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Key Points:

- Adjustments impact the Earth energy budget, but also circulation, precipitation and atmospheric structure
- Adjustments are classified into six different mechanisms and act at time scales ranging from seconds to multiple years
- Observational constraints can inform on some aspects of adjustments

Supporting Information:

Supporting Information may be found in the online version of this article.

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




















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Adjustments to Climate Perturbations—Mechanisms, Implications, Observational Constraints

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Abstract Since the 5th Assessment Report of the Intergovernmental Panel on Climate Change (AR5) an extended concept of the energetic analysis of climate change including forcings, feedbacks and adjustment processes has become widely adopted. Adjustments are defined as processes that occur in response to the introduction of a climate forcing agent, but that are independent of global-mean surface temperature changes. Most considered are the adjustments that impact the Earth energy budget and strengthen or weaken the instantaneous radiative forcing due to the forcing agent. Some adjustment mechanisms also impact other aspects of climate not related to the Earth radiation budget. Since AR5 and a following description by Sherwood et al. (2015, <https://doi.org/10.1175/bams-d-13-00167.1>), much research on adjustments has been performed and is reviewed here. We classify the adjustment mechanisms into six main categories, and discuss methods of quantifying these adjustments in terms of their potentials, shortcomings and practicality. We furthermore describe aspects of adjustments that act beyond the energetic framework, and we propose new ideas to observe adjustments or to make use of observations to constrain their representation in models. Altogether, the problem of adjustments is now on a robust scientific footing, and better quantification and observational constraint is possible. This allows for improvements in understanding and quantifying climate change.

Plain Language Summary Climate change is driven by perturbations to the atmospheric composition, to land use, or by changes of incoming solar radiation. It can be understood energetically by quantifying the perturbation to the Earth energy budget—the instantaneous radiative forcing—and the response of the climate system to this perturbation. This response can be split into feedbacks—mechanisms that act in response to global-mean surface temperature changes—and other processes that act independently of the global-mean surface temperature change. These latter processes are called adjustments. There is also a category of climate-relevant adjustments that is not directly related to the energy budget. This review documents the improved classification, understanding, constraint, and quantification of adjustments. A clearer picture of adjustments allows to better understand and quantify climate change.

1. Introduction

Climate change is caused by perturbations to the state or composition of the Earth system that change the balance between the net solar radiation absorbed and the terrestrial radiation emitted to space by the Earth's atmosphere and surface. Perturbations affecting the radiative budget include changes to atmospheric composition (especially greenhouse gases and aerosols or aerosol precursors), changes in land use, and changes to incoming solar

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radiation. The resulting change in the net imbalance of radiative flux is known as the *radiative forcing* \mathcal{F} . Radiative forcing was first identified as a means of quantifying the climate response to perturbations by Ramanathan (1975). Understanding was consolidated through the late 1970s and early 1980s (e.g., Dickinson & Cicerone, 1986; Ramanathan & Dickinson, 1979; Ramanathan et al., 1979) so that the concept of radiative forcing was in widespread use by the time of the first assessment of the Intergovernmental Panel on Climate Change (IPCC, 1990). Ramaswamy et al. (2019) provide further historical context.

A radiative imbalance will induce a surface temperature change as the system warms or cools toward a new equilibrium. The energy imbalance N caused by a forcing will be damped by the net climate feedback λ acting on the surface temperature perturbation ΔT_s , which can be described in a linearized form for small perturbations around an equilibrium state in the global mean:

$$N = \mathcal{F} + \lambda \Delta T_s. \quad (1)$$

Equation 1, which applies to the time-integrated global mean, embodies the “forcing-feedback” framework for temperature change (Gregory et al., 2004). The top-of-atmosphere imbalance N is taken up in Earth's climate system, with ocean heat uptake accounting for about 90% of the total (Forster et al., 2021) at present. Implicit in Equation 1 is the assumption that surface temperature change is uniquely determined by the feedback parameter and the forcing regardless of the mechanism inducing the forcing.

A literal interpretation of Equation 1 suggests that \mathcal{F} may be determined as the difference in top-of-atmosphere flux between two radiative transfer calculations that are identical except with respect to the forcing agent being present or not—the “instantaneous radiative forcing” at the top of the atmosphere. \mathcal{F} so computed is, however, an imperfect predictor of eventual temperature change (e.g., Hansen et al., 1997). Especially vexing is that the feedback parameter computed with this definition often depends on the forcing agent.

Computing radiative forcing as the flux change at the tropopause, rather than the top of the atmosphere, increases the degree to which the same forcing magnitude elicits the same feedback and surface temperature response, irrespective of the mechanism causing the forcing (Hansen et al., 1997). Early work showed further that “an improved measure of radiative forcing is obtained by allowing the stratospheric temperature to adjust to a radiative equilibrium profile” (Hansen et al., 1997). Stratospheric temperature adjustment is especially important for forcing by carbon dioxide, which acts to cool the stratosphere (Manabe & Wetherald, 1967), further reducing the terrestrial radiation lost to space (i.e., the stratospheric cooling increases the radiation imbalance N). This temperature *adjustment*, which returns the stratosphere to a state of (global-mean) radiative near equilibrium, acts on a more rapid timescale (weeks–months) than the surface temperature response (years to decades), and is independent of surface temperature change. This observation led early investigators (e.g., Ramanathan & Dickinson, 1979; Ramanathan et al., 1979) to view changes due to the stratospheric temperature adjustment as part of the forcing. Forcing including this adjustment is often reported at the tropopause but, since stratospheric radiative equilibrium requires no radiative flux divergence across the stratosphere, forcing at the tropopause and top-of-atmosphere are identical once the stratospheric temperature adjusts.

Further experience with climate models (starting with Hansen et al. (1997)) has demonstrated that stratospheric temperature adjustments in response to carbon dioxide and other forcings is one of many *adjustments* to forcing in the climate system. Adjustments are defined as changes in the state or composition of the Earth system, caused by an initial forcing agent, that modify the top-of-atmosphere energy imbalance in the absence of a change in global-mean surface temperature (Andrews & Forster, 2008; Bony & Stevens, 2020; Colman & McAvaney, 2011; Gregory & Webb, 2008; Sherwood et al., 2015). The term is used to note both the change in state and the resulting change in top-of-atmosphere flux. Cloud adjustments occurring in response to arbitrary forcings were recognized more than 15 years ago (Andrews & Forster, 2008; Gregory & Webb, 2008). Cloud adjustments to aerosol forcing as a consequence of aerosol-cloud interactions that was identified even much earlier (Lohmann et al., 2010; Rotstain & Penner, 2001), are thought to be an important component of total anthropogenic aerosol forcing (Bellouin et al., 2020).

The recognition of the substantial role of adjustments motivated the definition of *effective radiative forcing* (ERF) or \mathcal{E} in the notation of Bellouin et al. (2020) as the sum of the (purely radiative) instantaneous response and any adjustments. Climate model simulations follow Equation 1 more consistently when \mathcal{E} replaces \mathcal{F} , that is, effective

radiative forcing is a better predictor of global temperature change than the instantaneous radiative forcing computed with fixed state and composition. Adjustments are also useful in understanding precipitation responses to forcing (e.g., Richardson et al., 2018).

Here we assess and synthesize the current state of understanding on adjustments, updating the knowledge since the review by Sherwood et al. (2015). Section 2 highlights the evolving conceptual understanding of adjustments including limitations in existing diagnostic techniques, with a more explicit categorization than in Sherwood et al. (2015). The wide range of mechanisms by which adjustments can arise is surveyed in Section 3, highlighting several that have recently come to light. Section 4 takes a closer look at non-radiative adjustments such as atmospheric circulation and precipitation changes, and Section 5 highlights opportunities for constraining adjustments with observations, with a particular focus on aerosol-cloud interactions. Opportunities for developing understanding are highlighted in the final section.

2. Definitions and Framework

Measures of radiative forcing quantify the difference in energy balance between two states. As only one of these states is directly observable, estimates of radiative forcing necessarily rely on modeling. The character of the modeling depends on the choice of forcing metric and introduces uncertainties and constraints.

Instantaneous radiative forcing, as noted above, can be determined directly from radiative transfer calculations for which the physical basis is extremely strong. Stratospheric temperature adjustments may be estimated by assuming that the stratosphere is close to radiative equilibrium and that the resulting radiative heating rates remain constant after forcing. Forcing under this “fixed dynamical heating” assumption can be computed by iteratively adjusting the stratospheric temperature profile in the presence of the forcing agent until it reproduces the radiative heating rates in the unperturbed state (e.g., Fels et al., 1980; Forster et al., 1997). Stratospheric temperature adjustments substantially enhance forcing by carbon dioxide and perturb ozone forcing but have little impact on optically thinner gases such as methane and halocarbons (e.g., Pincus et al., 2020).

