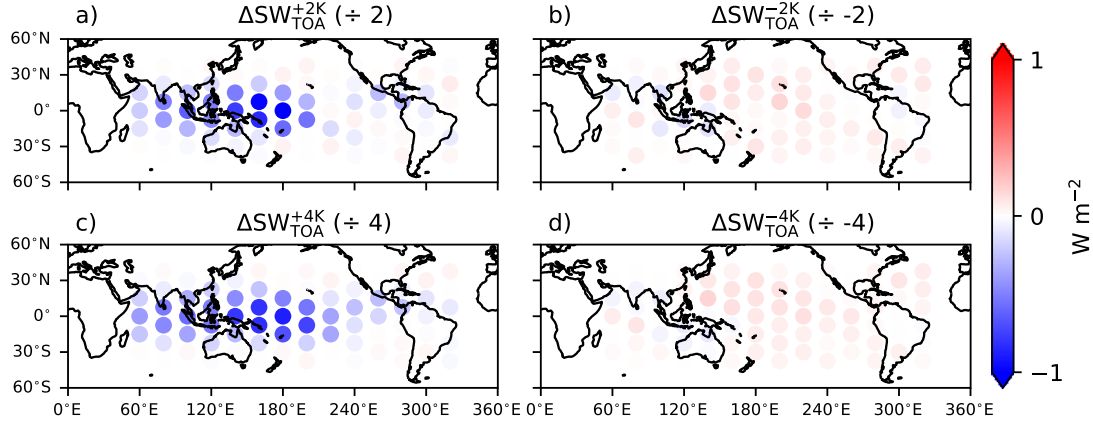


Figure 4.3: Change in global-mean SW TOA flux ($\Delta\text{SW}_{\text{TOA}}$) associated with each of the SST perturbation experiments. a) Changes for +2K patches, divided by 2. b) Changes for -2K patches, divided by -2. Panels c,d) as in a,b) but divided by ± 4 .

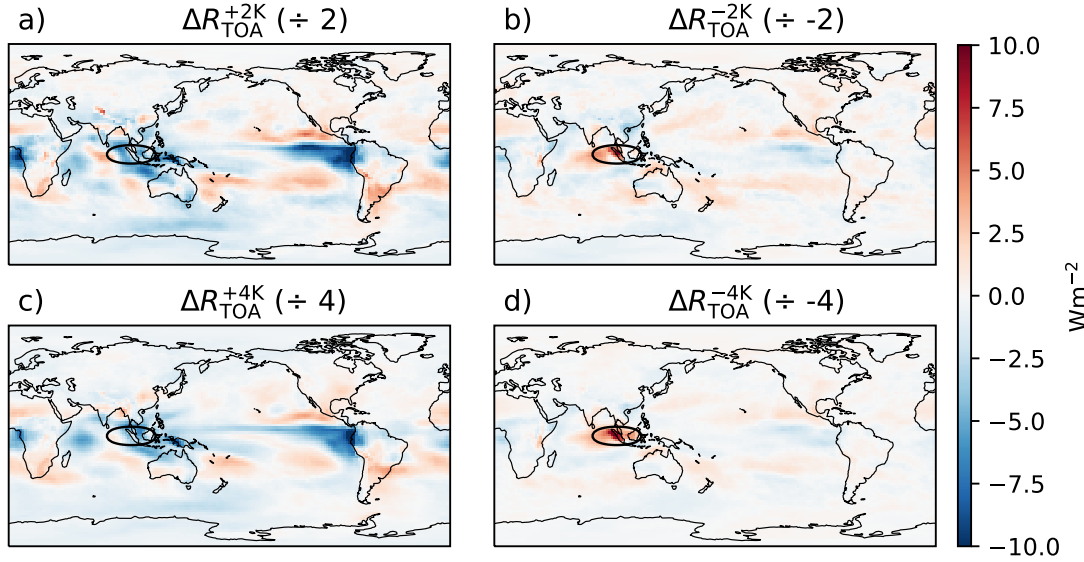


Figure 4.4: Spatial maps of the time-averaged ΔR_{TOA} for a patch in the Western Pacific warm pool (140E, 0N) for $\Delta\text{SST} = \pm 2\text{K}$ divided by ± 2 (a,b) and $\Delta\text{SST} = \pm 4\text{K}$ divided by ± 4 (c,d).

pattern effect, to convey that the strongly negative ΔR_{TOA} only kicks in for ΔSST above a certain magnitude in these moderately stable regions.

To dig into this more, we have performed additional simulations at $\pm 1\text{K}$ and $\pm 3\text{K}$ for a subset of four patches along the equatorial Pacific (Figure 4.5a). As indicated by the shading in the background of Figure 4.5a, these four patches cover

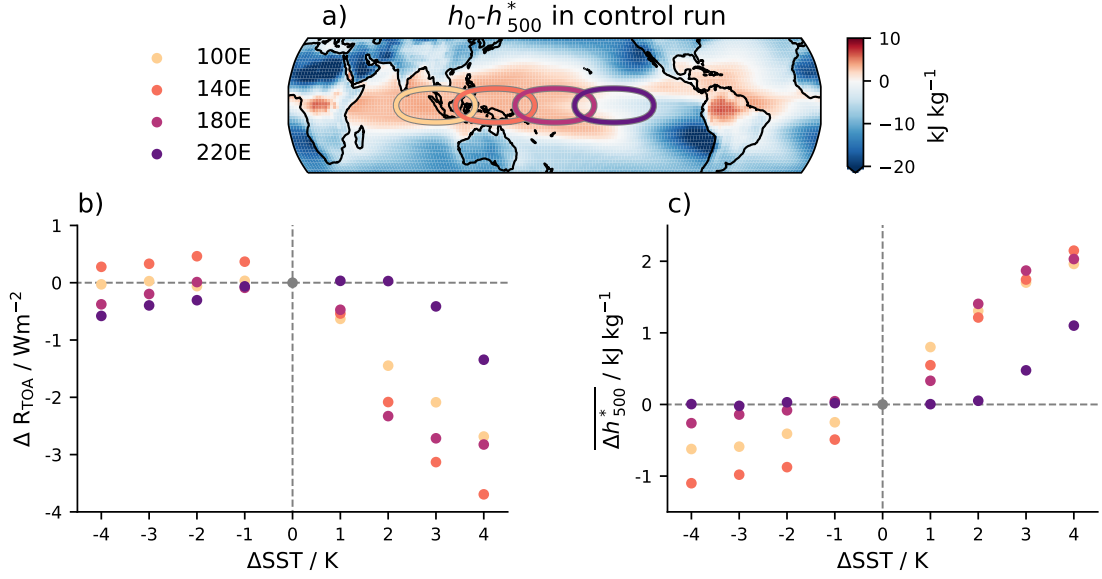


Figure 4.5: a) The half-amplitude contours of the four locations of the different equatorial sensitivity experiments are overlaid on top of the ‘convective instability’ parameter ($h_0 - h_{500}^*$) in the control run. b) The changes in global-mean R_{TOA} for each of the sensitivity experiments at $\pm 1\text{K}$, $\pm 2\text{K}$, $\pm 3\text{K}$, $\pm 4\text{K}$. c) as in b) but showing the change in tropical-average h_{500}^* , denoted $\overline{\Delta h_{500}^*}$. To account for the varying amounts of land in different patches, in panels b and c the ΔR_{TOA} and $\overline{\Delta h_{500}^*}$ have been divided by the ΔSST weighted land-fraction before plotting.

distinct convective regimes (as $h_0 - h_{500}^*$ is a measure of convective instability, see Theory). The patches at 100E and 140E are in strongly convective regions (where $h_0 - h_{500}^* > 0$), whereas the patch at 220E is in a region which is stable to deep convection (where $h_0 - h_{500}^* < 0$) and the patch at 180E is in a transition region. In Figure 4.5b we have plotted the global-mean ΔR_{TOA} for each of the four patches at $\Delta \text{SST} = \pm 1\text{K}$, $\pm 2\text{K}$, $\pm 3\text{K}$, $\pm 4\text{K}$. Again, the asymmetry in the TOA response between positive and negative perturbations is evident. Taking the 140E patch as an example, positive ΔSST perturbations induce a negative ΔR_{TOA} which scales quasi-linearly with the ΔSST magnitude, but for negative ΔSST the TOA response quickly saturates. It is a similar story for the 100E patch, but the signal is weaker for negative ΔSST . However, if we subset the TOA response only over subsiding regions we recover the same behaviour (Figure 4.6).

