Adjustments in the forcing-feedback framework for understanding climate change

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To: Editors, BAMS

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We hereby submit the first revision of the article “Adjustments in the forcing-feedback framework for understanding climate change”. The text length and figure number are unchanged from the previous draft. We attach a separate “response to reviewers” file.

Many thanks for considering this submission.

Sincerely,

Steven Sherwood

Adjustments in the forcing-feedback framework for understanding climate change

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ABSTRACT

The traditional forcing-feedback framework has provided an indispensable basis for discussing global climate changes. However, as analysis of model behavior has become more detailed, shortcomings and ambiguities in the framework have become more evident and physical effects unaccounted for by the traditional framework have become interesting. In particular, the new concept of adjustments, which are responses to forcings that are not mediated by the global mean temperature, has emerged. This concept, related to the older ones of climate efficacy and stratospheric adjustment, is a more physical way of capturing unique responses to specific forcings. We present a pedagogical review of the adjustment concept, why it is important, and how it can be used. The concept is particularly useful for aerosols, where it helps to organize what has become a complex array of forcing mechanisms. It also helps clarify issues around cloud and hydrological response, transient vs. equilibrium climate change, and geoengineering.

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More intensive analyses of climate simulations are revealing a need to revise definitions of forcing and feedback, and to recognize the new concept of rapid adjustments

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The traditional and now ubiquitous framework for understanding global climate change involves an external forcing, a response whereby the climate system opposes the forcing in order to regain equilibrium, and feedbacks which amplify or damp the response. The concept is most often applied to the global mean surface temperature $\overline{T}$, where the external forcing is a radiative perturbation (effective power input) $\mathrm{d}\overline{F}$, and $\mathrm{d}\overline{T}$ is the change in $\overline{T}$ produced by $\mathrm{d}\overline{F}$ at the new equilibrium[1]. This new equilibrium is achieved when the system response has caused a change $\mathrm{d}\overline{R}$ in net rate of energy loss by the planet that balances the effect of the imposed perturbation, i.e., so that $\overline{N} = \mathrm{d}\overline{F} - \mathrm{d}\overline{R} = 0$ where $\overline{N}$ is the net power into the planet from space. Feedback arises because there are various quantities X_i (atmospheric water vapor or sea ice cover for example) which depend on $\overline{T}$ and alter the planetary energy budget. The new equilibrium encompasses all of their effects as well:

$$\mathrm{d}\overline{R} = \mathrm{d}\overline{T}\left(\frac{\partial\overline{R}}{\partial\overline{T}} + \sum_i \frac{\partial\overline{R}}{\partial X_i}\frac{dX_i}{\mathrm{d}\overline{T}}\right) = \mathrm{d}\overline{F} \tag{1}$$

The ratio $(\mathrm{d}\overline{R}/\mathrm{d}\overline{T})^{-1}$ is called the "climate sensitivity parameter" (the "equilibrium climate sensitivity" being usually defined as $\mathrm{d}\overline{T}$ for a forcing equivalent to a doubling of CO_2). The term $(\partial\overline{R}/\partial\overline{T})$ is the "Planck response," or change that $\overline{R}$ would undergo if the climate system behaved as a black body with no feedbacks. The black-body system is stable to radiative perturbations because $(\partial\overline{R}/\partial\overline{T}) > 0$. This traditional approach is illustrated in

[1]Following the custom in climate literature, we use "equilibrium" in the loose sense of a system that is in a statistically steady state of energy balance. This is not a strict equilibrium since the Earth is constantly generating and exporting entropy.

Fig. 1b.

There are ambiguities across disciplines in what is meant by "feedback," a concept well known in electronics and control theory (Bates 2007). We will follow the usual custom of detailed climate feedback analysis studies (see Hansen et al. 1984; Schlesinger and Mitchell 1987; Soden and Held 2006) by referring to the amplifying role of a system property X_i, quantified by $\alpha_i \equiv -(\partial\overline{R}/\partial X_i)(\mathrm{d}X_i/\mathrm{d}\overline{T})$ as in (1), as a feedback. A feedback is "positive" if it amplifies the change in $\overline{T}$; in this case $\alpha_i > 0$. A system including feedbacks is stable if the sum of the α_i is smaller than the Planck response[2]. The forcing-feedback paradigm has helped establish, for example, the dominant role of water vapor in amplifying global temperature change and the role of clouds in accounting for its uncertainty (Cess 1990).