In general, however, assessing adjustments requires calculations with more complete models of the climate system. Two general approaches are available for estimating adjustments and the resulting ERF from climate models. One, introduced by Gregory et al. (2004), exploits the linearity of Equation 1 to determine λ as the slope and \mathcal{E} as the intercept of a linear regression of N against ΔT_s in simulations with abruptly increased forcing. The second suppresses ΔT_s in Equation 1 by fixing surface temperature at its control value and measuring the top-of-atmosphere imbalance in the presence of forcing agents (Hansen et al., 2005). Fixing land surface temperatures in climate models has historically been a technical challenge, so current practice is to calculate ERF by fixing sea surface temperatures and sea ice concentrations (Forster et al., 2016). Radiative kernels (Soden et al., 2008) can be applied to correct the ERF estimated with fixed sea surface temperatures to account for land surface temperature changes.

These approaches represent two approximations to a conceptual definition of adjustment. Because global-mean temperature increases even in the first year of a simulation with abrupt forcing, regression (the “Gregory” method), by defining ERF as the imbalance for a state with zero global-mean surface temperature change, requires adjustments to occur on rapid (sub-yearly) timescales. This definition inspired the use of the term *rapid adjustments* though this language is potentially misleading (Forster et al., 2021). Surface temperature change suppression (the “Hansen” approach), in contrast, imposes no constraint on the timescale of adjustments, but fixes surface temperature at each point (rather than only in the global average), potentially suppressing circulations and adjustments caused by local temperature anomalies. In practice the two approaches generally yield similar ERF values for individual models relative to the large inter-model spread (Chung & Soden, 2015; Forster et al., 2016).

The implications of such definitions can be understood by expanding Equation 1:

$$\bar{N}(t) = \bar{\mathcal{F}} + \sum_i \bar{A}_i(t) + \sum_j \left(\overline{\lambda_{A,j}(t) \Delta T_j'(t)} \right) + \sum_k \lambda_{fb,k} \bar{\Delta T}_s(t) \quad (2)$$

Here the imbalance includes the instantaneous radiative forcing \mathcal{F} damped by radiative feedbacks (k indexes feedback processes including the Planck, surface albedo, water vapor, lapse-rate and cloud feedbacks with feedback parameters $\lambda_{fb,k}$) that occur in response to the global-mean surface temperature change $\bar{\Delta T}_s(t)$. These

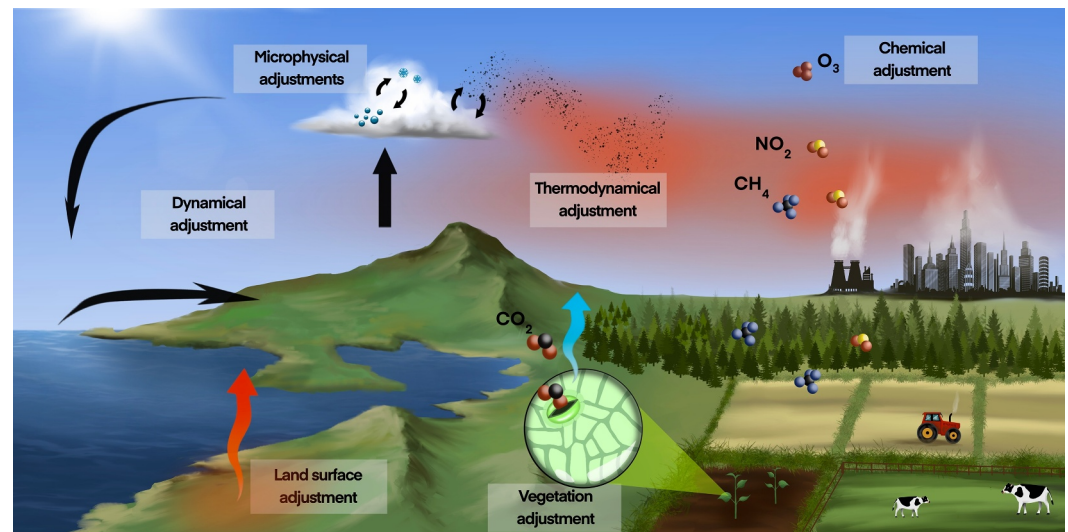


Figure 1. Mechanisms of adjustments: Thermodynamical adjustments, land surface temperature adjustments, vegetation and chemical adjustments, cloud microphysical adjustments in response to aerosol perturbations, and circulation adjustments in response to temperature anomalies.

feedbacks may occur across a range of time scales (e.g., Hansen et al., 2023). There are three classes of adjustments.

- *Adjustments independent of any surface temperature changes.* This is the best-defined class of adjustments. In Equation 2, it is the second term (A_i), where i indicates the summation over different adjustment mechanisms.
- *Adjustment mechanisms involving local temperature changes $\Delta T'$ with (near-)zero global-mean temperature change.* These are described by the third term. This term is often somewhat problematic in interpretation and in disentanglement from the adjustments A_i and from the feedbacks. We refer to these later as “land surface temperature adjustments” and “dynamical adjustments.”
- *Adjustment processes that are relevant for climate but not for the Earth radiation budget.* These do not appear in the above equation and thus are not damping or amplifying climate change. They are thus outside the forcing-feedback framework. It seems useful to nevertheless also consider them when discussing mechanisms acting independent of global-mean surface temperature change (Section 4).

The terms for the adjustments in Equation 2 are assumed to add up linearly in analogy to the feedback terms. An interesting manifestation of adjustments is the effect of orbital forcing that leads to ice age cycles. Over time scales of tens to hundreds of millennia, large climate changes such as the glacial—interglacial ages are caused by periodic change in properties related to the Earth’s orbit around the Sun such as its eccentricity, obliquity and precession. For the latter two, only the seasonal and latitudinal distribution of incident sunlight is affected but not the global annual mean. This implies that glacial-interglacial temperature changes arise almost entirely from adjustments (the growth or decay of ice sheets) which increase or decrease surface albedo. The differential absorption of sunlight could also drive changes in the Hadley circulation and in monsoons (Erb et al., 2015). While this is a rather extreme case in the context of adjustments and climate change since preindustrial times, it serves to highlight the limitations of the linearized forcing-feedback framework in Equation 1.

We note tangentially that Equation 2 has wider applications. Simple physical emulators of climate change adopt this energy budget framework to model global temperature projections in response to radiative forcing or emissions scenarios, with typically a two or three layer ocean model to represent different layers or timescales (e.g., Geoffroy et al., 2013). Recent characterizations of such models link emissions to ERF, explicitly accounting for adjustments within their ERF evaluation (Smith, Forster, et al., 2018).

3. Adjustment Mechanisms

Different drivers of climate change may cause different adjustment mechanisms (Figure 1). The mechanisms discussed below are included in the second term on the right-hand-side of Equation 2 (thermodynamical,

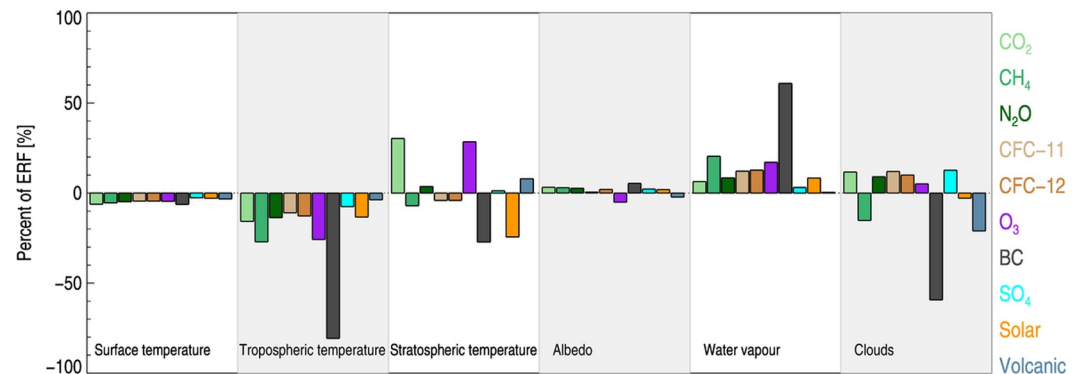


Figure 2. The relative importance (in %) of adjustment processes to ERF for an increase in 10 different climate drivers. These include atmospheric concentrations of greenhouse gases (CO₂, methane/CH₄, nitrous oxide/N₂O, halocarbons/CFC-11 and CFC-12, and ozone/O₃), absorbing (BC) and scattering (SO₄) aerosol as well as volcanic aerosol, and incoming solar radiation changes. Note that SO₄ and volcanic aerosols have negative ERF. Numbers are taken from Hodnebrog et al. (2020), Marshall et al. (2020), Skeie et al. (2020), and Smith, Kramer, et al. (2018).

vegetation, microphysical and chemical adjustments) and in the third term (land surface and dynamical adjustments). Most of the former mechanisms (in particular, microphysical and chemical adjustments, but also vegetation adjustments to a large extent) are due to the nature of the driver of the forcing, rather than due to the instantaneous radiative forcing itself.