This asymmetry is also evident in how the ΔSST perturbations impact on the

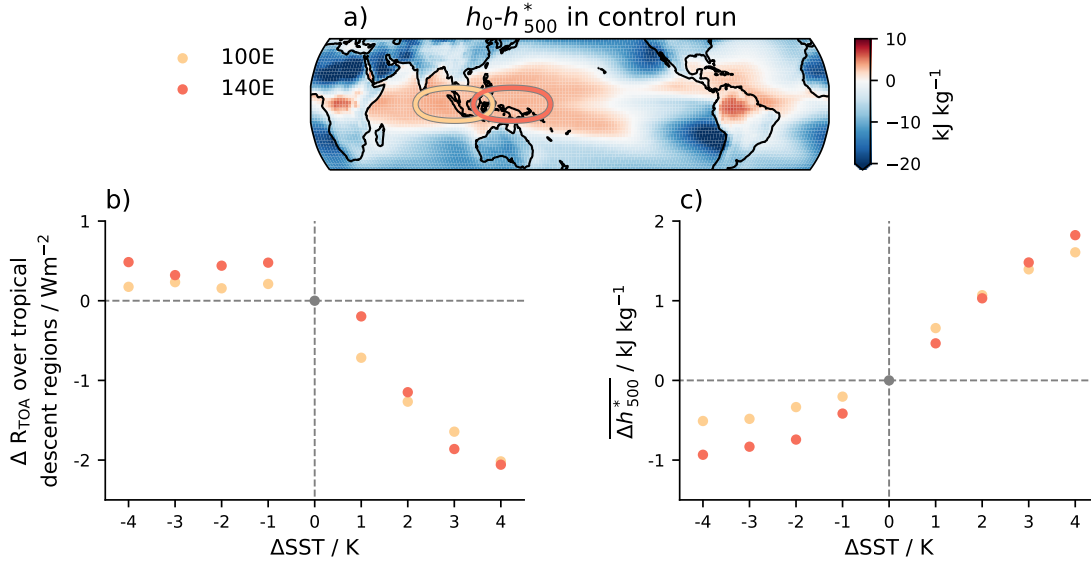


Figure 4.6: As in Figure 4.5 of the main text, but in panel b we plot the change in R_{TOA} averaged over tropical regions where $\omega_{500} > 0$ in the control run (to pick out low cloud subsidence regions). We also plot the two patches in deeply convective regions. This figure illustrates how the ΔR_{TOA} response to negative ΔSST anomalies in convective regions is linear over a very small region, but quickly saturates.

tropical-average saturation moist static energy, $\overline{h_{500}^*}$ (Figure 4.5c), which makes sense as a higher $\overline{h_{500}^*}$ indicates a warmer and more stable free-troposphere which tends to increase the inversion over low-cloud regions. Looking again at the 140E patch confirms our suspicion that greater $\overline{\Delta h_{500}^*}$ is associated with greater ΔR_{TOA} for positive values of ΔSST . Conversely, for negative ΔSST there is an initial decrease in $\overline{h_{500}^*}$, which quickly saturates. For the other ‘convective’ patch at 100E it is a similar story. It is also worth noting here that in non-convective regions, negative ΔSST also causes a negative ΔR_{TOA} , which strengthens with larger ΔSST (Figure 4.5b), even without impacting on the $\overline{h_{500}^*}$ (Figure 4.5c).

Previously we also noted a ‘magnitude dependence’ of ΔR_{TOA} on the magnitude of positive ΔSST changes in moderately stable regions (Figure 4.2a,c) and this is again present in Figure 4.5. Looking at the 220E patch results in Figure 4.5b, we can see that the sharp decrease in R_{TOA} does not occur until $\Delta \text{SST} \gtrsim 2\text{K}$, whereas for the other patches the quasi-linear decrease occurs for all $\Delta \text{SST} > 0\text{K}$. A similar picture exists for the $\overline{\Delta h_{500}^*}$, which is unaffected until $\Delta \text{SST} \gtrsim 2\text{K}$ for the 220E

patch (Figure 4.5c).

4.3.2 A conceptual model for the non-linear response

To understand this behaviour, it is helpful to introduce a conceptual picture which builds on our earlier discussion of convective quasi-equilibrium and weak (but non-zero) temperature gradients in the free-troposphere. We call this the ‘circus tent’ model and it is sketched in Figure 4.8a. This conceptual model was originally suggested by Isaac Held in a 2011 blog post entitled ‘Atlantic Hurricanes and Differential Tropical Warming’. In this model, the temperature of the tropical free-troposphere (or equivalently, its h^*) can be thought of as being a tight fabric supported by ‘convective tent poles’ of different ‘heights’ corresponding to their subcloud h_0 . Where there is deep convection, convective quasi-equilibrium ensures the height of the fabric (h_{500}^*) is roughly equal to h_0 , and as one moves away from the convective center the h_{500}^* profile relaxes somewhat until it comes under the influence of a different tent pole. The ‘tightness’ of the fabric is related to the efficient homogenization of h^* anomalies by gravity waves, and the pattern of Δh_{500}^* appears to be related to the Matsuno-Gill response to tropical heating (e.g., Figure 1 of Gill, 1980), with an equatorially-confined lobe extending to the East and two off-equatorial maxima slightly to the West of the heating. This pattern can be seen in our maps of Δh_{500}^* (Figure 4.7).

At this point it is worth briefly discussing the relative roles of gravity waves and large-scale equatorial waves (i.e. the Matsuno-Gill response) in homogenizing temperatures across the tropics. Typically, studies focus on the role of gravity waves in setting the weak temperature gradients in the tropical free-troposphere. However, it is likely that both gravity waves and the Matsuno-Gill response are acting to homogenize the free-tropospheric temperatures. Bao et al. (2022) recently investigated the mechanisms behind free-tropospheric temperature gradients in reanalysis data, and found that the associated hydrostatic pressure gradients

were balanced predominantly by nonlinear terms in the momentum equation, in particular via zonal wind advection. They further highlighted that it is the rotational component of the flow which is dominant, suggesting a role for equatorial waves (Sardeshmukh and Hoskins, 1988). Overall, this suggests that the Matsuno-Gill response and gravity waves both play a large role in setting the homogeneity of free-tropospheric temperatures in the tropics. At typical GCM resolution, gravity waves are not explicitly resolved and so need to be parameterized, lest the GCM miss a crucial form of vertical momentum transfer (Giorgetta et al., 2018). We have discussed in the introduction the uncertainty associated with subgrid-scale parameterizations, but it is worth noting here that the degree to which the free-tropospheric temperature gradients are determined by large-scale waves versus gravity wave parameterization will likely depend sensitively on the parameterization scheme used.

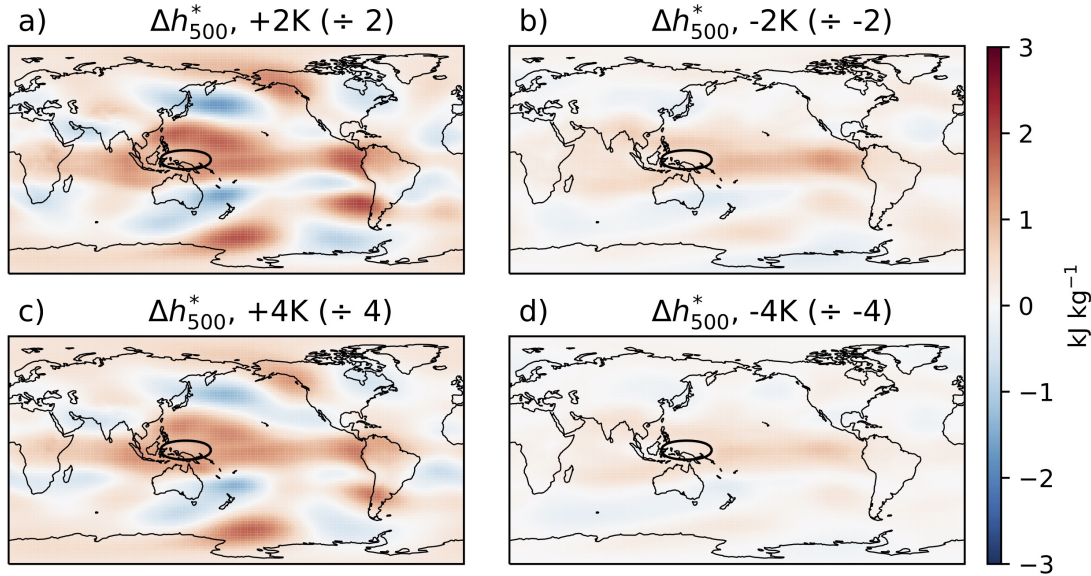


Figure 4.7: Spatial maps of the time-averaged Δh_{500}^* for a patch in the Western Pacific warm pool (140E, 0N) for $\Delta SST = \pm 2K$, divided by ± 2 (a,b) and $\Delta SST = \pm 4K$ (c,d) divided by ± 4 .

Using this model we can now understand the impact of warming and cooling in convective regions we saw in Figure 4.5. The case of positive ΔSST in convective