Many potential feedbacks within the Earth system can be conceived, involving system components having a wide range of characteristic response times (Dickinson and Schaudt 1998; Jarvis and Li 2011). Clouds and water vapor can respond to climate changes in days to weeks, whereas the deep ocean or ice sheets may require centuries or millennia. Feedbacks that are not fast enough to fully keep pace with responses of interest, due to the involvement of a slowly-varying component such as the deep ocean or ice sheets, may appear to have time-varying strengths (e.g. Senior and Mitchell 2000); those may be better treated as exogenous forcings in transient calculations. The "equilibrium" (sometimes called "Charney") climate sensitivity (ECS) has become the standard measure of the climate sensitivity of the Earth system relevant to the anthropogenic warming problem. It was adopted from early slab-ocean model experiments that were run to a less-complete equilibrium in which ice sheets, vegetation and atmospheric composition were all specified. The paradigm can however be extended to include, for example, changes to natural sources of CO_2 as "carbon cycle"

[2]Alternatively, the Planck response can also be considered as a strong negative feedback that is more direct and simpler than the others. This avoids giving it a special status and thus makes (1) more symmetrical (Gregory et al. 2009) in the sense that the system is stable if the sum of all the feedbacks, including the Planck response, is negative. However, use of the word "feedback" to describe the Planck response is confusing because there is nothing being "fed back upon."

feedbacks (e.g. Gregory et al. 2009; Arneth et al. 2010; Raes et al. 2010). Additional feedbacks not considered in the ECS will enter on longer (e.g., geologic) time scales.

In applying (1) to the global climate one normally assumes that all partial derivatives represent constants of the climate system, but this implies at least two bold assumptions. The first is that responses vary *linearly* with perturbation amplitudes (or equivalently, are not state dependent). Formally this must hold for sufficiently small perturbations, but possibly not for multiple doublings of CO_2 as some feedbacks may become stronger or weaker in significantly different climates (Crucifix 2006; Caballero and Huber 2013; Colman and McAvaney 2009).

The second assumption is that all responses are uniquely determined by the scalar $\mathrm{d}\overline{T}$ regardless of how the temperature change is brought about, a situation that may be called *fungibility*. Complete fungibility requires either that temperature changes always occur with the same spatial and seasonal pattern, or that different patterns produce the same $\mathrm{d}\overline{R}/\mathrm{d}\overline{T}$ where $\mathrm{d}\overline{R}$ is the change in global- and annual-mean radiation balance. However this will not be the case since different forcings will generally produce different warming patterns. Moreover, during transient warming regardless of how it is forced, some parts of the ocean may warm more slowly than others, temporarily producing anomalous warming patterns. In the absence of feedbacks these pattern differences should not strongly affect the paradigm, since the Planck non-linearity is sufficiently weak (Bates 2012). Many quantities X that affect the global radiation budget, however, are sensitive to spatial or seasonal variations in temperature. This sensitivity lies at the root of difficulties that have emerged over the years with the traditional framework (Hansen et al. 1997).

WHY ADJUSTMENTS? The radiative forcing concept is, in effect, a "common currency" that we may use to compare various types of perturbation: emissions of CO_2 or other pollutants, changes in land use, solar activity, etc. The concept is useful only to the extent that it accurately predicts the magnitude of the response without having to worry about any other details of the perturbation—that is, insofar as feedbacks are independent of the perturbation.

While one might imagine that the instantaneous impact of a perturbation on the top-of-atmosphere (TOA) radiation balance would be a good measure of its radiative forcing, early studies quickly recognized that this measure was not optimal. The temperature of the stratosphere, in particular, was not closely tied to that of the surface. For example it warms under a positive solar forcing, yet cools under a positive greenhouse gas forcing (Fels et al. 1980) therefore requiring the surface and troposphere to warm more to balance the same instantaneous TOA net flux perturbation (Hansen et al. 1997). This problem was resolved by allowing for a "stratospheric adjustment" prior to calculating the radiative forcing, which has been the standard approach at least since the first IPCC report (IPCC 1990).

Recent work reveals that heterogeneous responses also occur within the troposphere and can produce similar, but more subtle problems. Some of the most important mechanisms by which this can occur are illustrated in Fig. 2. For example, increasing the concentration of CO_2 in the atmosphere affects longwave radiative fluxes and slightly warms the mid- and lower troposphere, even with no surface temperature change (see Fig. 2b). In models this subtle change in stratification and relative humidity reduces middle and low-altitude cloud cover (Fig. 3), further altering the TOA net flux even before any global warming or cloud feedbacks take place (Andrews and Forster 2008; Gregory and Webb 2008; Colman and McAvaney 2011; Kamae and Watanabe 2012a; Wyant et al. 2012). This change in cloud cover is quite different to that which occurs subsequently due to the increase in $\overline{T}$(compare Fig. 3a and Fig. 3b). Likewise, any forcing that is horizontally inhomogeneous (for example changes in tropospheric ozone, or aerosols, discussed below) or that significantly affects the tropospheric radiative cooling will drive changes in atmospheric circulation that may alter the planetary albedo by changing patterns of cloud cover. Conceptually these complications are no different from the one long recognized for the stratosphere.