1. *Thermodynamical adjustments.* Some climate drivers interact with solar or terrestrial radiation to cause heating or cooling within the atmosphere. This can trigger further changes to surface fluxes, clouds, water vapor and precipitation.
2. *Vegetation adjustments.* Vegetation reacts to elevation in CO₂ concentrations (fertilizing effect), to tropospheric ozone (damaging effect), and to aerosols (increase in diffuse radiation) altering surface energy fluxes, emissions of radiatively active atmospheric constituents (e.g., dust) and emissions of ozone and aerosol precursors (e.g., biogenic volatile organic compounds (BVOCs)).
3. *Microphysical adjustments.* Clouds and precipitation respond to changes in the abundance of cloud condensation nuclei (CCN) and ice nucleating particles (INPs), altering their microphysical state and thus their radiative properties in response to anthropogenic aerosol and aerosol precursor emissions.
4. *Chemical adjustments.* Atmospheric chemistry acting on anthropogenic emissions of chemically active trace gases leads to changes in oxidizing capacity and other forcing constituents (e.g., ozone).
5. *Land surface adjustments.* Heating or cooling of the surface by for example, atmospheric aerosols or greenhouse gases, aerosol-cloud interactions or surface albedo changes, may lead to adjustments by modifying the surface energy budget and surface fluxes even with no change in global-mean surface temperature or, depending on definition, if changes in surface temperature are limited to land only.
6. *Dynamical adjustments.* Anomalies in atmospheric circulations are induced by changes in surface and atmospheric temperature patterns, with consequences for the radiation budget for example, by changes in cloud patterns.

These six mechanisms of adjustments will be discussed in the following subsections.

Another way of analyzing adjustments is by separating their effects on the different state variables relevant for radiation, as is commonly done in feedback analysis (e.g., Bony et al., 2006). Several studies have investigated adjustments to different climate drivers in this way, by using radiative kernels (Hodnebrog et al., 2020; Marshall et al., 2020; Skeie et al., 2020; Smith, Kramer, et al., 2018; Smith et al., 2020). Figure 2 illustrates how they diagnose the adjustment processes, and their relative importance compared to the ERF. The surface temperature and albedo adjustments are weak for all drivers shown in Figure 2. BC stands out with a relatively large change for tropospheric temperature, water vapor and cloud adjustments. Stratospheric temperature adjustment is important for CO₂, but also for other drivers such as ozone, BC and changes to insolation.

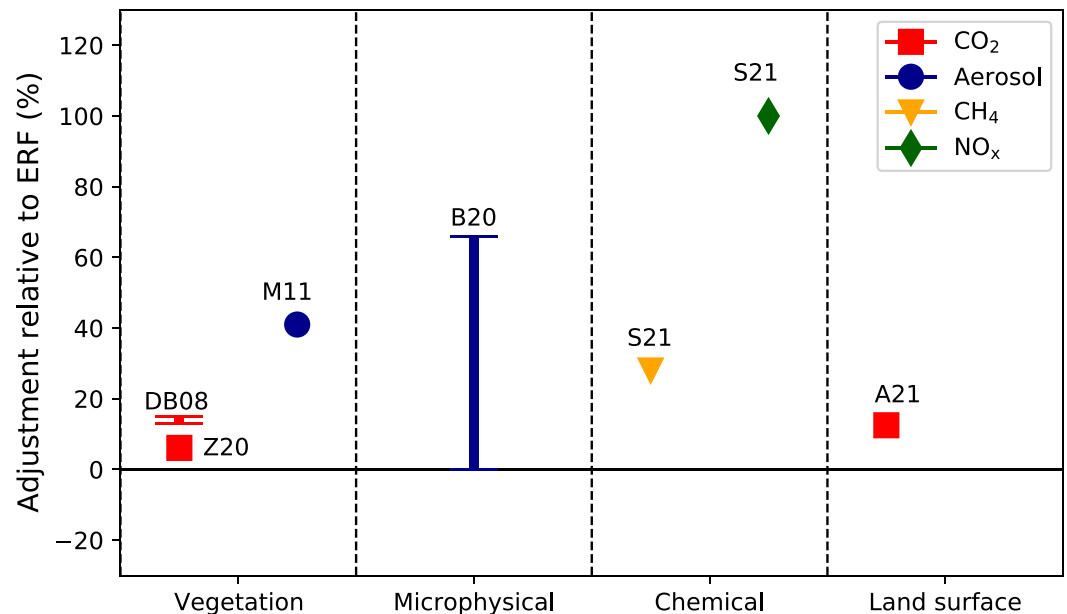


Figure 3. The relative importance (in %) of adjustment processes to ERF for four different mechanisms. The values are taken from the literature. Adjustments to increases in atmospheric CO₂ concentrations, to changes in aerosol concentrations, and the chemical adjustments contributing to the effective forcing by methane and NO_x emissions are considered as examples. Values are from Doutriaux-Boucher et al. (2009, DB09), Zarakas et al. (2020, Z20), Mahowald (2011, M11), Bellouin et al. (2020, B20), Szopa et al. (2021, S21), and Andrews et al. (2021, A21).

Figure 2 addresses the thermodynamic adjustments. For the other adjustment mechanisms, no systematic multi-model assessment is available yet. A compilation of some studies on relative magnitudes of the different adjustment mechanisms for different drivers is shown in Figure 3.

A summary of the level of scientific understanding and the level of quantification is provided in Table 1.

3.1. Thermodynamical Adjustments—Adjustments of Temperature, Humidity, and Cloud Profiles

Thermodynamical adjustments are the changes in atmospheric heating rates and, subsequently, in temperature, humidity and cloudiness, in response to absorption or emission of radiation by climate drivers. Figure 4 illustrates how atmospheric absorption causes adjustments and feedbacks on different timescales following increases in three very distinct climate drivers, namely CO₂, BC and sulfate aerosols (based on results from Stjern, Forster, et al. (2023)). CO₂ causes an immediate radiative cooling in the stratosphere and a weaker radiative heating in the troposphere. The resulting stratospheric temperature reductions occur on a timescale mainly from weeks to months, while most of the tropospheric temperature increase is slower. The CO₂-driven temperature adjustments cause changes in atmospheric static stability and in humidity, triggering cloud changes in the lower and upper troposphere within days to weeks. A further effect is the change in cloud-top radiative cooling in response to altered greenhouse gas concentrations with consequences for cloud lifetime (Bretherton, 2015). Larger cloud changes, on much longer timescales, follow as surface temperature-driven feedbacks increase with time (see Section 3.5). Whereas CO₂ mostly interacts with longwave radiation, BC interacts with solar radiation. The solar absorption by BC causes a rapid temperature increase in the atmosphere, with a vertical structure depending on the vertical distribution of BC (Stjern et al., 2017). Large cloud reductions in most of the troposphere and cloud increases near the surface occur within days to weeks. Sulfate aerosols have initially a small impact on the absorption of radiation within the atmosphere and adjustments are very weak. Changes to the incoming solar radiation similarly mostly act on the surface energy budget but have some effect on atmospheric heating in the troposphere via changes in the ultraviolet (Gray et al., 2009).