4. A new conceptual model for tropical SST pattern effect and its non-linearity 185

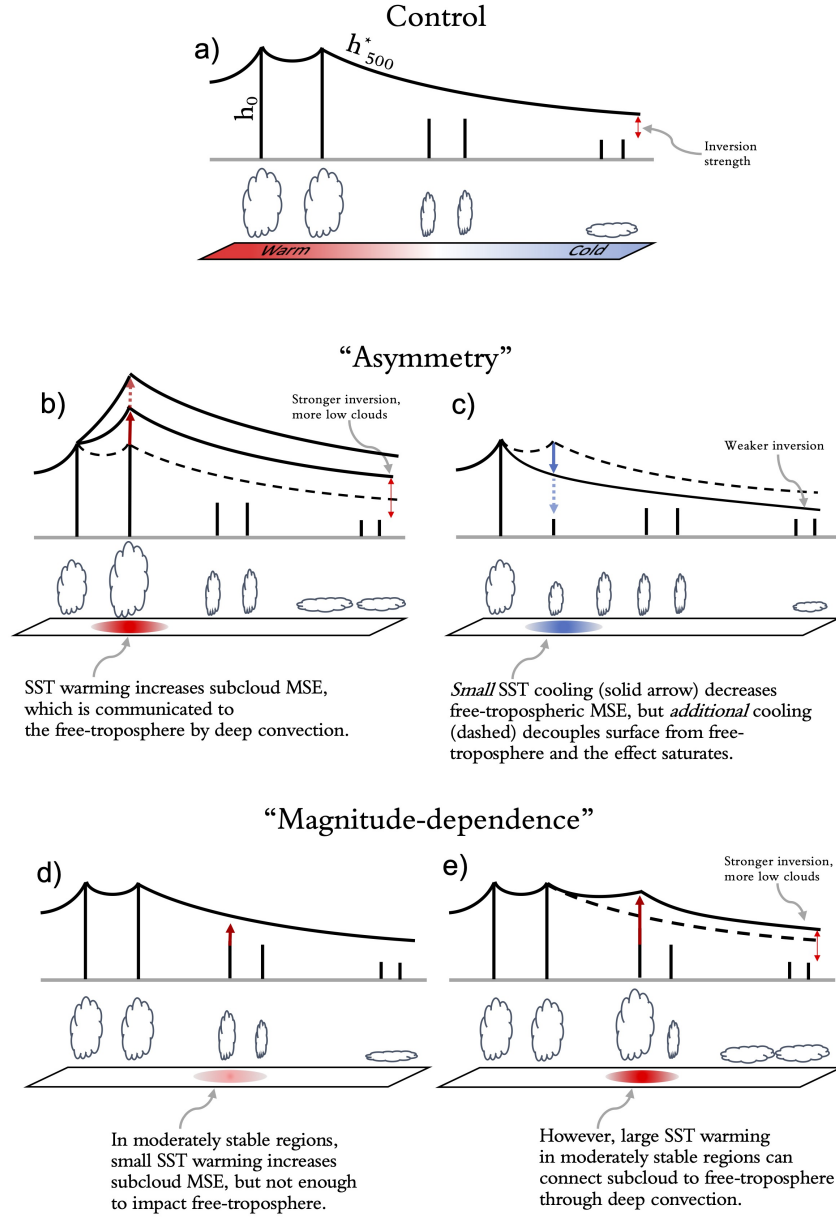


Figure 4.8: Schematic representation of the processes controlling the non-linear TOA response to SST perturbations in different convective regimes. a) Illustrates the ‘circus tent’ model of the tropical atmosphere in the presence of an SST gradient, where deep convection occurs over the warmest SSTs with highest h_0 and is able to perturb the free-tropospheric temperature structure (as measured by h_{500}^*). b) and c) conceptually illustrate how a tropical circus tent responds ‘asymmetrically’ to warming and cooling in convective regions. d) and e) illustrate how a tropical circus tent responds non-linearly with respect to the magnitude of the SST perturbation in moderately stable regions.

regions is sketched in Figure 4.8b. In this case, because the region is already convecting, we are increasing the height of a tent pole which is already ‘in contact’

with the tent fabric, which raises the h_{500}^* throughout the free-troposphere, including over regions of low-clouds where it strengthens the inversion. Because the tent fabric is tight, this also explains why the changes in h_{500}^* and R_{TOA} are approximately linear (Figure 4.8b). On the other hand, Figure 4.8c illustrates how for negative ΔSST in convective regions the tent pole may be lowered sufficiently to lose contact with the fabric. At this point, further decreases in SST are not communicated to the free-troposphere, which explains the ‘saturation’ of $\overline{\Delta h_{500}^*}$ and ΔR_{TOA} at negative ΔSST (Figure 4.5b, Figure 4.7). Additionally, for non-convective regions we previously noted that negative ΔSST causes a negative ΔR_{TOA} even without changing the tropical-average h_{500}^* . To understand this, we note that non-convective regions do not ‘make contact’ with the tent fabric, and that the difference between the height of the tent pole and the overlying fabric in this case ($h_0 - h_{500}^*$) is a measure of inversion strength. Hence, negative ΔSST in non-convective regions perturbs R_{TOA} not by altering h_{500}^* , but by *lowering* the local values of h_0 through cooling the surface. This increases the inversion strength over this region and locally increases low-cloud coverage, leading to a negative ΔR_{TOA} .

The circus tent model of the tropics is also useful for understanding the ‘magnitude dependence’ we have noted earlier, where even for positive ΔSST , the relationship between ΔSST and $\overline{\Delta h_{500}^*}$ or ΔR_{TOA} can be highly non-linear. This phenomenon is most pronounced in moderately stable regions such as the Central Pacific, which correspond to the middle tent poles in Figure 4.8a which are not quite tall enough to make contact with the h_{500}^* fabric (i.e., they sit below the local convective threshold). In this situation, small SST warming raises the subcloud h_0 , but may not raise it sufficiently to overcome the convective threshold and make contact with the tent fabric (Figure 4.8d). For the SST warming to be able to substantially alter the $\overline{\Delta h_{500}^*}$ or ΔR_{TOA} , it must be strong enough to raise this tent pole (increase the h_0) enough to make contact with and subsequently raise the height of the tent fabric, as in Figure 4.8e. When this condition is met, the local increase in h_{500}^* results in a stronger inversion over low-cloud regions due to the

tightness of the fabric. This explains the sudden decrease in ΔR_{TOA} for the 220E patch at $\Delta \text{SST} > 2\text{K}$.

4.3.3 A linear model for the TOA response to SST warming in convective regions

Given the non-linearities we have highlighted in the previous section, a natural question is: “Why do previous Green’s function methods work at all?”. One possibility is that the regions of strongest sensitivity in Green’s function studies tend to be strongly convective, such as the West Pacific (Dong et al., 2019c). Our own analysis shows that the relationship between ΔR_{TOA} and ΔSST is reasonably linear in convective regions (e.g., compare Figure 4.5b and Figure 4.5c), and in this section we explore this link in more detail using the +2K and +4K experiments from Figure 4.2. To do this, in Figure 4.9a we first plot the change in patch-averaged h_0 against the patch-averaged changes in h_{500}^* for each of +2K and +4K experiments from Figure 4.2 which are deemed to be ‘convective’ (i.e., if $\langle h_0 \rangle > \langle h_{500}^* \rangle$ either in the control run or in the perturbed run). The results of Figure 4.9a act as a test of the convective quasi-equilibrium hypothesis mentioned earlier, and confirms that changes in subcloud h_0 are efficiently transported into the local free-troposphere in regions of deep convection. Next, in Figure 4.9b we check to what extent these local changes in h_{500}^* relate to broader changes across the tropics. In the limit of zero horizontal temperature gradients (perfect WTG) the points in Figure 4.9b would lie on the one-to-one line (because any localized variations in h_{500}^* would be communicated perfectly across the tropics), but we actually find that they lie on a line of constant, but shallower, slope (gradient of about 1/3). The shallow slope indicates that the changes in h_{500}^* are not spread uniformly across the tropics (motivating our conceptual model which includes horizontal h^* gradients), and the fact that the slope is constant with increasing local forcing indicates that there is no ‘state-dependence’ in the relationship between local and tropical-average changes in h_{500}^* . To confirm the role the changes in h_{500}^* play in the ‘non-local stability-inversion’

mechanism, in Figure 4.9c we scatter the changes in $\overline{\Delta h_{500}^*}$ against ΔR_{TOA} and the strong linear relationship indicates that larger changes in free-tropospheric temperature do indeed alter the inversion strength and TOA radiation. Taken together, these results suggest that ΔR_{TOA} should be linearly related to the $\langle \Delta h_0 \rangle$ in regions of deep convection, which we find holds reasonably well in our experiments (Figure 4.9d).

Since local $\langle \Delta h_0 \rangle$ accounts for much of the scatter in ΔR_{TOA} for positive ΔSST in convecting regions (Figure 4.9d), a natural question is: can we relate $\langle \Delta h_0 \rangle$ to the ΔSST perturbation more directly? As we defined the subcloud moist static energy at a given geopotential height, we can write $\langle \Delta h_0 \rangle$ as:

$$\langle \Delta h_0 \rangle = \langle c_p \Delta T_0 + L_v \Delta q_0 \rangle.$$

Writing $q_0 = \text{RH}_0 q_0^*$ and linearizing, we can further approximate this as:

$$\langle \Delta h_0 \rangle \approx \left\langle \left(c_p + L_v \text{RH}_0 \frac{dq_0^*}{dT} \right) \Delta T_0 + L_v q_0^* \Delta \text{RH}_0 \right\rangle.$$

If we also assume Clausius-Clapeyron scaling of q_0^* (i.e., $\frac{dq_0^*}{dT} = \frac{L_v q_0^*}{R_v T^2}$), we arrive at:

$$\langle \Delta h_0 \rangle \approx \left\langle \left(c_p + \frac{\text{RH}_0 L_v^2}{R_v T_0^2} q_0^* \right) \Delta T_0 + L_v q_0^* \Delta \text{RH}_0 \right\rangle. \quad (4.2)$$

In Fig. 4.10a, we show that Eq. 4.2 can capture most of the variations in $\langle \Delta h_0 \rangle$ across the SST patch experiments ($\pm 2\text{K}$ and $\pm 4\text{K}$) conducted for Figure 4.2, although there are slight errors at large, positive ΔSST which we think are because we only use a first-order Taylor expansion of $\frac{dq_0^*}{dT}$ in deriving Eq. 4.2. Next, if we assume the surface temperature is perfectly communicated across the sub-cloud layer, we can replace $T_0 = \text{SST}$, which also reproduces the modeled $\langle \Delta h_0 \rangle$ well (Fig. 4.10b), with a slight overestimate at higher $\langle \Delta h_0 \rangle$. We get similar levels of skill if we assume constant relative humidity but include ΔT_0 (Fig. 4.10c). This purely ‘thermodynamic’ scaling also becomes less accurate at strong, positive ΔSST , possibly because of changes in moisture convergence. Finally, if we combine both of