We refer to changes that occur directly due to the forcing, without mediation by the global-mean temperature, as "adjustments" and the accordingly modified top-of-atmosphere radiative imbalance as the "effective" radiative forcing, following Boucher et al. (2013). Their role is illustrated in Fig. 1a.

Most adjustments are rapid, but there is no fundamental time scale that separates rapid adjustments from feedback responses. The time scales of the two can, in principle, overlap significantly. Some adjustments can for example occur through state variables X that respond very slowly (e.g., vegetation cover or soil humidity responses to CO_2-induced stomatal closure (Doutriaux-Boucher et al. 2009) or aerosol-induced diffuse sunlight (Mercado et al. 2009)—see Fig. 2d—or responses involving stratospheric composition and chemistry). Meanwhile some feedbacks, such as the water-vapor feedback, can be triggered by warming of the land and atmosphere (Colman and McAvaney 2011) that occurs within days or weeks of an applied forcing. Indeed the original stratospheric adjustment requires several months to complete and has to be calculated by a special model run with the troposphere and surface held fixed. However, the largest tropospheric adjustments are likely due to changes in clouds driven by changes in tropospheric radiative fluxes, and appear to occur within days (Dong et al. 2009).

ADJUSTING OUR VIEW OF AEROSOLS It turns out that the climate community has been grappling with tropospheric adjustments for years, but without calling them by this name: they play a dominant role in the climatic impact of aerosols. One example is the "semi-direct effect" of aerosols, triggered by the uneven distribution of tropospheric radiative heating by the aerosol. This can subtly alter atmospheric stability which will affect convection (Fig. 2b), and because it is horizontally heterogeneous, it can drive circulations (Fig. 2a) that alter both the global cloud radiative effect and patterns of temperature and rainfall. This response should be regarded as a rapid adjustment to aerosol perturbations to the radiation field, since it occurs even in the absence of a change in $\overline{T}$.

Likewise, the cloud-mediated (or "indirect") impact of aerosols, which serve as cloud condensation nuclei (CCN), on climate involves rapid adjustments. This impact begins with an increase in the number of nucleated droplets which, in the absence of any changes to the water content or circulation of air within the cloud, would produce a cloud with higher albedo (often called the "Twomey effect"). Model studies indicate however that the knock-on effects that occur via changes in the flow field or the microphysical evolution of clouds can lead to final changes in albedo that differ substantially from this initial droplet number effect. A number of such knock-on effects have been articulated in the literature including "lifetime effect," liquid water path effect, etc. Most of these are based on idealized conceptual models, and their applicability to real clouds remains controversial (Boucher et al. 2013).

The initial and the various knock-on effects are often conceptualized as each having distinct physical significance, but in more realistic simulations it is typically not possible to distinguish them individually, and the assumptions under which they were deduced often do not hold. Only their combined effect can be properly diagnosed. We argue that the subsequent change in $\overline{T}$ is a response to this net radiative effect of aerosol—the aerosol effective radiative forcing or ERF, which includes the initial droplet-number effect and all adjustments. This concept is not new, and has been referred to in the literature before as a quasi-forcing (Rotstayn and Penner 2001) or a radiative flux perturbation (Lohmann et al. 2010).

In both the CO_2 and aerosol cases, adjustments are more uncertain than instantaneous forcings because they involve cloud and other dynamical responses that models may not calculate reliably. Regardless of this, models suggest these effects can be large, so they cannot be ignored.

WAYS OF DEFINING AND CALCULATING ADJUSTMENTS The ERF concept is motivated mainly by the desire to improve fungibility within the forcing-response framework, that is, minimize the quantitative differences of $\mathrm{d}\overline{T}$ to various types of forcing $\mathrm{d}\overline{F}$. Ideally we would choose an adjustment framework that optimises this, aiming for the ERF to be the forcing experienced by the system when $\mathrm{d}\overline{T} = 0$. There is however no unambiguous way to specify this, because regionally heterogeneous surface temperature changes occur immediately after a forcing is applied.