Table 1
Summary of Adjustment Mechanisms

Adjustment	Level of scientific understanding	Level of quantification
Thermodynamical	<p>Medium. (T)—Clear concept for radiative heating due to absorption, less clear for cloudiness changes. (M)—Qualitative agreement across models for large-scale averages for radiative heating, broadly also for cloud changes. (O)—Observational evidence clear for stratosphere, but not yet for the troposphere</p> <p>Low. Basic physiological responses of vegetation to CO₂ are understood but radiative effects and further climate responses are not well understood. (M) Very limited analysis of the contribution of vegetation changes to radiative forcing exists and uncertainties have not been evaluated. (O) Observations suggest that vegetation influences Earth's radiative budget but understanding is limited and not specific to radiative forcing</p>	<p>Medium. Large model diversity for cloud adjustments. However, no observational estimate yet</p>
Vegetation	<p>Low. Basic physiological responses of vegetation to CO₂ are understood but radiative effects and further climate responses are not well understood. (M) Very limited analysis of the contribution of vegetation changes to radiative forcing exists and uncertainties have not been evaluated. (O) Observations suggest that vegetation influences Earth's radiative budget but understanding is limited and not specific to radiative forcing</p>	<p>Low. Estimates of the contribution of vegetation changes to radiative forcing has only been diagnosed in a few cases and no quantification of uncertainties or inter-model spread have been conducted</p>
Cloud microphysical	<p>Medium. (T)—Multiple mechanisms explained, partly of opposite sign. Less clear but likely much smaller signal for mixed- and ice-phase clouds. (M)—Models tend to agree on increases in horizontal and vertical extent as aerosol concentrations increase. (O)—Mixed results on vertical extent</p>	<p>Low. The magnitude, and in some cases even the sign, of microphysical adjustments, is still very uncertain. Observational estimates are available but no consensus reached yet</p>
Chemical	<p>Medium in general, with uncertainties in secondary aerosol effects. (T)—Good theoretical understanding of chemical responses to emissions. (M)—Good qualitative agreement on ozone and methane lifetime adjustments; Poor agreement for other trace gases (e.g., VOCs) and secondary aerosol effects. (O)—Strong evidence in some cases (e.g., stratospheric ozone); Inconsistencies in tropospheric ozone observed trends</p>	<p>Medium. Good quantitative agreement in some cases (e.g., ozone ERP); Large model diversity in aerosol-mediated cloud adjustment from oxidizing capacity changes. Observational constraint feasible in some cases (e.g., ozone adjustment from halocarbon perturbation), no estimate of aerosol response to oxidizing capacity</p>
Land surface temperatures	<p>Medium. (T) Clear understanding of thermodynamic underpinnings as surface temperature warms with a positive instantaneous forcing but (M) uncertainty in the magnitude of the direct surface temperature adjustments remains and uncertainty of indirect adjustments associated with the temperature change has not been explored. (O) No observational support yet. Isolating these adjustments may not be possible in observations</p>	<p>Low. Only the total surface temperature adjustment is regularly quantified. Isolation of the contributions from the land temperature change to other adjustments has been infrequent. No observational quantification exists</p>
Dynamical	<p>Low in general, though some mechanisms better understood. (T)—Good understanding of circulation responses to land-sea contrast. Many mechanisms for eddy-driven circulation shifts have been proposed, but exact mechanism uncertain. (M)—Generally considerable model uncertainty in circulation adjustments, although most models simulate a poleward shift of the jet and storm track for CO₂. (O)—No observational evidence yet</p>	<p>Low. Qualitative model agreement on CO₂ adjustments, but greater uncertainties for aerosols. No observational estimate yet</p>

Note. The level of scientific understanding is considered “high” where a sound theoretical (T) basis is available and is supported by modeling (M) and observational (O) evidence for all aspects of the mechanism. “Low” applies where only concepts are available. The level of quantification is considered “high” where quantification from observations and modeling agrees across multiple studies. “Low” is where studies are unavailable or inconsistent.

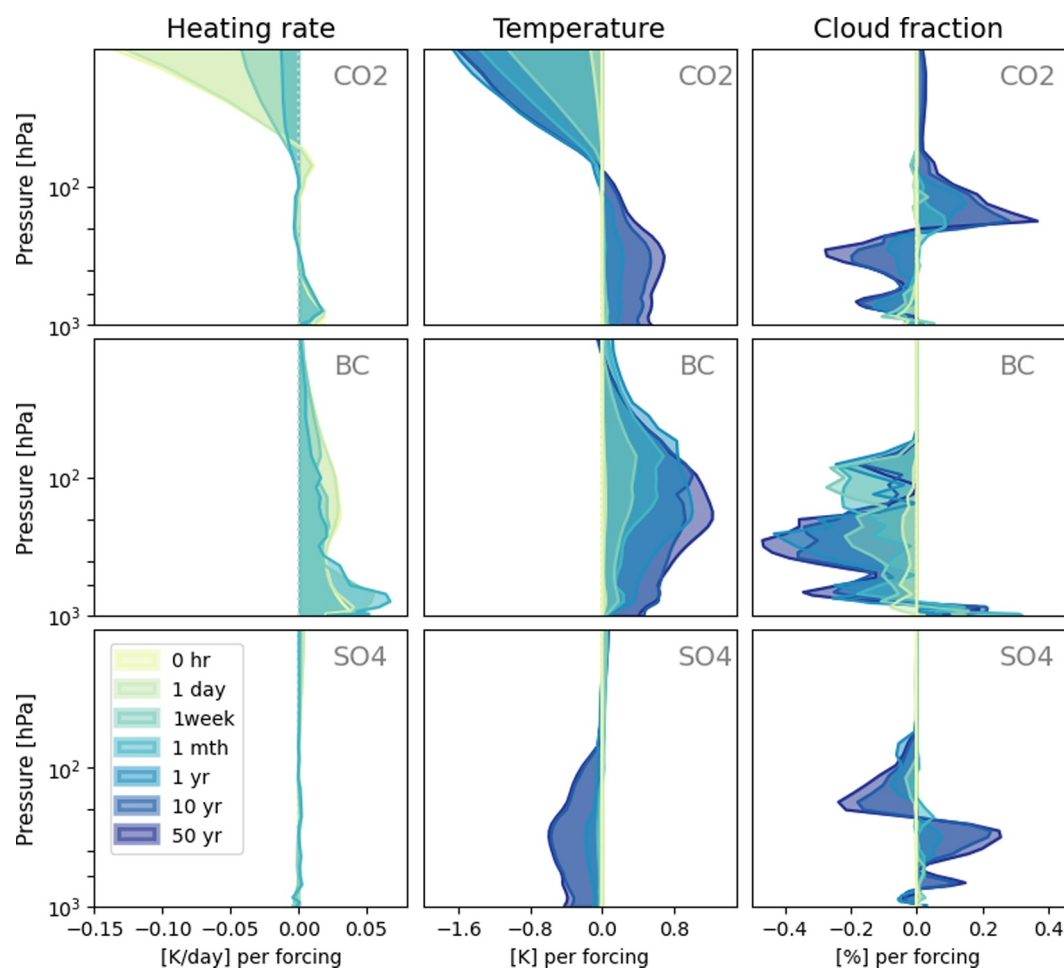


Figure 4. Vertical profiles of (left column) heating rates (in K day^{-1}), (middle column) temperature changes (K) and (right column) cloud fraction changes (%) in response to defined, sustained changes in three different climate drivers: increase in atmospheric carbon dioxide concentrations (CO_2 , top row), increase in atmospheric black carbon concentrations (BC, middle row) and increase in atmospheric sulfate concentrations (SO_4 , bottom row). Changes are normalized to the IRF imposed by the perturbation. Global-mean results averaged over the outcome from six climate models are shown for seven time horizons given by color code. Based on the results analyzed by Stjern, Forster, et al. (2023).

3.2. Vegetation Adjustments

Vegetation may respond to various climate drivers. Due to the impact of vegetation on albedo and on atmospheric humidity, as well as CO_2 concentrations, but also the emission of soil dust and BVOCs, this implies a radiative adjustment. Increases in atmospheric ozone concentrations may damage vegetation (Reich & Amundson, 1985) and thus reduce the vegetation effect on climate. Increases in aerosol lead to enhanced fraction of diffuse versus direct radiation, implying more efficient photosynthesis and thus reduction in atmospheric CO_2 concentrations (Mercado et al., 2009) with further consequences on albedo.

A further important adjustment process is due to the physiological response. As CO_2 concentrations increase, leaf stomatal conductance is reduced. This suppresses the amount of water transpiration that occurs when the plant absorbs the excess CO_2 (e.g., Leakey et al., 2009). In consequence, surface latent and sensible heat fluxes are affected, and thus boundary layer temperature and humidity, and subsequently cloud cover, each impacting the CO_2 ERF as adjustments (Andrews et al., 2012; Doutriaux-Boucher et al., 2009). Likewise, increased CO_2 fertilization by plants generally decreases surface albedo by increasing leaf area, leading to associated radiative changes (Bala et al., 2006; Zarakas et al., 2020). Increased CO_2 fertilization may also decrease dust emissions through changes in bare soil fraction, but there is no consensus across models on the sign of that effect (e.g., Thornhill, Collins, Olivie,

et al., 2021). Vegetation also responds to precipitation changes (Section 4), but such effects have not yet been investigated very much.