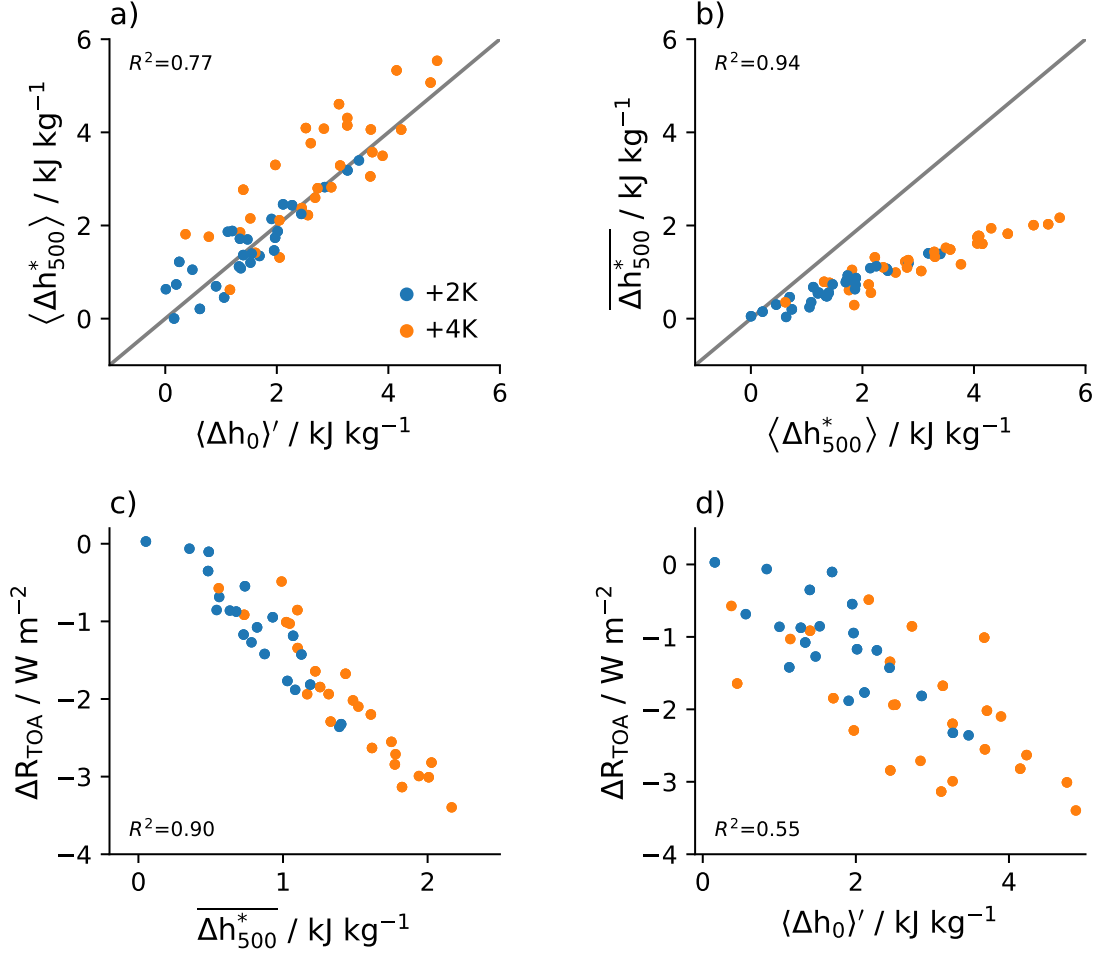


Figure 4.9: Evaluating the canonical model of the SST pattern effect for warming in convective regions. a) shows a scatter plot of $\langle \Delta h_0 \rangle'$ vs $\langle \Delta h_{500}^* \rangle$, as a test of convective quasi-equilibrium. $\langle \Delta h_0 \rangle'$ is equal to $\langle \Delta h_0 \rangle$ except for if that local patch region does not ‘convect’ in the control climate (i.e., $\langle h_0 \rangle < \langle h_{500}^* \rangle$), in which case we subtract the absolute value of $\langle h_0 - h_{500}^* \rangle$ from the control run to adjust for the fact that the tent fabric only ‘feels’ the amount by which the convective threshold is exceeded. b) shows a scatter plot of $\langle \Delta h_{500}^* \rangle$ vs $\overline{\Delta h_{500}^*}$, as a test of WTG. c) shows $\overline{\Delta h_{500}^*}$ vs the global-mean ΔR_{TOA} , with the strong correlation indicating support for the ‘non-local stability-inversion’ mechanism. Finally, d) shows a scatter plot of the global-mean ΔR_{TOA} vs $\langle \Delta h_0 \rangle'$. Numbers indicate the Pearson coefficient of determination, R^2 .

these approximations ($\Delta \text{RH} = 0$ and $T_0 = \text{SST}$ in Eq. 4.2) we can still reconstruct changes in subcloud h_0 reasonably well (Fig. 4.10d), even though we are only using information about the SST distribution. This provides support for work which has used SST as a proxy for changes in h_0 , however for +4K perturbations we only achieve $R^2 = 0.49$ in this case, which suggests caution should be taken when using SST as a proxy for subcloud MSE at large, positive ΔSST .

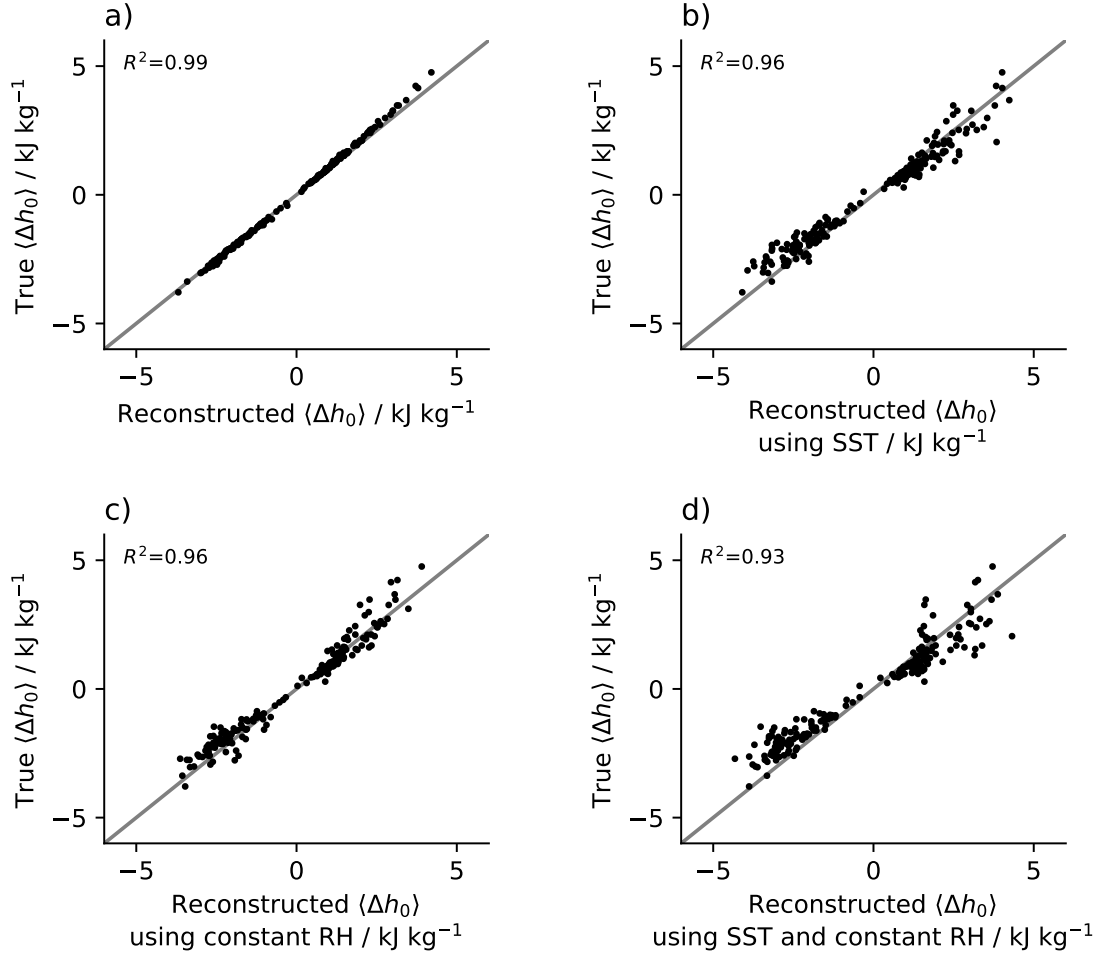


Figure 4.10: Estimating the true $\langle \Delta h_0 \rangle$ in each of the $\pm 2\text{K}$ and $\pm 4\text{K}$ SST patch experiments from Figure 4.2: a) Eq. 4.2. b) Eq. 4.2 with ΔT_0 replaced by ΔSST . c) Eq. 4.2 with $\Delta \text{RH}=0$. d) Eq. 4.2 after making both approximations. Numbers indicate the Pearson coefficient of determination, R^2 .

4.4 Discussion and Conclusions

In this work we have shown that the climate response to isolated tropical SST perturbations exhibits strong non-linearities with respect to their sign, magnitude and location. We argue that these non-linearities arise primarily due to the fact that identical SST perturbations do not necessarily perturb the saturation moist static energy of the tropical free-troposphere equally, which is important for setting the inversion strength over low-cloud regions. For example, in moderately stable

regions such as the Central Pacific negative SST anomalies have little impact on the global-mean TOA radiation or tropical h_{500}^* , and positive SST anomalies only have a strong effect when the ΔSST magnitude exceeds a certain value determined by the local convective threshold.