Two common approaches are available for quantifying the adjustment, with different advantages and disadvantages. The first or "regression method" (sometimes called Gregory method) is to regress the net TOA flux perturbation $\overline{N}$ onto $\mathrm{d}\overline{T}$ in a transient warming simulation, yielding a plot, (see Fig. 4) in which the $\mathrm{d}\overline{T} = 0$ intercept is the ERF (Gregory et al. 2004). The second or "fixed-SST" method (sometimes called the Hansen method) diagnoses ERF, $\mathrm{d}\overline{F}$ and $\overline{N}$ from a simulation including the forcing agent but with sea-surface temperatures and sea ice prescribed to their unperturbed climatology (Cess and Potter 1988; Hansen et al. 2005, see also Fig. 1c).

The regression method implicitly defines adjustments as those changes which occur relatively soon (within a few years), including those mediated by regional variations in SST change. The latter are excluded by the fixed-SST approach, which does on the other hand include all other forcing-related adjustments no matter how long they take to occur (provided the simulation is long enough). The regression method can be thrown off by time-varying feedbacks, in which case $\overline{N}$ versus $\mathrm{d}\overline{T}$ will not be a straight line. However this method satisfies the principle of cleanly separating adjustments from global mean temperature change $\mathrm{d}\overline{T}$, whereas the fixed-SST method permits land temperature change which contributes to $\mathrm{d}\overline{T}$ and affects the air-sea temperature difference over oceans (see Kamae and Watanabe 2012b; Shine et al. 2003; Vial et al. 2013). This enhances the global Planck response and triggers some warming-related changes such as an increase in global atmospheric water vapor (Colman and McAvaney 2011), the effects of which should be subtracted out if one wishes to isolate true adjustments from changes that result from feedbacks.

These two methods are shown for a typical CMIP model in Fig. 4. A third method that has been used in the literature for precipitation responses (examined further below) is to assume that the change during some limited time period (e.g., one year) following an abrupt forcing, compared to the climatology before, is due to adjustments. However, the $\mathrm{d}\overline{T}$ during this period is substantial, making it difficult to quantitatively compare with the other approaches.

The regression-based ERF estimate from a single simulation is inherently noisier than the fixed-SST one and is best suited for global-mean rather than regional responses. However, it can be made more precise by averaging across an ensemble of at least 5-10 shorter coupled simulation pairs of 10-20 years in which the step change in CO_2 from the control is made at different times to average over natural climate variability (e.g. Watanabe et al. 2012, and see next section). The fixed-SST ERF estimate is naturally more robust to internal climate variability because it takes advantage of the long averaging time, and the fact that the interannual ocean variability is either absent or identical in the perturbed and control simulations.

To the extent that the temperature-mediated response of the climate system is linear and invariant to the warming pattern, these methods should give almost identical results when the latter is corrected for the change in $d\overline{T}$ at fixed SST. As seen in Fig. 4, this is approximately true for global mean quantities, but there are noteworthy differences.

In all CMIP5 models for which the needed output has been published, the fixed-SST $4\times CO_2$ ERF exceeds the regression-based one, usually by a statistically significant margin. This point has been obscured because the literature has reported the former without the aforementioned $\sim$0.5 W m^{-2} feedback correction for land warming. Applying this correction to the seven CMIP5 models in Table 1 of Andrews et al. (2012) reporting both estimates, the fixed-SST ERF exceeds the regression ERF by about 15% (0.2-1.6 W m^{-2} at $4\times CO_2$, with a mean of 1 W m^{-2}). The HadGEM2-ES model exhibits a particularly large discrepancy due to a somewhat nonlinear response of $\overline{N}$ to the warming $d\overline{T}$ during the first year or two. Andrews et al. (2012) traced this response to an increase over time in cloud shortwave radiative feedback over oceans. Since this increase seems to occur in many models, it merits further study.

Watanabe et al. (2012) showed in one model how an ENSO-like SST anomaly can set up in the first year or two after CO_2 increase due to weakening of the Walker circulation (see also Bony et al. 2013); this is an example of an adjustment not captured by the fixed-

SST framework. In parts of the oceans with relatively shallow mixed layers the SST can respond more rapidly than in others, leading to the emergence of fast changes in SST patterns while the global $\mathrm{d}\overline{T}$ is still small (Armour et al. 2013). These changes influence cloud and circulation patterns (see next Section), amplify the atmospheric adjustments and can be aliased onto changes in global-mean cloud radiative effects in some models.