Both the reduction in stomatal conductance and the increased CO₂ fertilization are a direct consequence of the CO₂ concentration increase, independent of T_s change. However, physiological changes also induce land temperature change as the suppression of transpiration by stomata closure reduces evaporational cooling at the surface. This implies further adjustments as discussed below (Section 3.5).

Physiological changes to other forcing agents also may be considered (e.g., fertilization by dust, or vegetation changes in response to microphysical precipitation changes) but there is no evidence these are non-negligible in terms of their impact on the Earth energy budget. We note some of these physiological processes, such as changes in leaf area index or plant functional type, evolve over generations of plants and thus on time scales of years to decades.

3.3. Cloud and Precipitation Microphysical Adjustments to Aerosol Perturbations

In addition to their interaction with radiation (Section 3.1), aerosols serve as nuclei for the formation of cloud droplets and ice crystals. Increases in aerosol concentration typically result in increases in the cloud droplet number concentration (N_d ; Squires, 1952), which leads to an increase in cloud reflectivity (Twomey, 1974). Aerosol perturbations may also lead to changes in ice crystal number concentration, but only a small fraction of anthropogenic aerosols act as ice nucleating particles (Burrows et al., 2022; Kanji et al., 2020; Murray et al., 2012; Vergara-Temprado et al., 2018).

The changes to the hydrometeor size distribution in response to aerosol concentration changes and the resultant radiative effect is quasi-instantaneous and thus is the IRF due to aerosol-cloud interactions (IRFaci). It is noted that one could take the view that the IRFaci is an adjustment to the aerosol perturbation, but the community, including IPCC, has chosen to regard it as an IRF.

Changes to the hydrometeor size distributions produce changes in cloud microphysical processes and in turbulence. In liquid clouds, this includes a modification of precipitation formation (Albrecht, 1989), droplet sedimentation and evaporation (Ackerman et al., 2004; Bretherton et al., 2007) and turbulent entrainment (Small et al., 2009). These changes in processes lead to adjustments through modifications in cloud macrophysical properties, such as cloud fraction (Albrecht, 1989) and water path (Pincus & Baker, 1994). These adjustments act on a timescale of hours to days (Dagan et al., 2017; Glassmeier et al., 2021; Gryspeerd et al., 2021; Seifert et al., 2015). With strong dependence on cloud type, they exhibit large regional variation (Bellouin et al., 2020) and temperature dependence, with the potential to modify cloud feedbacks and climate sensitivity (Dagan, 2022a; Murray-Watson & Gryspeerd, 2022; Zhang et al., 2022).

Microphysical adjustments are hypothesized to be important also in mixed-phase and ice clouds (DeMott et al., 2010; Fan et al., 2013; Lohmann, 2002, 2017). The adjustments in mixed-phase and ice clouds can be caused by mechanisms different from the adjustment mechanisms in liquid water clouds. Anthropogenic INPs can lead to the glaciation of supercooled cloud droplets as INPs are needed for ice nucleation at temperatures warmer than about -36°C (DeMott et al., 2010). Glaciation can lead to changes in precipitation, cloud lifetime and radiative fluxes (Lohmann, 2002; Storelvmo et al., 2008). Anthropogenic INPs and associated cloud adjustments could also impact cloud phase feedback to global warming as higher INP concentrations would mean more ice clouds that can be transformed into liquid clouds with global warming (Murray et al., 2021). In addition, INPs could alter the properties of cirrus clouds (DeMott et al., 2010; Kärcher & Lohmann, 2003). Finally, additional CCN might considerably affect deep convective clouds, associated latent heating, precipitation and radiative fluxes through various mechanisms (Fan et al., 2013; Koren et al., 2010; Rosenfeld et al., 2008; E. Williams et al., 2002).

While the magnitude of IRFaci remains uncertain (Bellouin et al., 2020), the processes driving it have a moderate level of understanding and confirmation by observations. The adjustments to aerosol-cloud interactions are much less well understood. With a complex array of interacting processes (Stevens & Feingold, 2009), the magnitude (and in some cases the sign) of these adjustments is uncertain (Bellouin et al., 2020) and differs between observations and global models (Bender et al., 2019; Gryspeerd et al., 2020; Malavelle et al., 2017; Possner et al., 2020; R. Wood, 2007). Adjustments in mixed-phase and ice clouds are even less well understood and the

confidence in the magnitude of these adjustments remains low (Bellouin et al., 2020). Some quantitative assessments of cloud microphysical adjustments are discussed in Section 5.

3.4. Chemical Adjustments

Atmospheric chemistry plays a role when reactive compounds are emitted to the atmosphere. These emitted compounds may exert an IRF (e.g., halocarbons) or may be relatively radiatively inactive (e.g., nitrogen oxides; NO_x). In both cases, chemical adjustments following their emission alter the abundance of radiatively active trace gases (e.g., O_3) and lead to additional perturbations to the Earth's radiation balance.

Anthropogenic nitrous oxide (N_2O) emissions exert an IRF through absorption of terrestrial outgoing radiation (O'Connor et al., 2021), with both chemical adjustments—arising from a reduction in stratospheric O_3 and a shortening of the CH_4 lifetime—and cloud adjustments having consequences for the radiation budget (O'Connor et al., 2021; Szopa et al., 2021; Thornhill, Collins, Kramer, et al., 2021). A similar effect is found with halocarbon emissions (Szopa et al., 2021).

The ERF from anthropogenic CH_4 emissions is also strongly modulated by chemical adjustments (Szopa et al., 2021). CH_4 -driven increases to its own lifetime, tropospheric O_3 , and stratospheric water vapor contribute substantially to the CH_4 ERF (Myhre et al., 2013; Shindell et al., 2005). CH_4 lifetime and abundance are also modulated by chemical adjustments from other emitted compounds (Thornhill, Collins, Kramer, et al., 2021). For example, NO_x reduces CH_4 lifetime (resulting in a negative NO_x ERF overall, despite a positive tropospheric O_3 chemical adjustment to NO_x ; Szopa et al., 2021, their Figure 6.12), while emissions of carbon monoxide (CO) and volatile organic compounds (VOCs) increase CH_4 lifetime (contributing approximately 30% to their positive ERF; Szopa et al., 2021).

Anthropogenic changes in stratospheric and tropospheric O_3 arise from chemical adjustments. Of the resulting ERF, 45% are attributable to anthropogenic CH_4 , a third due to CO and VOCs, and a quarter due to NO_x (Stevenson et al., 2013), but with some offsetting from the impact of halocarbons on stratospheric O_3 and subsequent reduction in tropospheric O_3 (Szopa et al., 2021). Stratospheric ozone also responds to changes in incoming solar radiation, particularly at ultraviolet wavelengths, as a result of changes in photochemistry. This reduces the modeled surface warming impact of solar perturbations by about one third (Chiodo & Polvani, 2016).

Changes in tropospheric O_3 and hydroxyl (OH) radical production arising from anthropogenic emissions can also affect secondary aerosol production and aerosol size distributions. This may reduce the ERF from aerosol-cloud interactions by 20% (Karset et al., 2018). It may also offset the positive tropospheric O_3 chemical adjustment from anthropogenic NO_x (O'Connor et al., 2021) and change the sign of the cloud adjustment attributable to anthropogenic CH_4 from negative to positive (O'Connor et al., 2022).

3.5. Land Surface Temperature Adjustments

The IRF due to most climate drivers is spatially heterogeneous. Thus, even for unchanged global-mean near-surface temperatures, a pattern of temperature changes occurs. Beyond this, in practice, the community diagnoses ERF using simulations where sea surface temperatures are fixed (fixed-SST), but land temperatures are free to respond, largely due to technical hurdles in fixing surface temperatures over land (Forster et al., 2016). In such cases, the ERF is diagnosed from conditions where $\overline{\Delta T_s} \neq 0$, even if inconsistent with the formal definition. In particular over land, temperatures respond at rapid time scales. The land temperature change (ΔT_L) occurs within several days after the imposed perturbation (quantified at ~ 5 days by Dong et al. (2009)). Using high-temporal-resolution output from models contributing to the Precipitation Driver Response Model Intercomparison Project (Myhre et al., 2017), Stjern, Forster, et al. (2023) find the land begins to cool to an extent distinguishable from the inter-model variability within 1 day after a five-fold increase in SO_4 . After a doubling of CO_2 , the land begins to warm within just a few hours. These temperature changes imply non-negligible radiative adjustments.