To understand these results, we have introduced the ‘circus tent’ model of the tropical atmosphere, which brings together the twin pillars of convective quasi-equilibrium and weak (but non-zero) temperature gradients in the tropical free-troposphere. In this model, local SST perturbations only alter the h_{500}^* if the subcloud h_0 exceeds a local ‘convective threshold’, and then proceed to perturb the h_{500}^* quasi-linearly for positive perturbations. Negative ΔSST perturbations in convective regions can also decrease the h_{500}^* (Figure 4.5c), generating positive ΔR_{TOA} as a result of low-cloud changes. However, the effect saturates for sufficiently negative ΔSST because eventually the subcloud layer becomes decoupled from the free-troposphere (Figure 4.5b, Figure 4.8c). These concepts are understood implicitly in the tropical dynamics community (Flannaghan et al., 2014; Fueglistaler et al., 2009; Fueglistaler et al., 2015; Zhao et al., 2009), but to our knowledge this is the first time they have been *explicitly* invoked to understand the pattern effect.

Our work has implications for studies which construct SST Green’s functions by demonstrating that the TOA response is not always linear in ΔSST , even for a given sign, and that the character of the non-linearity varies depending on the convective regime being perturbed. Preliminary work as part of the Green’s Function Model Intercomparison Project (GFMIP, Bloch-Johnson et al., 2023) has also demonstrated similar non-linearities in five other GCMs (Zhang et al., 2023), suggesting our results are not model-specific. This does not mean the Green’s function approach is without merit, but suggests that future work should focus on mapping the TOA response across multiple ΔSST values for each location and understanding the responses in isolation before combining them so as to minimize the risk of introducing compensating errors. This is currently being undertaken in a

multi-model context as part of the GFMIP project (Bloch-Johnson et al., 2023). A particular focus of future work should be on understanding how SST perturbations alter the distribution of subcloud moist static energy, particular over the perturbed region, and understanding what factors set the shape of the tropical ‘circus tent’ and its response to forcing.

Finally, although our work has focused on the climate response to *isolated* SST perturbations, the ‘circus tent’ framework we propose also predicts that the response to SST perturbations need not *combine* linearly either (see schematic in Figure 4.11), which is another key assumption of Green’s function reconstructions (e.g., Zhang et al., 2023, Dong et al., 2019c).

While Dong et al., (2019) showed that the TOA responses combined linearly when simultaneously warming the Eastern and Western tropical Pacific, our conceptual model suggests this is a fortuitous outcome of warming one region already ‘in contact’ with the tent fabric, and another which sits well-below it (Figure 4.8a, Figure 4.11). Generally our results suggest that positive SST patches in convective or marginally stable regions have a weaker impact on free-tropospheric h^* when applied simultaneously than when considering a linear sum of the individual patches. In other words, the impact of positive SST anomalies is “less than or equal to the sum of its parts”. On the other hand, our framework predicts that simultaneously *cooling* multiple convective regions yields a response which is “greater than or equal to the sum of its parts”.

To assess whether the “less than the sum of its parts” paradigm is applicable to our ICON GCM simulation, we also conducted a series of “two-patch” experiments where we simultaneously perturbed pairs of the four equatorial patches which we highlighted in Figure 4.5. We then calculated the change in free-tropospheric h^* for all of the experiments, and compared them against the changes one would infer if the system was indeed perfectly linear, with the results plotted in Figure 4.12.

“Non-additivity”

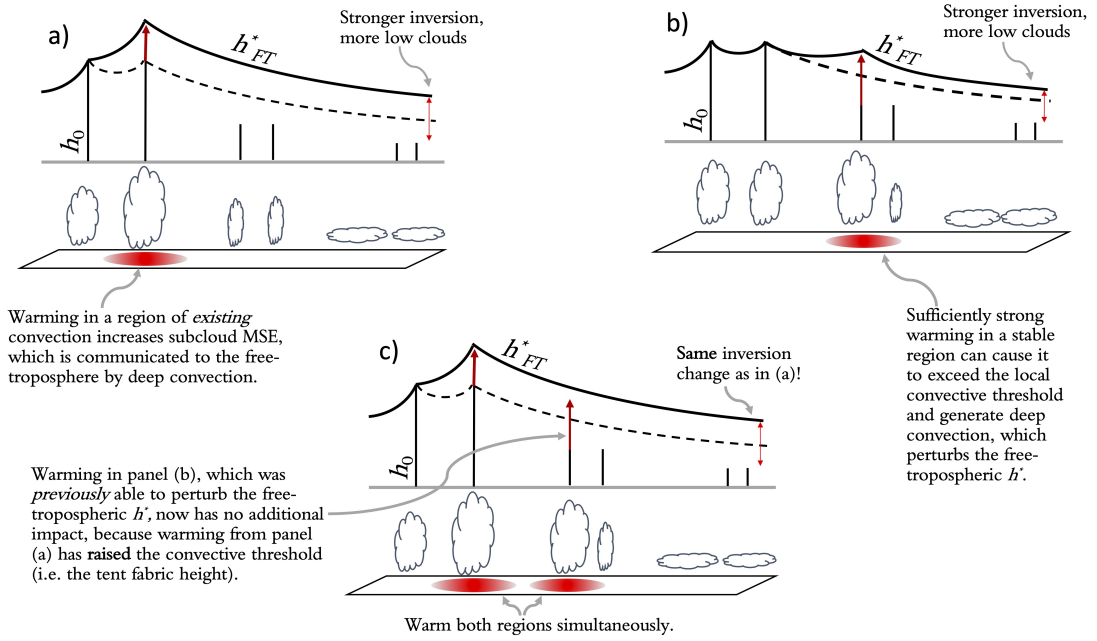


Figure 4.11: Schematic representation of how the ‘circus tent’ framework predicts the climate impact of isolated SST perturbations may not combine linearly. In panel (a) we show a similar sketch to Figure 4.8b of the main text, with warming in a convective region being communicated across the free-troposphere and strengthening the inversion. In panel (b) we show a similar sketch to Figure 4.8e of the main text, which depicts how warming can cause a region which was previously not deeply convecting to exceed the convective threshold and influence the free-tropospheric saturation MSE (h_{FT}^*). In panel (c) we show what would happen if we applied the warming perturbations from panels (a) and (b) simultaneously. In this framework, while the warming in panel (b) was originally sufficient to exceed the convective threshold and perturb the free-tropospheric h^* , is now insufficient because warming in a different region has raised the threshold subcloud MSE required for deep convection.

The main takeaway from Figure 4.12 is that the simulated values of $\overline{\Delta h_{FT}^*}$ tend to be *less* than the inferred values in response to simultaneous warming of multiple regions. For $\Delta SST = +2K$, there are some cases where the inferred and simulated values are similar (e.g., the 140E+220E case), which corresponds to the sensitivity test of Dong et al. (2019b). However, as the size of the SST forcing is increases, there are eventually no cases where the simulated and inferred values are similar, and this is because $\Delta SST = +4K$ is strong enough to generate deep convection even in the 220E case in our model (compare with Figure 4.5c), and thus perturb h_{FT}^* in

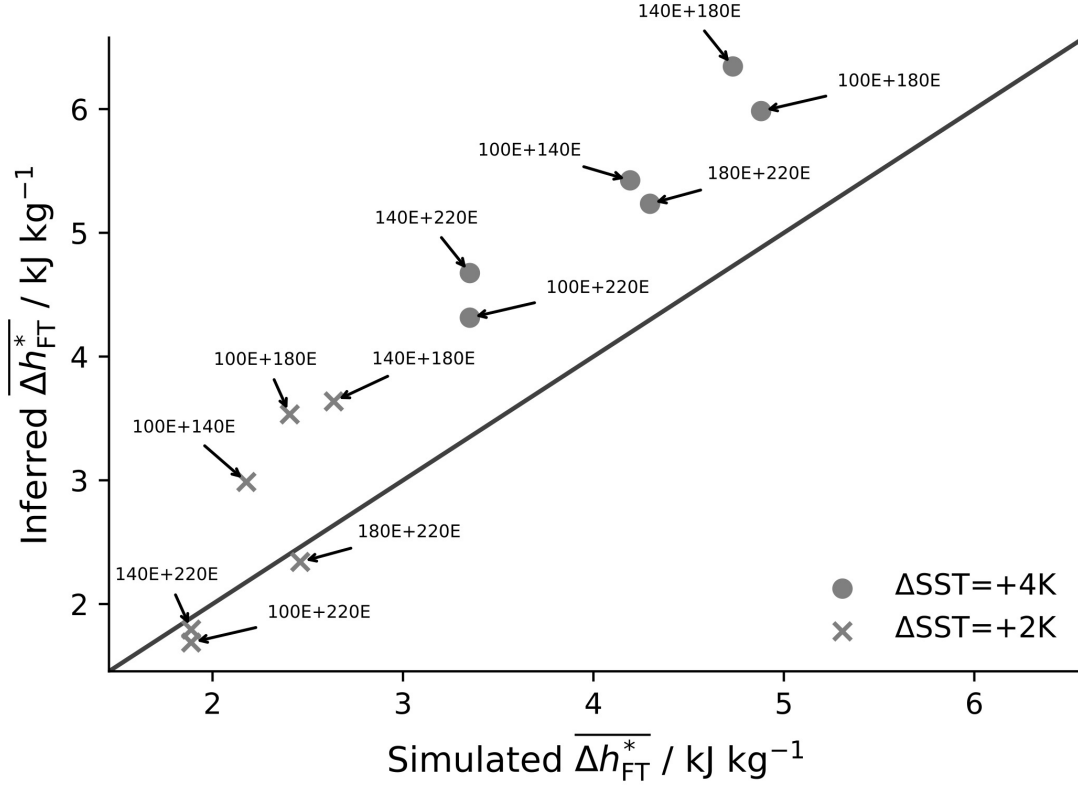


Figure 4.12: Results from “two-patch” experiments using the four equatorial patch locations highlighted in Figure 4.5 of the main text. Plotted are the changes in tropically-averaged saturation moist static energy of the free-troposphere ($\overline{\Delta h_{FT}^*}$) from simulations with two patches imposed simultaneously (x-axis), and the “inferred” values of $\overline{\Delta h_{FT}^*}$ obtained by adding the changes from two simulations with each patch imposed separately (y-axis). The longitudes of the two patches are marked in each case (latitude is 0° in all experiments), and the magnitude of the patch experiments is either +2K (crosses) or +4K (circles). The one-to-one line is shown. Note that in this figure, h_{FT}^* is defined using a bulk average of h^* between 700-300 hPa, as opposed to the 500hPa level as used in the main text.

a manner similar to our sketch in Figure 4.11.