GCM-based estimates of the radiative forcing of anthropogenic aerosol on climate have often been based on comparison of fixed-SST simulations with pre-industrial and present-day aerosol emissions. These estimates are universally uncorrected for the associated change in global surface air temperature due to land temperature change; in principle an approximately $1\ \mathrm{W\ m^{-2}}$ correction per K of $\mathrm{d}\overline{T}$ should be added to them to be fully consistent with the ERF paradigm. However, at least for one model checked (CAM5), the surface air temperature change is less than 0.1 K, so this correction is negligible for most purposes.

In summary, ERF is a construct designed to fit the global radiative response of a model as a linear function of $\mathrm{d}\overline{T}$ over timescales of decades to a century. From this perspective, the regression ERF is preferable to the fixed-SST one since it is based on precisely the linear fit which is used for global feedback analysis, but this fit is imperfect, especially if applied to regional responses to $\mathrm{d}\overline{T}$ rather than global-mean ones. The difference in results between the two methods can be interpreted as an indicator of short-term deviations from linearity in the relation of $\overline{N}$ to $\mathrm{d}\overline{T}$. Such deviations seem to arise from the knock-on effects of inhomogeneous surface warming. The attribution of this to adjustment or feedback is inherently ambiguous, should depend on the circumstances and goals of the analysis, and will be different between the two methods considered here.

PRECIPITATION Rapid adjustments to $\overline{N}$ caused by CO_2 and aerosol are difficult if not impossible to detect in observations. We can however look for these physical effects in quantities other than the TOA radiative flux. Notably, we can consider the direct impact of a CO_2 change on precipitation, in the absence of any global-mean (or ocean-mean) $\overline{T}$ change. We should note however that because precipitation patterns are sensitive to small changes in the temperature pattern, we would expect regional precipitation changes to be relatively forcing-dependent even in the absence of adjustments—for example, a forcing that causes warming asymmetrically distributed between the hemispheres shifts tropical rain maxima toward the hemisphere of greater warming(Seo et al. 2014). Thus rapid adjustments alone may not explain all forcing-dependence of precipitation responses.

Possible adjustments of precipitation to aerosol perturbations (both radiative and cloud-microphysical) are now well recognized in principle, but poorly understood, hence controversial. For instance, by absorbing solar radiation, increased black carbon aerosols will cause a slight decrease in global-mean precipitation for the same surface temperature (Andrews et al. 2010). However, regional precipitation changes may be more important, and can occur far away from the aerosol that drives them (Wang 2013). Models suggest that, due to their heterogeneous heating of the atmosphere and surface, aerosol-radiation interactions can affect monsoons (Ramanathan et al. 2005; Lau et al. 2006), shift the inter-tropical convergence zone (Rotstayn and Lohmann 2002), and displace atmospheric jets poleward (Allen et al. 2012). Some of these studies have argued that these effects can be detected in observed rainfall trends. CCN-mediated effects on precipitation also have attracted great attention but are even more controversial (e.g. Tuttle and Carbone 2011; Tao et al. 2012).

Less recognized are the direct effects of solar or greenhouse-gas perturbations on precipitation. CO_2 warms and stabilizes the lower troposphere, slowing the global hydrological cycle for a given $\overline{T}$ (Allen and Ingram 2002; Andrews et al. 2010) and slowing and causing a redistribution of the tropical overturning circulation (Andrews et al. 2010; Wyant et al. 2012; Bony et al. 2013). The shifts in tropical rainfall associated with this effect make up a substantial part of the total circulation-driven rainfall change in climates simulated by the end of the 21st century (Bony et al. 2013). The change in global-mean rainfall is also nontrivial compared to that from warming. These effects on rainfall are somewhat more pronounced than those on TOA radiative balance, where adjustments appear to account for no more than 20% of global-mean $d\overline{T}$ in a multi-model average (though also contributing to forcing uncertainty, Forster et al. 2013; Vial et al. 2013). Much of the precipitation adjustment to CO_2 occurs very rapidly, within a week (Fig. 5). Thus, precipitation adjustments to CO_2 may stand a better chance of eventually being detectable in observations than would the TOA radiation adjustments.