The ΔT_L primarily influences the ERF by directly modifying longwave emissions (Planck effect or surface temperature in Figure 2). This direct effect can be substantial, accounting for nearly all of the tropospheric temperature adjustment for perturbations in CO_2 , CH_4 , and SO_4 (Smith, Kramer, et al., 2018). The ΔT_L also causes indirect radiative changes as the Earth system rapidly responds to the temperature change. These effects are not separable in coupled or fixed-SST simulations, so they have remained largely undiagnosed. Andrews et al. (2021) provide the most comprehensive attempt to date, isolating all ΔT_L -induced radiative effects using

simulations by Ackerley et al. (2018) with both SSTs and land temperatures fixed in a single general circulation model (following Shine et al. (2003)). In the model evaluated by Andrews et al. (2021), the ΔT_L -induced reduction in the CO₂ ERF manifests itself through increased emission of terrestrial radiation from the land surface and troposphere, reduced land albedo, increased moisture and through cloud adjustments associated with the land warming. When compared against traditional fixed-SST simulations, they find land warming reduces the ERF by 1 W m⁻² for a quadrupling of CO₂, or roughly 14% of the total ERF. This suggests a proportional underestimate of equilibrium climate sensitivity occurs when fixed-SST simulations are used to calculate the ERF, highlighting the importance of understanding and quantifying the radiative effects of ΔT_L . As a solution that allows for the freedom of the model to develop diurnal cycles and to respond to synoptic-scale weather variability, the temperature could be fixed at a soil level below the surface. Alternative approaches use a model's feedback parameter scaled by ΔT_L (Hansen et al., 2005), or use radiative kernels to subtract out portions of the adjustments that are likely to be ΔT_L related (Smith et al., 2020; Tang et al., 2019; Thornhill, Collins, Kramer, et al., 2021). However, simulations with fixed land temperatures provide the most accurate estimate of the ERF and we encourage additional groups to perform these experiments.

Some of the ΔT_L , and its associated radiative effects, however, occur due to local processes and are independent of a change in \bar{T}_s (i.e., at global-mean $\Delta \bar{T}_s = 0$). This component of the land surface temperature adjustment thus is a radiative adjustment in the strict sense. However, it has not yet been quantified in detail. Cloud adjustments arise from reduced stability and cloudiness over land (a positive adjustment) counteracted by an increase in lower tropospheric stability and low-altitude cloudiness over the oceans (negative adjustment). In total these cloud adjustments act to reduce the ERF, but the magnitude and even sign is model dependent. The contrasting changes over land and ocean suggest that the land warming causes land-sea circulation changes (see Section 3.6).

3.6. Dynamical Adjustments

Dynamical adjustments arise from spatial heterogeneity in surface and atmospheric temperature changes (third term on the right-hand side of Equation 2). Non-uniform aerosol forcing can cause changes in regional and global circulation, as several studies have shown (Bollasina et al., 2011; Chemke & Dagan, 2018; Ganguly et al., 2012; Johnson et al., 2019; G. Persad et al., 2023; Roeckner et al., 2006; Xie et al., 2022). For example, land surface cooling induced by aerosols can alter monsoon circulations even in a situation where SSTs are held fixed (Bollasina et al., 2011; Dong et al., 2015; Ganguly et al., 2012; Guo et al., 2013; Voigt et al., 2017). Adjustments to CO₂ also involve a land-sea contrast and a change in the monsoon circulations, which in some regions are opposite in sign to the changes driven by subsequent sea surface warming (Shaw & Voigt, 2015).

In addition, absorbing aerosols lead to atmospheric heating, which can modify large-scale atmospheric circulations (Johnson et al., 2019; Sand et al., 2020; T. Wood et al., 2020; Xie et al., 2020), with the perturbation's location being a crucial factor in this modulation (Dagan et al., 2019; A. I. Williams et al., 2022). Specifically, tropical atmospheric heating can cause direct thermally driven circulations because of the absence of a significant Coriolis force, while the response in the extra-tropics is more locally focused (Dagan et al., 2019). Furthermore, the strength of these large-scale circulation adjustments varies depending on the background circulation at the site of the added aerosols (A. I. Williams et al., 2022). Consequently, the magnitude and even sign of the generated ERF can be influenced by the perturbation's location (G. G. Persad & Caldeira, 2018; A. I. Williams et al., 2022). The potential implications of large-scale circulation adjustments for ACI are much more challenging to elucidate due to the wide range of scales involved and the need to consider microphysical processes while accounting for large-scale circulation. As a result, little is known about ACI-driven large-scale circulation adjustments. Nevertheless, idealized convective-permitting simulations that use parameterized (Abbott & Cronin, 2021; Dagan, 2022b) or directly resolved (Dagan et al., 2023) large-scale vertical velocity have shown that a local ACI perturbation could strengthen large-scale circulations and subsequently drive an increased cloudiness that results in stronger (more negative) ERF (Dagan et al., 2023).

Changes in the eddy-driven circulation, including the width of the Hadley cells, as well as the position and strength of the midlatitude jet and storm tracks and the southern hemisphere polar jet, are also subject to adjustments to forcing (Grise & Polvani, 2014, 2016; Morgenstern et al., 2020; Staten et al., 2014; T. Wood et al., 2020). For example, the warming effects by additional BC lead to a widening of the tropical belt and poleward shift of the midlatitude storm tracks, while additional scattering aerosol has the opposite effect (Johnson et al., 2019; T. Wood et al., 2020; Zhao et al., 2020). CO₂ adjustments typically manifest as a poleward expansion

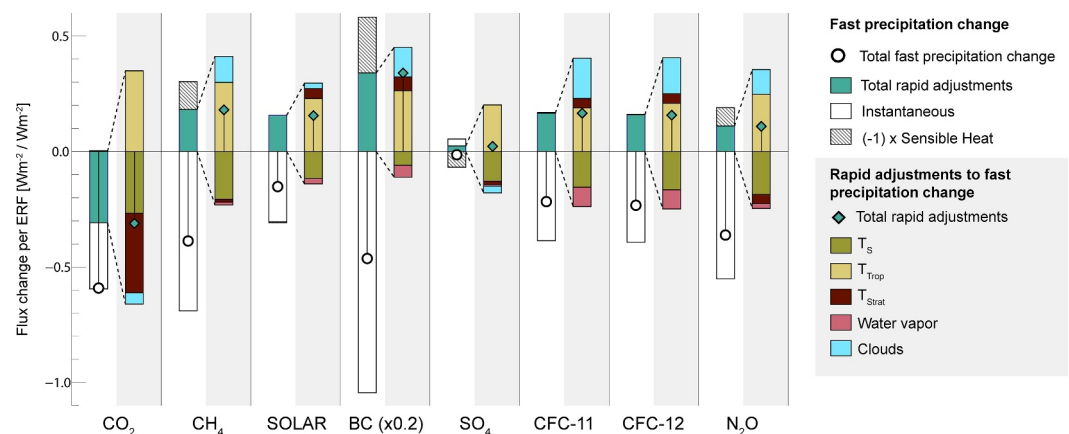


Figure 5. Fast precipitation changes in response to a range of climate forcings in energy flux density units. For each forcer, fast precipitation changes are broken down into contributions from instantaneous changes, sensible heat and adjustments (left bar for each forcer). The adjustments are further decomposed into contributions from surface (T_s), tropospheric (T_{Trop}) and stratospheric (T_{Strat}) temperature changes, water vapor and cloud changes (right bars). Based on Myhre et al. (2018) and Hodnebrog et al. (2020), who used multi-model simulations (Myhre et al., 2017) combined with radiative kernels. All numbers have been normalized by the multi-model mean ERF. BC has been further scaled by 0.2. The underlying experimental setups were $CO_2 \times 2$, $CH_4 \times 3$, insolation + 2% (SOLAR), BC $\times 10$, $SO_4 \times 2$, CFC-11 $\times 8$, CFC-12 $\times 9$, $N_2O \times 3$.

of the extratropical circulation (Deser & Phillips, 2009; Grise & Polvani, 2014; T. Wood et al., 2020). This is thought to be related to enhanced meridional temperature gradients in the upper troposphere, although the detailed mechanisms remain unclear (Staten et al., 2014). Perturbations in stratospheric water vapor and ozone also have the potential to induce adjustments in large-scale circulation through altering meridional temperature gradients in the upper troposphere-lower stratosphere (e.g., Charlesworth et al., 2023; Maycock et al., 2013; Thompson & Solomon, 2002) though these composition changes may themselves occur as part of climate feedbacks (e.g., Nowack et al., 2015) or as an indirect response following the introduction of a forcing agent; for example, CO_2 driven stratospheric cooling alters ozone photochemistry, leading to tropospheric and stratospheric circulation adjustments (Chiodo & Polvani, 2016).