Future work could build on these results by mapping this non-additivity across regions and by applying these results to understand the response of climate to uniform warming, a problem which Green’s function approaches generally struggle with (Dong et al., 2019c; Zhang et al., 2023).

The problem with quotes in a thesis is that no one knows if they are real or not.

— Abraham Lincoln

5

Conclusions

Uncertainty surrounding the climate response to anthropogenic activity is a key challenge for the scientific community. In particular, given that Earth's climate is home to a rich variety of regional circulations, one uncertainty is the exact size and location/pattern of the external perturbations. This uncertainty in the climate response to regional patterns of forcing provides strong motivation for systematic examination of the sensitivity of both global- and regional-climatic changes to the location, magnitude and type of the perturbations. Therefore, as discussed in Chapter 1, this thesis aimed to use a systematic approach to understand how the climate system responds to spatial patterns of forcing. In particular, we have chosen to focus on the climate response to spatial patterns of absorbing aerosol and sea-surface temperatures. These research topics were chosen as they represent sources of intense uncertainty in the literature, and are thus ripe for examination with an idealized modeling approach to map their sensitivities and to better understand the mechanisms behind them.

Furthermore, as noted in the introduction, the climate change problem is often broken down into understanding the *forcing* of the climate system, the *feedbacks* which act as the system moves away from its prior equilibrium state, and the *impacts* of these changes for human society. Another overarching rationale for this

thesis was therefore to explore the sensitivity of forcing, feedback and impacts to variations in the spatial pattern of anthropogenic perturbations.

5.1 Findings

To help formalize our conclusions, we revisit our original main research questions, which formed the basis for Chapters 2, 3 and 4, respectively:

1. Does the global-mean ERF from aerosol-radiation interactions depend on the location of absorbing aerosol? If so, why?
2. How do fast-changes in global and local precipitation depend on the location of absorbing aerosol?
3. Is the global-mean radiative response to an SST linear with the respect to the sign and magnitude of the perturbation? What implications does this have for Green's function approaches?

In Chapter 2, we investigated the sensitivity of global-mean aerosol effective radiative forcing (ERF) to the location of absorbing aerosol. This question was motivated by a desire to investigate the assumptions often made in budgetary assessments of aerosol ERF, and by questions about whether the ‘location dependence’ of aerosol ERF found in previous work was simply a consequence of differences in aerosol lifetimes and transport between regions or arose due to a more fundamental mechanism related to localized absorbing aerosol. To tackle this question, we conducted an ensemble of sensitivity experiments in an atmosphere-only configuration of the ICON GCM, where we systematically place idealized plumes of absorbing aerosol in different locations across the globe. The absorbing aerosol perturbations are designed such that the global-mean AOD perturbation is the same in every experiment, but only their location is different. Furthermore, the aerosol radiative properties are prescribed and non-interactive, which allows us to focus on the essential physics of how regional absorbing aerosol and subsequent

adiabatic heating impacts on the global-mean ERF.

Our main finding in Chapter 2 is that there is no simple ‘one-to-one’ relationship between global-mean absorbing AOD and global-mean ERF. To the contrary, we found that the global-mean aerosol ERF varied by an order of magnitude depending on the location of the absorbing aerosol plume, in addition to changing *sign*. Generally we found a positive global-mean ERF in response to midlatitude absorbing aerosol and absorbing aerosol in regions of strong tropical descent, but a *negative* global-mean ERF when the absorbing aerosol plume was placed in regions of large-scale tropical ascent such as the Western tropical Pacific. We also found that the ‘location-dependence’ was driven primarily by rapid adjustments of cloud liquid water path and circulation, with the dependence of IRF on underlying surface albedo playing a comparatively small role.

To understand these results, we first invoked the seminal work of Hoskins and Karoly (1981), who used linear quasi-geostrophic theory to show predicts that diabatic heating (such as through absorbing aerosol) tends to generate a downstream low pressure anomaly and subsequent cyclonic circulation (in the Northern Hemisphere). Informed by this, we argue that the subsequent equatorward flow induced by the downstream low will reduce the relative humidity local to the plume by drawing down cold, dry air from higher latitudes. We find support for this explanation in our simulations, and thus provides a robust constraint on the ERF from midlatitude absorbing aerosol.

In the tropics, we find that the sign and magnitude of the global-mean ERF depends strongly on where the absorption plume is placed relative to the climatological overturning circulation, as diabatic in the tropics is balanced predominantly by vertical motion. When the plume is placed in regions of climatological descent, the anomalous vertical motion induced by the plume heating interferes destructively with the overturning circulation and generates a localized warming, which reduces

the ambient relative humidity and obstructs cloud formation - a positive, localized ERF. On the other hand, when the plume is placed in regions of climatological ascent, the anomalous upwards vertical motion from the plume heating projects onto the existing ascent structure and strengthens the tropical overturning circulation (that is, the Walker Circulation). The strengthening of the Walker Circulation then generates positive cloud liquid water content anomalies throughout the tropics, which constitutes a negative ERF through increased shortwave reflection.

The most striking feature of our experiments is the strongly negative ERF in response to absorbing aerosol over the Western tropical Pacific. As such, Chapter 2 concludes by examining whether Indonesian wildfires may act as a real-world counterpart for our results using idealized absorbing aerosol perturbations. To investigate this we conducted further experiments using a climatology of Indonesian biomass burning, which was then perturbed so as to produce AOD values comparable to observational estimates of the 2015/2016 wildfire season. These experiments also include a more realistic vertical profile, a seasonal cycle and a less strongly absorbing aerosol burden ($\text{SSA}_{550} = 0.87$) compared to the ensemble simulations. We find that our main result - that absorbing aerosol over the Western tropical Pacific generates negative global-mean ERF through increased low-cloud water content - is robust to these changes and in addition is robust to realistic variations in the vertical profile of the aerosol.

There are two caveats to the work in Chapter 2 which are worth discussing in further detail. The first is that we conducted the ICON plume simulations with no background aerosol concentrations; that is, the control run contains no aerosol whatsoever and the perturbation run contains only the AOD from a single plume. This was done in order to reduce the complexity and potential variability of the simulations, under the assumption that any interactions between background and anthropogenic aerosol would be linear and small. However, as shown by Samset and Myhre (2011), background aerosol can enhance the direct radiative effect of

absorbing aerosol, even in the context of pure radiative transfer calculations, by increasing the amount of solar flux which is backscattered into the aerosol layer. As a result, we expect that the instantaneous radiative forcing from our simulations would be amplified by the addition of a background aerosol distribution. Nevertheless, further work is needed to test this and to establish whether background aerosol distributions also modulate the *rapid adjustments* from aerosol, in addition to their instantaneous radiative forcing.

A second caveat, in a similar vein, arises as Samset and Myhre (2011) also showed that the vertical distribution of cloudiness strongly influences the sensitivity of instantaneous radiative forcing to vertical variations in absorbing aerosol (for example, in Fig. 3 of their paper). As a result, it is clear that any biases in the ICON model with regards to simulating cloudiness will also influence the *sensitivity* of instantaneous radiative forcing to aerosol. Crueger et al. (2018a) thoroughly evaluated the cloud fields simulated by the ICON model in atmosphere-only mode (as used in this thesis) and found that while the net SW and LW cloud radiative effects were well-represented in ICON compared to observational data, the vertical structure of cloud cover was strongly biased. This is a bias common to many CMIP-class GCMs (Giorgetta et al., 2018) and manifests in the ICON model as a ‘too few, too bright’ bias in the subtropical low cloud regions, and an over-representation of middle and upper tropospheric clouds in the tropics and midlatitudes. As we discussed in the introduction, clouds act as an underlying high-albedo surface which strongly controls the instantaneous radiative forcing from absorbing aerosol layers (Haywood and Shine, 1995). Therefore, ICON’s ‘too bright’ bias means that the strength of the cloudy-sky instantaneous radiative forcing over subtropical regions may be biased strong compared to a model with a better climatology of cloudiness.

Furthermore, as ICON simulates too few subtropical low clouds, it may be possible that the *sensitivity* of these regions to Walker Circulation changes is too high. One way of thinking about this is that the cloud cover in a GCM grid-box

cannot exceed 100%, and so regions which already have a large low cloud cover will be less sensitive to strengthening of the inversion because they cannot increase their low cloud cover very much. However, as the ICON model has an under-representation of subtropical low clouds, it is possible that the SW cloudy-sky adjustments we found in Chapter 2 (Fig. 2.5) are too strong. However, without doing similar experiments in a different model (ideally one with a markedly different set of cloud biases) it is difficult to confirm this suspicion. We will return to the question of model diversity and the need for multi-model comparisons later in the conclusions.