Determining the spatial distribution of the precipitation adjustment is challenging because precipitation is highly variable on interannual and longer time scales, and sensitive to gradients in tropical sea surface temperature. Unforced anomalies that happen to occur after a step increase in forcing will be confounded with adjustments, necessitating an ensemble average to obtain the latter accurately from abrupt-forcing scenarios (Fig. 6). Moreover, atmospheric responses to forcings can quickly drive changes to the surface oceans, especially near the equator, that can strongly amplify or otherwise alter regional precipitation responses (compare panels a,b of Fig. 6; see also Chadwick et al. 2014). Such knock-on responses (which also affect top-of-atmosphere radiation) should be regarded as part of the adjustment to the extent that they involve SST gradients rather than the global mean $\overline{T}$, although again there is no unambiguous separation.

CONCLUSION In response to changing concentrations of CO_2 or other forcings, the climate system changes in ways that are independent of any global-mean surface temperature change, but which subsequently influence the global-mean radiation budget and hence surface temperature. These adjustments also appear to affect other climate quantities significantly, in particular precipitation. They are physically significant, depend on the forcing agent, and need to be accounted for when computing the radiative forcing of the agent. Many of them develop on a time scale of days (Cao et al. 2012; Kamae and Watanabe 2012a; Bony et al. 2013). Accounting more appropriately for adjustments offers new opportunities to better understand, predict, and evaluate impacts of different perturbations.

The fact that adjustments scale with the amplitude of the forcing rather than that of the global warming response means that even if global-mean temperature were for some reason very insensitive to forcing—due for example to some hypothetical strong negative feedback from clouds—the adjustments would remain unaffected. It also implies that part of the climate response to forcing is independent of the long-standing uncertainty in climate sensitivity.

While the forcing-feedback paradigm has always been recognised as imperfect, such discrepancies have previously been attributed to variations in "efficacy" (Hansen et al. 1984), which did not clarify their nature. Decomposing the climate response into a forcing-specific adjustment and a $\overline{T}$-mediated response that is more forcing-independent provides a clearer way of understanding climate changes, especially transient ones.

The adjustment concept needs to be fully integrated into energy budget studies (e.g. Otto et al. 2013). Estimates of radiative forcings and climate sensitivity ought to be defined consistently when observations are used to constrain estimates of the radiative forcings, climate sensitivity or both quantities. The traditional climate sensitivity is actually the product of two quantities, the radiative forcing of a doubling of CO_2 and a climate sensitivity parameter in units of K (W m^{-2})$^{-1}$, where it has been assumed that the former is known exactly. In fact it is not, especially when adjustments are considered as part of the forcing (see Webb et al. 2013; Stevens and Schwartz 2012). Because adjustments make the forcing uncertain, future studies should distinguish between the traditional climate sensitivity, which depends on adjustments, and the climate sensitivity parameter, which does not.

This decomposition may clarify some past reports of feedbacks appearing to be state-dependent, forcing-dependent or time-dependent, although not all such complexities are likely be resolved and some variations in efficacy will remain. Studies already show that transient climate changes at arbitrary times while the system is out of equilibrium, can be approximately recovered by adding the rapid adjustment to CO_2 at that time to the (lagging) temperature-mediated response (Andrews et al. 2010). To a considerable extent this also works for multi-model mean precipitation responses (Bony et al. 2013). This leads to a considerable simplification and will be useful to those exploring climate change using simple models that are only a function of global-mean radiative forcings (e.g., Huntingford and

Cox 2000), especially if such models explore scenarios (e.g., overshooting of carbon targets, ramping up and down of greenhouse gas forcings) that stray from those for which they have been fitted to the behavior of a more detailed climate model.

This approach also helps us to understand and anticipate the limitations of potential geo-engineering strategies. Idealized climate simulations of solar radiation management (Bala et al. 2008; Schmidt et al. 2012) show that when the warming due to CO_2 increase is counteracted by a decrease in the solar constant, the warming-induced impacts are reversed to a large extent but adjustment responses may linger. In the case of precipitation, the rapid adjustment to the higher CO_2 amount is not counteracted by a commensurate adjustment in response to lower solar radiation, leaving a net decrease in global-mean precipitation as well as larger residual changes at the regional level.

Indeed changes to solar radiation, volcanic aerosol, and orbital properties are each likely to lead to distinct adjustments to global rainfall and regional climate patterns in addition to their common impact on global-mean temperature. Recognizing and understanding the adjustments may be crucial in helping to make sense of both present and past climate changes. For instance accounting for adjustments helps interpret differences in the precipitation response to natural vs anthropogenic forcing (Liu et al. 2013).