4. Adjustments Beyond the Earth Radiation Budget

In the main framework (Section 2), adjustments are considered as a modulating influence on the Earth radiation budget. However, adjustments can also manifest as perturbations to other climate parameters, notably precipitation, circulation patterns and lower atmosphere turbulence without immediate consequences for the Earth radiation budget. The influence of adjustments on precipitation can occur via any mechanism that affects the local tropospheric energy balance, or the atmospheric circulation. Recently, this has been investigated with climate models, for a range of climate individual drivers, by using a combination of radiative kernels (e.g., Kramer et al., 2019; Soden et al., 2008) and idealized experiments with Earth System Models (e.g., Hodnebrog et al., 2020; Myhre et al., 2018; Smith, Kramer, et al., 2018).

Precipitation change in response to a forcing can usefully be split into two components—in response to surface temperature change, called slow precipitation change, and in response to the initial forcing but without a change in surface temperature, called fast precipitation change (Bony et al., 2013). While the slow changes have been found to scale closely with ERF, which includes the effect of radiative adjustments (Andrews et al., 2010; Samset et al., 2016), the fast precipitation changes are adjustment processes that do not involve top-of-atmosphere radiation changes. Their underlying mechanisms are diverse and driver-specific. Figure 5 shows multi-model estimates of fast precipitation response that have been broken down into the influence of instantaneous changes, changes to sensible heat, and adjustments. This is possible because any precipitation change is constrained by the atmospheric energy balance, following the energetic relation for steady-state conditions (O’Gorman, 2012; Pendergrass & Hartmann, 2014):

$$L\Delta P = \Delta Q - \Delta S \quad (3)$$

Here, L is the latent heat of condensation, P is precipitation, Q is atmospheric radiative flux divergence (in general, cooling by emission of terrestrial radiation), and S is sensible heat. In Figure 5, the fast precipitation responses ($L\Delta P_{\text{fast}}$) are normalized by the ERF of the model experiment, in order to make them comparable. Note that this normalization is imperfect as ΔP_{fast} does not generally scale with ERF, but it still serves to provide consistent signs and to remove some of the differences in forcing strength between the model experiments used. Further, the adjustment contribution to ΔP_{fast} has been expanded into its contributions from surface, tropospheric and stratospheric temperature change, cloud changes, water vapor and clouds. These components are also scaled by the ERF, for comparability. Surface albedo changes can also contribute, but are negligible for all the experiments shown here and is therefore not shown.

When normalized by ERF, all components have negative total ΔP_{fast} . Broadly, adjustments are driven by a positive contribution from tropospheric temperature and cloud changes, counteracted by negative influences from surface temperature and water vapor. CO_2 change is the only climate forcer where the total adjustments are negative, which can be seen to be primarily due to the strong negative contribution from stratospheric cooling. The mechanism by which stratospheric cooling acts to reduce precipitation on fast time scales is via alteration of the atmospheric energy budget (Allen & Ingram, 2002; Bony et al., 2013).

The influence of adjustments on precipitation response is particularly strong for black carbon aerosols (Sand et al., 2020). Note that in Figure 5, the ΔP_{fast} from BC has been scaled by 0.2. Myhre et al. (2018) showed how BC is the driver with the highest model diversity, while Samset (2022) pointed out how aerosol absorption of solar radiation is very poorly constrained in the most recent set of coordinated Earth system model simulations—both in absolute magnitude, and in terms of changes over the historical era. The geographical pattern of the BC perturbation plays a large role (Dagan et al., 2019; A. I. L. Williams et al., 2023). Increases of BC concentrations in the Tropics lead to enhancement of precipitation and also precipitation extremes (Dagan & Eytan, 2024), while increases in the extratropics reduce precipitation. This makes further investigations into both the instantaneous and adjustment influence of BC on precipitation a key topic for constraining projections of precipitation under future emissions and climatic changes.

Changes in greenhouse gas and aerosol concentrations can also influence turbulence in the planetary boundary layer (PBL) independent of sea surface temperature changes. The influence of BC on PBL dynamics and turbulence is strongly dependent on the altitude of added BC (Slater et al., 2022; Wang et al., 2018). In many cases, however, the change in the vertical temperature profile following BC-induced heating causes an increase in stability, which leads to suppressed turbulence and a shallower PBL (Ding et al., 2016). This link between absorbing BC aerosols and suppressed turbulence has been identified from observations (Wilcox et al., 2016), case-studies using large-eddy simulation (LES) and regional modeling (Slater et al., 2022; Wang et al., 2018), and recently using global modeling (Stjern, Hodnebrog, et al., 2023). Furthermore, dimming by absorbing aerosol causes the downward solar radiation at the surface to decrease, resulting in changes of surface energy balances and turbulent exchange fluxes. As shown in Senf et al. (2021), in addition to stabilizing the PBL, this effect can weaken turbulence and convective development, negatively affecting the formation of shallow convective clouds (Feingold et al., 2005). Recent model results indicate that the fast response of scattering sulfate aerosols on lower tropospheric turbulence is weak, while the response of increased CO_2 concentrations is to enhance near-surface turbulence over land, opposite to that of BC (Stjern, Hodnebrog, et al., 2023). Changes in turbulence could influence exchange processes between the Earth's surface and the atmosphere, and human health through the dilution of air pollutants.

5. Observations of Adjustments

Observing adjustments is not straightforward. Most drivers of forcing change gradually rather than abruptly. Also, it is challenging to separate the adjustments in the observations from the meteorological variability, and it is also challenging to disentangle adjustments from IRF as well as from feedbacks. The time dependence of adjustments decouples them from the initial causal factor. At longer timescales, feedbacks and forced change of the climate system often dominate observed change (a notable exception is the stratospheric cooling in response to CO_2 , see Section 5.4). Nevertheless, there are cases where adjustments have been observed, typically in response to a specific perturbation, where the timescale of the change can be clearly determined and the response—at least, approximately—isolated from other factors.

5.1. Response to Volcanic Eruptions

Volcanic eruptions are an external perturbation to the climate system and are often well-defined. Large eruptions may affect the stratosphere and provide clues about stratospheric adjustments. On average, the stratosphere maintains an approximate balance between emission and absorption of radiation (temperature profile of radiative equilibrium; Section 1). If the emissivity or absorptivity of the stratosphere changes, then radiative balance is shifted to a new temperature. Changes in stratospheric composition can therefore give rise to stratospheric temperature adjustments. These adjustments can occur in response to direct injections of radiatively active gases or aerosols to the stratosphere, changes in the radiation incident on the stratosphere, or in response to changes in stratospheric chemistry. For example, increasing tropospheric ozone cools the stratosphere due to reduced upwelling radiation at the tropopause. If the stratospheric temperature adjustment is not uniform in space, then it can change stratospheric circulation.

Several examples of stratospheric temperature adjustments have been observed. The 1982 El Chichón and 1991 Pinatubo volcanic eruptions produced a layer of sulfate aerosol in the stratosphere that increased the absorption of longwave and solar near-infrared radiation. This caused significant stratospheric warming and tropospheric cooling that was detected in radiosonde (Labitzke & McCormick, 1992) and satellite observations (Röck, 2000). In addition to aerosol absorption (Stenchikov et al., 1998), heating, heterogeneous reactions on aerosol and changes in circulation combined to produce reductions in stratospheric ozone (Grant et al., 1992). The observed tropical stratospheric cooling of about 4 K following the 2022 Hunga Tonga-Hunga Ha'apai underwater eruption has been attributed to the large injection of water vapor into the stratosphere due to the eruption (Schoeberl et al., 2022).

Also the cloud microphysical adjustments to volcanic aerosol emissions into the troposphere may be observed and may serve to improve process understanding and as constraints for models (Malavelle et al., 2017). These analyses suggest a small change in cloud liquid water path (Haghighatnasab et al., 2022; Malavelle et al., 2017), but potentially a strong adjustment of cloud horizontal cover (Chen et al., 2022).