Following on from Chapter 2, in Chapter 3 we used the same set of idealized absorbing aerosol experiments to investigate the *impacts* of regional absorbing aerosol on both global and local precipitation changes. This is motivated by work which shows that absorbing aerosol has plays a strong role in the historical and future evolution of global precipitation (McCoy et al., 2022; Samset, 2022). However, these studies have assumed a ‘one-to-one’ relationship between global-mean absorbing AOD and global-mean precipitation changes, which is not obvious *a priori* due to the fact that global-mean precipitation is constrained by shortwave absorption, longwave cooling and sensible heat fluxes, all of which may respond differently to absorbing aerosol in different locations.

In addition to probing the sensitivity of global-mean precipitation to the geographical location of absorbing aerosol, in Chapter 3 we also set out to explore the sensitivity of *local* precipitation changes (i.e., changes in precipitation local to the AOD perturbation) to the location of absorbing aerosol. In particular, we were interested in investigating whether the ‘tropical-extratropical’ contrast in the local precipitation response to aerosol found by previous studies (Dagan et al., 2019, 2021), where tropical aerosol drives a local *increase* in precipitation but extratropical aerosol drives a local *decrease* in precipitation, holds across all locations. Furthermore, the studies of Dagan et al. (2019) and Dagan et al. (2021) explained the local *decreases* in precipitation in response to extratropical absorbing

aerosol by appealing to the local atmospheric energy budget and the process of ‘geostrophic adjustment’. However, while geostrophic adjustment is a ubiquitous feature of rotating fluids, it can not on its own explain the extratropical precipitation decreases. As such, in Chapter 3 we also aimed to provide a *dynamical* argument for why diabatic heating from absorbing aerosol in the extratropics would drive local precipitation decreases.

Our main finding with respect to the global-mean precipitation changes is that, even though the global-mean AOD perturbation is identical across all of our experiments, the changes in global-mean precipitation vary by over an order-of-magnitude. Although the absorbing aerosol perturbations always cause a decrease in global-mean precipitation, as expected through energy budget arguments (Pendergrass and Hartmann, 2014), we find that the magnitude of these decreases varies significantly in response to changes in the location of the AOD perturbation. In particular, the response is much weaker over the Sahara and Eurasia than over the ocean or other land regions. To understand this, we decompose the global-mean precipitation changes into contributions from global-mean shortwave absorption, longwave cooling and sensible heat fluxes. We find that changes in shortwave absorption are relatively agnostic to the longitudinal position of the aerosol plume, although there is a slight latitudinal dependence arising due to less incident solar radiation at higher latitudes. As such, shortwave absorption is not able to explain the ‘location-dependence’ of changes in global-mean precipitation. On the other hand, changes in global-mean sensible heat fluxes and atmospheric longwave cooling both show a marked ‘location-dependence’.

Sensible heat fluxes tend to decrease in the global-mean, due to surface dimming and cooling from enhanced aerosol extinction, which is a positive contribution to global-mean precipitation changes. This positive sensible heat contribution is particularly strong in response to aerosol plumes over the Sahara and Eurasia, where the surface cooling is particularly strong. An outlier in this regard is the

global-mean sensible heat change in response to absorbing aerosol over the Maritime continent region, where aerosol drives an *increase* in global-mean sensible heat flux due to an intensification of the Walker Circulation, and thus a negative contribution to global-mean precipitation changes.

Changes in longwave cooling generally contribute positively to changes in global-mean precipitation, offsetting some of the negative contribution from shortwave absorption. We find that the dominant term in the longwave cooling contribution comes from increases in downwelling, clear-sky emission from the atmosphere to the surface, which is consistent with the fact that absorbing aerosol perturbations heat the atmosphere aloft. We also find a large, positive contribution of the longwave cooling in response to absorbing aerosol over the Maritime continent, which we argue is due to the fact that absorbing aerosol over this region generates a strong convective response and warms the whole of the tropical free-troposphere. We also speculate that enhanced Rossby wave activity from this region may contribute to the increases in global-mean temperature and longwave cooling by increasing the poleward eddy heat flux into the midlatitudes. However, further work is needed to confirm this link.

Our main results with respect to the *local* precipitation changes are as follows. Broadly, our experiments demonstrate a ‘tropical-extratropical’ contrast in the local precipitation response, with negative local precipitation changes in the extratropics and positive changes over most of the tropics. This is in agreement with previous work which has also found a ‘tropical-extratropical’ contrast when looking in a model with fixed-SSTs, but we demonstrate it over a much wider range of climatic conditions. Furthermore, we argue that the negative precipitation changes in response to extratropical aerosol plumes is due (again) to the downstream low pressure system induced by the diabatic heating perturbation (following Hoskins and Karoly (1981)). This downstream low advects dry air towards the aerosol plume and thus drives a local reduction in moisture and inhibits precipitation formation local to the plume. Previous work has pointed out that the local precipitation response to

absorbing aerosol is positive in the tropics as diabatic heating is mostly balanced by upwards vertical motion in the tropics, as a result of the weak-temperature gradient constraint (Dagan et al., 2019; Sobel et al., 2001). However, although we do indeed find a tendency towards a positive local response in the tropics, there are still significant variations between regions. To explain these variations, we introduced a simple measure of subcloud buoyancy which is able to approximately identify the regions where the local precipitation response is largest by highlighting the regions most unstable to deep, precipitating convection. We then manipulate this metric analytically to gain insights into why the local precipitation response is negative over parts of tropical land, which we find is due to a combination of surface cooling over this regions and that these regions are warm and moist in the climatology.

It is worth noting that the mechanism we invoke for explaining the local precipitation response in the midlatitudes is entirely linear and does not take into account the fact that the heating may also be balanced by storm track anomalies which alter the eddy heat flux (e.g., Chemke and Polvani (2020)). The fact that we do not currently *require* such anomalies to explain our results may be due to the fact that we only consider annual-averages (which we expect to smear out the influence of the storm tracks) and also because our aerosol perturbations are circular and not zonally elongated. Zonally-elongated aerosol plumes are more realistic near to storm track entrances as aerosol is advected by the prevailing winds and thus if we used zonally-elongated aerosol plumes we might expect a stronger storm track response (and a larger role for transient eddy heat fluxes in balancing the heating anomaly) due to the fact that diabatic heating projects throughout the baroclinic zone.

Finally, in Chapter 4, we move on from absorbing aerosol and instead focus on the climate impact of spatial patterns of sea-surface temperature anomalies. In particular we are interested in the mechanisms by which localized SST anomalies generate perturbations in Earth’s TOA radiation budget, a process often termed

the ‘pattern effect’. Previous studies have attempted to quantify the pattern effect using so-called ‘SST Green’s Functions’, which are constructed by applying localized SST perturbations in an atmosphere-only GCM. However, such approaches have assumed (either explicitly or implicitly) that the relationship between global-mean TOA fluxes and local SSTs is linear. These assumptions of linearity are stringent, and have not yet been rigorously tested. Thus, the aims of this chapter were two-fold; firstly, to evaluate the assumptions of ‘linearity’ which are essential to prior SST Green’s function studies, and secondly, to explain why deviations from linearity occur by forming a more robust conceptual model of the SST pattern effect.

To investigate the linear assumptions made by previous studies, we conducted a series of 400 SST perturbation experiments in an atmosphere-only configuration of the ICON GCM, where the SST perturbations varied in sign, magnitude and location. Our headline result is that the relationship between global-mean TOA fluxes and local SST changes is highly non-linear, especially when the SST perturbations are in moderately stable regions such as the Central Pacific. We classified these non-linearities into two categories; an ‘asymmetry’ in the TOA response to positive and negative SST perturbations, and a ‘magnitude-dependence’ of the global-mean TOA response on the size of the SST perturbation. Both of these classes of non-linearity have significant implications for Green’s Function approaches to the pattern effect, which have all assumed symmetry with respect to positive and negative SST perturbations, and have often used large SST perturbations (e.g., with $\pm 4\text{K}$ amplitude) in order to maximize the signal in their atmosphere-only runs, even though the SST patterns of interest have amplitudes which are much smaller.

To understand these non-linearities, we proposed a revised conceptual model of the SST pattern effect, which we term the ‘circus tent’ model. In this model, the only way in which SSTs can meaningfully impact the tropical free-tropospheric temperature (or equivalently, h^*), and thus the TOA fluxes, is by enhancing the

subcloud moist static energy (MSE or h_0) in regions of deep convection. Non-linearities arise because the atmosphere is only unstable to deep convection in regions where the subcloud MSE exceeds the saturation MSE of the free-troposphere (the local ‘convective threshold’). As such, a positive SST anomaly in the Central Pacific (where the subcloud MSE often lies very close to this threshold) may sufficiently increase the subcloud MSE such that deep convection occurs and increases the free-tropospheric MSE. On the other hand, a negative SST anomaly in the same region would lower the subcloud MSE and thus further stabilize this region against deep convection, and generate negligible changes in the free-tropospheric MSE. We find that this conceptual, ‘circus tent’ model can explain both the asymmetry and magnitude-dependence we find in our GCM simulations.