Rapid adjustments involve rapid processes, and may present an opportunity to use model evaluations on very short time scales to constrain processes that are also important for longer-term climate change. Such systematic weather-forecast type of verification has been suggested as a possible way to improve the representation of model processes (Brown et al. 2012), but may also aid our understanding of the multiple adjustments associated with aerosol-cloud interactions (Boucher 2012) or the physical processes that control the direct effect of CO_2 on circulation and precipitation (Bony et al. 2013). Since rapid adjustments closely track the forcing variations in time, they may contribute substantially to initial transient climate changes even if making up a relatively small part of the long-term equilibrium response. This may offer an opportunity to better detect and disentangle the relative role of different forcings on climate, including on regional responses, and thus to facilitate detection and attribution studies. There are thus two broad reasons for future studies to distinguish more carefully between forcing-specific adjustments and temperature-driven feedback responses: to clarify our understanding of past climate changes, and to exploit what the relationships among various adjustment and feedback responses may be able to tell us about the climate system and how it will respond in the future.

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List of Figures

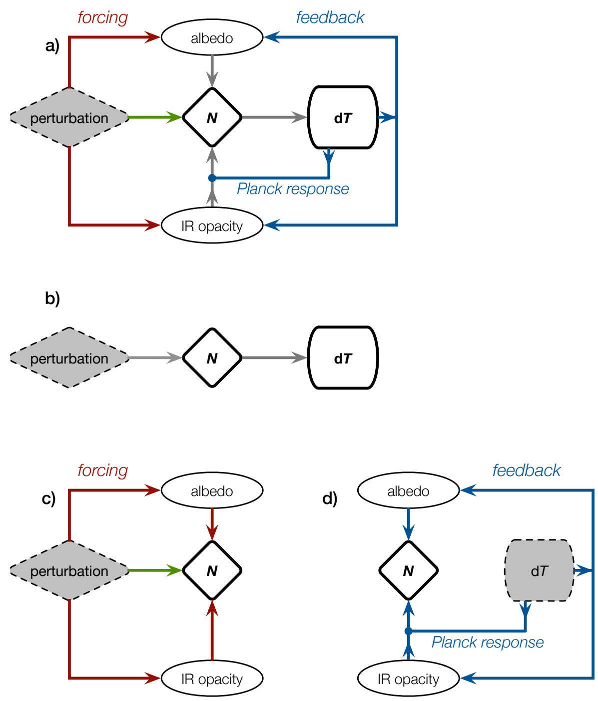

FIG. 1. Diagram showing forcing-feedback concepts for global temperature and methods of diagnosing them. (a) Full system, with shortwave albedo effects in upper part and longwave in lower part. Traditionally-defined forcing occurs via green arrow (in the case of solar forcing) or red arrows (other forcings) from perturbation to the top-of-atmosphere energy imbalance $\overline{N}$. Adjustments also occur via red arrows. Feedbacks occur via blue arrows, with the Planck response shown by the direct arrow from d$\overline{T}$ to $\overline{N}$. Feedbacks and adjustments can be diagnosed simultaneously by the regression method. (b) Traditional view of Planck system with no adjustments (nor feedbacks). (c-d) Reduced atmosphere-only system with fixed SST. Adjustments can be diagnosed by observing change in $\overline{N}$ after applying a perturbation with SST fixed (c); feedbacks can be diagnosed by observing changes in $\overline{N}$ after changing the SST with no (other) perturbation (d).