5.2. Response to Point Sources of Aerosols

Some of the clearest observations of adjustments come from the microphysical adjustments to localized aerosol emissions. Aerosol sources such as ships, effusive volcanoes and industry can produce observable changes in N_d , along with accompanying changes in macrophysical cloud properties (sometimes termed “opportunistic experiments,” Christensen et al., 2022). Typically occurring as curvilinear polluted cloud features, the impact of the cloud microphysical adjustments are clear when compared to the relatively unpolluted clouds nearby. Increases in cloud height and decreases in precipitation have been observed in response to these aerosol perturbations (Christensen & Stephens, 2011), along with both increases and decreases in cloud water path (Gryspeerd et al., 2019; Toll et al., 2017, 2019; Yuan et al., 2023). The spatially limited nature of these perturbations also highlights the time evolution of these adjustments (Goren & Rosenfeld, 2012; Gryspeerd et al., 2021), evolving over hours to days. However, only a small fraction of the clouds polluted by shipping show visible ship tracks. It was recently shown that even in the case of ship track not distinguishable in the cloud microstructure, the clouds are affected by the ship emissions (Manshausen et al., 2022). This study found a positive adjustment of cloud liquid water path to increases in N_d , which may, however, be weak (Tippett et al., 2024).

Unique patterns of emissions also leave observable fingerprints of cloud microphysical adjustments at larger scales. The shipping corridor in the South-East Atlantic produces a clear change in N_d , but with less clear evidence of a change in cloud water. This implies a small role for cloud microphysical adjustments (Diamond et al., 2020). Enhanced lightning in the shipping corridors over the South China Sea and the northeastern Indian Ocean suggests a response of convective clouds to aerosol, with more vigorous convective activity at larger aerosol concentrations (Thornton et al., 2017).

5.3. Response to Wildfires

Emitting large amounts of absorbing aerosol, wildfires have the capability to alter atmospheric stability (Koch & Del Genio, 2010). Increases in smoke are correlated to reductions in cumulus cloud amount, through suppression of convective activity (Koren et al., 2004). In contrast, stability strengthening can increase cloud cover for stratocumulus clouds (Wilcox, 2010), expanding their cooling effect as a negative thermodynamic adjustment.

Recent extreme wildfires, such as the Australian pyroconvection events in 2019/2020, impressively demonstrated that local wildfire outbreaks can have global impacts (Senf et al., 2023; Yu et al., 2021). Much of the Australian smoke was emitted into the lower stratosphere, where it spread throughout the Southern Hemisphere after a short time (Ohneiser et al., 2020), leading to observable temperature increases in the stratosphere (Stocker et al., 2021) and dynamically driven adjustment processes (Kablick et al., 2020; Khaykin et al., 2020).

5.4. Trend Analysis of Recent Climate Change

Furthermore, anthropogenic emissions of greenhouse gases have increased the emissivity of the stratosphere and reduced the absorption of solar radiation by ozone, causing the stratosphere to cool. Significant long-term trends of stratospheric cooling have been detected in multi-decade satellite and radiosonde records (Maycock et al., 2018; Randel et al., 2009; Santer et al., 2023). The response of ozone concentrations to perturbations of the incoming solar radiation has been documented from satellite retrievals of the 11-year solar cycle (Dhomse et al., 2022; Hood et al., 2015; Maycock et al., 2016).

Short-lived climate forcers (SLCFs), such as aerosols and ozone, are spatio-temporally unevenly distributed. Thus, adjustments due to SLCFs are generally larger in regions and seasons with higher concentrations. It is expected that emissions of anthropogenic SLCFs will be further reduced as a result of air pollution control, but the progress of emission reductions much depends on the circumstances of each country. As countries are required to take action on climate change, a quantitative assessment of near-term climate change due to changes in emissions related to SLCFs by country or region is needed as a scientific basis. A quantification of adjustments is particularly important over such time scales. Since climate adjustments and feedbacks in a given country are not only due to emissions by this country and by a specific type of SLCFs, and since also impacts of specific emissions are not limited spatially, it is essential to consider multiple influencing factors to isolate adjustments attributable to specific emissions. There were attempts to analyze the microphysical adjustments to aerosol perturbations from the patterns of increasing versus decreasing aerosol emissions. The results are consistent with a small liquid water path adjustment but a positive relation between cloud fraction and drop number albeit with substantial uncertainty (e.g., Quaas et al., 2022).

6. Conclusions, Research Gaps, and Recommendations for Future Work

In the last decade, adjustments became an established part of the classical forcing—feedback framework to understand climate change from an energetic perspective. Substantial progress can be documented in terms of (a) analytical understanding, by classifying specific mechanisms by which adjustments act, (b) approaches to quantify adjustments across models and across the influences of specific state variable changes, and (c) the exploitation of observations to assess adjustment and constrain model estimates of these. An increasing number of studies aim for a more explicit analysis of adjustments that do not effect the Earth radiation budget immediately, but other components, like for example, precipitation fluxes.

Adjustments to changes in the Earth's orbit around the Sun, specifically the Milanković cycles that do not alter annual-, global-mean incoming solar radiation (changes to obliquity and precession of the orbit), are an evident manifestation of adjustments being relevant for large climate changes even in the absence of global instantaneous forcing.

Estimates of adjustments often come from idealized climate model simulations that employ large perturbations to, for example, greenhouse gas and aerosol concentrations, in order to improve signal-to-noise ratios. This leads to questions as to the linearity of adjustments and hence whether the size of adjustments found in such simulations are relevant for smaller, more realistic, forcings. Similar considerations apply to the additivity of adjustments when more than one climate forcing mechanism is present.

Climate model simulations can be used to understand the factors driving adjustments, to enable the development of more generic understanding of adjustment mechanisms (e.g., their dependence on the vertical profile of the instantaneous radiative heating perturbation), building on idealized studies including Hansen et al. (1997) and Salvi et al. (2021). An example of where this understanding would have great utility is in understanding the adjustments to halocarbon radiative forcing; there are a large number of halocarbon gases of potential climate importance, and it would be a challenge to accurately include these in climate model radiative transfer codes. To

date, adjustments have only been calculated for two CFCs (Hodnebrog et al., 2020) and the extent to which these adjustments are representative of a wider set of halocarbons is not known.

While this review has made a clear case that ERF is the preferred definition of forcing, the requirement to use climate models for the computation of adjustments does lead to some limitations. Climate model radiative transfer codes have limited spectral resolution, as well as requiring other approximations because of computational speed requirements; this necessitates continuing assessment of the quality of these codes for a wide range of forcings (including, separately, longwave and shortwave forcings) against benchmark radiation, via intercomparison projects such as the Radiative Forcing Model Intercomparison Project (Pincus et al., 2016). In addition, exploratory studies of new potential climate change mechanisms can be most easily done within an IRF or stratospheric adjusted radiative forcing framework using stand-alone radiative transfer codes, at the required spectral resolution, ahead of decisions on whether incorporation into climate model radiative codes is justified.

Adjustments to aerosol perturbations remain to be of particular importance. This is especially relevant for adjustments to aerosol-cloud interactions. Some aspects of these are especially poorly quantified such as the adjustments of deep convective clouds to aerosol-cloud interactions.

Adjustments are a sizable aspect in the forcing—feedback framework. Due to the typically rather short time scales on which they act, they lend themselves to constraints from observations more than feedbacks, despite the challenge in distinguishing IRF, adjustments and feedbacks in observations. Several examples for this are reported in this review. The analysis may investigate specific state variables (e.g., Figure 2) or specific adjustment mechanisms (Figure 1). In order to isolate, to better characterize, and to quantify specific mechanisms, one may analyze specific perturbations (as has been performed for aerosols), or one may use high-resolution, limited-area models thanks to the local nature in space and time of many adjustment processes (Nam et al., 2018). Model—data synergy arises when constraining models using observations, but also model studies may be used for a detection-attribution analysis by identifying fingerprints of adjustment mechanisms in models and assessing these in observations.

Conflict of Interest

The authors declare no conflicts of interest relevant to this study.

Data Availability Statement

No new data are produced for this review. The data for Figure 2 stems from previous publications by Hodnebrog et al. (2020), Marshall et al. (2020), and Skeie et al. (2020), and Smith, Kramer, et al. (2018). The data for Figure 3 were extracted from Doutriaux-Boucher et al. (2009), Zarakas et al. (2020), Mahowald (2011), Bellouin et al. (2020), Szopa et al. (2021), and Andrews et al. (2021). The data in Figure 4 are from Stjern, Forster, et al. (2023). The data for Figure 5, finally, are from Myhre et al. (2018) and Hodnebrog et al. (2020). For convenience, all data are tabulated and made available again in a concise way as Supporting Information S1.

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