In addition to explaining the non-linear TOA response to *isolated* SST anomalies, our circus tent model also predicts a non-*additivity* of tropical SST anomalies. In simple terms, the model predicts that the response of free-tropospheric MSE (and thus TOA flux) will be lower when multiple SST patches are imposed simultaneously in a GCM than when they are imposed separately, and their results summed up. We test this prediction using an additional set of experiments where we choose pairs of SST warming patches and impose them simultaneously. We find that for situations where the SST warming is small (+2K amplitude), the non-additivity only sets in when simultaneously perturbing two ‘convective’ regions. However, for larger SST warming perturbations (+4K amplitude), we find that the non-additivity is present for all combinations of patches we tested. For example, the inferred h_{FT}^* perturbation from the additivity assumption is approximately 20% larger than the true h_{FT}^* perturbation found when two SST patches are imposed simultaneously in the ICON model (Figure 4.12). This is because a +4K SST anomaly is sufficient to trigger deep convection in all the regions tested in our model, and deep convection is the crucial ingredient in driving this non-additivity.

5.2 General conclusions and recommendations for future work

This thesis has explored and quantified how the climate response to localized absorbing aerosol and SST perturbations depends on their geographical location.

Firstly, this thesis has demonstrated that both the global-mean effective radiative forcing (ERF) and global/local precipitation responses from absorbing aerosol depend strongly on the location of the aerosol plume. These results run counter to the assumptions made in many prior studies (e.g., Stevens (2015a)), who ignored this ‘location-dependence’. The sensitivity of ERF to the location of absorbing aerosol in particular has potential implications for regional emissions mitigation, as our results demonstrate that an identical AOD burden over one country may not necessarily be as impactful (as measured by the global-mean ERF) as if it were placed over a different country. One general limitation of our work is that we have not tested these mechanisms in a model with a full aerosol-climate model, which would introduce new dimensions to this problem such as the translation from emissions to concentrations, and would also introduce new sources and sinks of aerosol to the problem (notably wet deposition). Despite this, our idealized approach (where we prescribe aerosol radiative properties) provides a firm baseline for future studies with more comprehensive models.

Furthermore, in our idealized framework, we have focused exclusively on strongly absorbing aerosol mixtures (with a single-scattering albedo = 0.8), and as such our results are not immediately applicable to historical aerosol loading, which tend to be less absorbing. One simple way to build upon our work would thus be to redo a set of similar experiments to ours, but with a higher single-scattering albedo. This would allow a quantification of the role of scattering vs absorbing aerosol in setting any potential ‘location-dependence’ of aerosol ERF or precipitation responses in future climates. Additionally, we have represented our absorbing aerosol plumes as two-dimension Gaussian functions, but as we show in Chapter 4 it is not necessarily

true that one can ‘add up’ these plumes in order to recreate the response to arbitrary aerosol distributions. The degree to which the climate response to a realistic distribution of aerosol can be considered as the sum of these Gaussian functions (the ‘additivity’ problem) should be explored in future work.

Our results in Chapters 2 and 3 also demonstrate that the ‘location-dependence’ of the ERF/precipitation responses to absorbing aerosol arise largely due to the ways in which the large-scale atmospheric circulation adjusts to the aerosol plume. The results of this thesis thus represent the first time that circulation adjustments have been directly implicated in the aerosol ERF, expanding our conception of ‘rapid adjustments’ to climate forcings such as black carbon. However, this result runs counter to earlier work by Johnson et al. (2019), who argued that circulation adjustments are not relevant for the global-mean ERF from black carbon. As we point out in Chapter 2, this could potentially be due to the fact that they choose to run their experiments with emissions from a year without significant biomass burning over the Maritime continent, which is a region we find to be particularly sensitive. In addition, Johnson et al. (2019) used a comprehensive aerosol scheme, as opposed to the idealized aerosol parameterization we used. Testing the sensitivity to aerosol representation and, more generally, understanding the sources of discrepancy between our idealized work and previous work with comprehensive models would be useful for future work to explore. We hypothesize that the inclusion of wet deposition in Johnson et al. (2019) is a strong source of disparity in our results due to the fact that wet deposition causes preferential aerosol washout in convectively active regions.

In addition to demonstrating a robust ‘location-dependence’ of the climate impacts of absorbing aerosol, we have also investigated how the Earth’s TOA energy budget responds to localized SST perturbations. A concept of ‘location-dependence’ is already established in this context, and often termed the ‘SST pattern effect’. However, studies of the SST pattern effect have often assumed that the global-mean TOA response is linear with respect to the sign and magnitude of the SST anomaly.

Our main finding in Chapter 4 is that this assumption fails quite spectacularly, particularly in regions only moderately stable to deep convection, such as the Central Pacific. For example, the TOA response to a -4K SST perturbation in the Central Pacific is negligible, whereas the response to a +4K perturbation is strongly negative, around -3Wm^{-2} in the global-mean (Figure 4.5). As a result, our results in Chapter 4 suggest that SST Green's Functions constructed with only SST anomalies of a single sign will not generally be skillful in reproducing the TOA flux response to SST anomalies of the opposite sign. One interesting avenue for future work would thus be to test if constructing a 'piecewise' Green's Function, where separate Green's Functions are constructed for positive and negative SST perturbations, improves their skill in reconstructing climate features such as variations in TOA fluxes.

Our SST perturbation experiments also demonstrate a robust 'magnitude-dependence' of the TOA flux response on the magnitude of the imposed SST anomaly. That is, the TOA flux response can be non-monotonic even for a fixed sign of the SST perturbation. Hence, another recommendation from Chapter 4 is that SST Green's Functions be constructed using perturbation magnitudes as close as possible to the magnitude of the anomalies they aim to reconstruct the climate response to. This may sound trivial, but many studies impose very large perturbation magnitudes (4K amplitude) in order to maximize the signal-to-noise ratio in their atmosphere-only GCM runs. Our work suggests that this practice may fundamentally bias the Green's Functions due to the magnitude-dependence we demonstrate.

Preliminary work (Jonah Bloch-Johnson, personal communication) has also demonstrated that the degree of non-linearity in the TOA response to positive and negative SST perturbations also depends strongly on the *spatial scale* of the perturbations, with smaller perturbations tending to generate a stronger non-linearity than perturbations on larger spatial scales. This scale-dependence also

implies that future SST Green's Functions should be constructed using perturbations of a similar spatial scale to the SST anomalies they will be projected onto.

The 'circus tent' model we propose in Chapter 4 for understanding these non-linearities also predicts that the climate response to SST anomalies is non-*additive*. That is, the response to imposing two SST perturbations simultaneously in a GCM is different from imposing them in two separate GCM runs and then adding together the climate responses (up to a 20% difference). This notion of 'additivity' is also a fundamental assumption of SST Green's Functions constructed with isolated 'SST patch' experiments. As such, our results suggest that this SST patch approach may be limited in comparison to other approaches such as the 'random field' method of Li and Forest (2014), and further work is necessary in order to understand the benefits of such an approach. This will be possible in the upcoming Green's Function Model Intercomparison Project (GFMIP, Bloch-Johnson et al. (2023)), which I am involved with and which my work in Chapter 4 has helped to motivate.

Three general limitations of the work in this thesis should also be addressed. Firstly, in this thesis I have only conducted experiments with a single GCM. This means there is uncertainty attached to all the results in this thesis as GCMs exhibit distinct biases in their mean state, variability and response to forcing. Future work should thus aim to test the mechanisms put forward in this thesis in a variety of different GCMs from different modeling centres.

Secondly, my analysis has primarily focused on the time-mean response of the atmosphere to spatial patterns of perturbations and in doing so I have not focused on how these spatial sensitivities may vary from season-to-season. A logical next step would thus be to extend my analysis to look at how these spatial sensitivities vary throughout the annual cycle. In regions which have a pronounced seasonal cycle, such as monsoonal climates, we may expect qualitatively different sensitivities in the monsoon season compared to the dry season, in particular when considering

the local precipitation responses in Chapter 3. Furthermore, the seasonality of the storm track regions may also project onto our spatial sensitivities as the baroclinic zone shifts either towards or away from the imposed heating anomaly. Lin et al. (2018) demonstrated that a wave-like response is favoured when a heating anomaly is coincident with the baroclinic zone and the maximum in the Eady growth rate, but that the response is weaker when the heating anomaly moves away from this region. Hence, it is possible that in our simulations may exhibit more wave-like responses to the imposed aerosol perturbations when looking on seasonal scales, but that this response is muted in the annual-mean compared to the linear mechanisms we highlight in Chapters 2 and 3.

Thirdly, our experiments have exclusively been conducted using fixed-SSTs. While this is recommended for the ERF calculations we present in Chapter 2, the use of fixed SSTs does limit us to studying only the ‘fast’ response of precipitation to aerosol in Chapter 3. More broadly, the use of fixed-SSTs as opposed to a slab ocean or coupled atmosphere-ocean model means that the surface fluxes are unphysically constrained and not fully able to respond to the imposed aerosol burdens or circulation changes. For example, in both Chapter 2 and Chapter 3 we note that absorbing aerosol over the Maritime continent induces a strong convective response and a strengthening of the Walker Circulation. However, in a model with interactive SSTs, the surface cooling due to aerosol extinction aloft would likely reduce the underlying SSTs and thereby suppress convection. Furthermore, in a fully-coupled model the surface wind stress imposed by a strengthened Walker Circulation would likely cause a shoaling of the thermocline and enhanced upwelling of colder water, which may also act as a negative feedback on the Walker Circulation changes. Ongoing work imposing large increases in biomass burning emission over Indonesia in a fully-coupled climate model has also demonstrated a strengthening of the Walker Circulation (Matthew Kasoar, personal communication), but the timescales over which these changes occur are still unclear. As such, a fruitful avenue for future work would be to repeat the idealized absorbing aerosol experiments I

conducted, but in a slab ocean model and then test a few regions of particularly high sensitivity in a coupled model to assess their consistency.

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