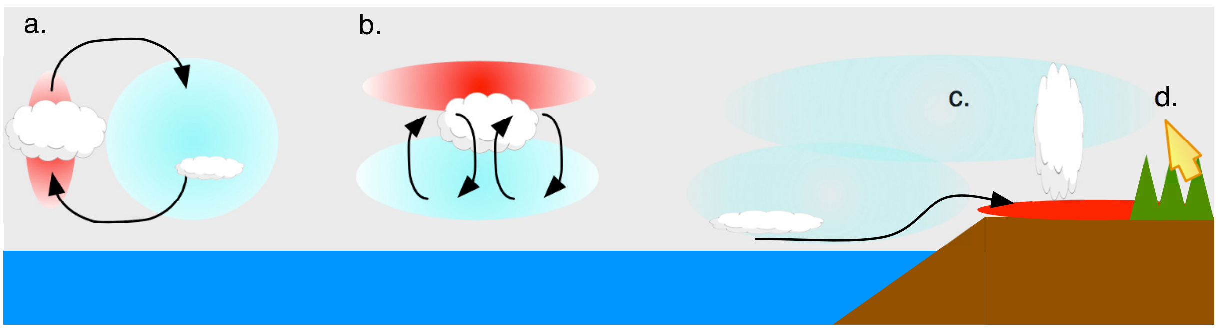

Fig. 2. Several examples of forcing adjustment mechanisms. (a) Solar, aerosol and greenhouse gas perturbations each can cause horizontal variations (red/blue regions) in net radiative heating of the atmosphere, which can drive circulations that alter cloud cover regionally and possibly change the global-mean radiative effect of clouds, modifying the conventional radiative forcings of these perturbations. (b) These perturbations can also cause vertical variations in the heating rate, altering atmospheric stratification, and affecting convection and local cloud development. (c) Perturbations may affect land and ocean surfaces differently, further affecting cloud cover. (d) CO_2 and aerosol perturbations can increase the growth of plants (affecting land albedo), or increase their water-use efficiency, affecting fluxes of water vapor (yellow arrow) and ultimately cloud cover.

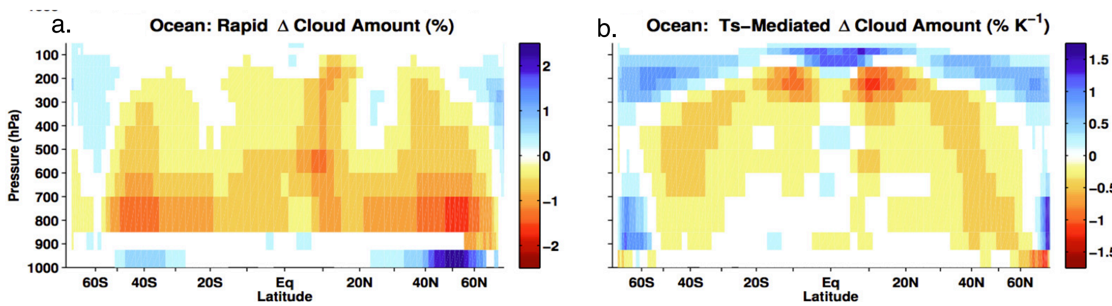

FIG. 3. Response of zonal-mean cloud cover over oceans to (a) a quadrupling of CO_2 (with warming effects removed), and (b) warming (with CO_2 effects removed), averaged over several GCMs, determined by regression method. Because cloudiness is generally reduced in both cases, these responses produce increased effective radiative forcing and positive cloud feedback, respectively, although the details of the cloud changes vary among models and can be seen here to differ significantly between the adjustment and feedback responses. Reproduced from Zelinka et al. (2013).

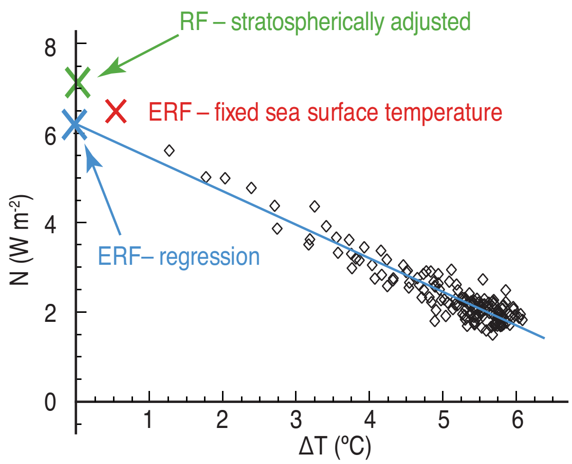

FIG. 4. Stratospherically-adjusted RF and ERF estimates by regression and fixed-SST methods using instantaneous $4{\times}CO_2$ experiments, for a typical CMIP5 climate model. N is the net radiative flux imbalance at the top of atmosphere and $d\overline{T}$ the global mean surface temperature change. The green cross gives the stratospherically-adjusted radiative forcing of CO_2 quadrupling, 7.1 W m^{-2} in this model. This is estimated as the instantaneous net downward radiative flux change at the tropopause when CO_2 is quadrupled. The black diamonds are annual means of the differences between the first 150 years of step- $4{\times}CO_2$ and control simulations. The blue line is the regression fit to these points. Its $d\overline{T} = 0$ intercept (blue cross) is the Gregory estimate of ERF (6.5 W m^{-2} in this model). The red cross is a 30 year mean difference of a pair of fixed sea surface temperature runs, one with standard CO_2 and one with quadrupled CO_2. To make it consistent with our basic definition, it needs to be adjusted to $d\overline{T} = 0$ by adding 0.5 W m^{-2} thereby removing the feedback contribution (dashed red line), giving a fixed-SST ERF estimate of 7.0 W m^{-2} for this model. Adapted from Fig. 7.2 of Boucher et al. (2013).

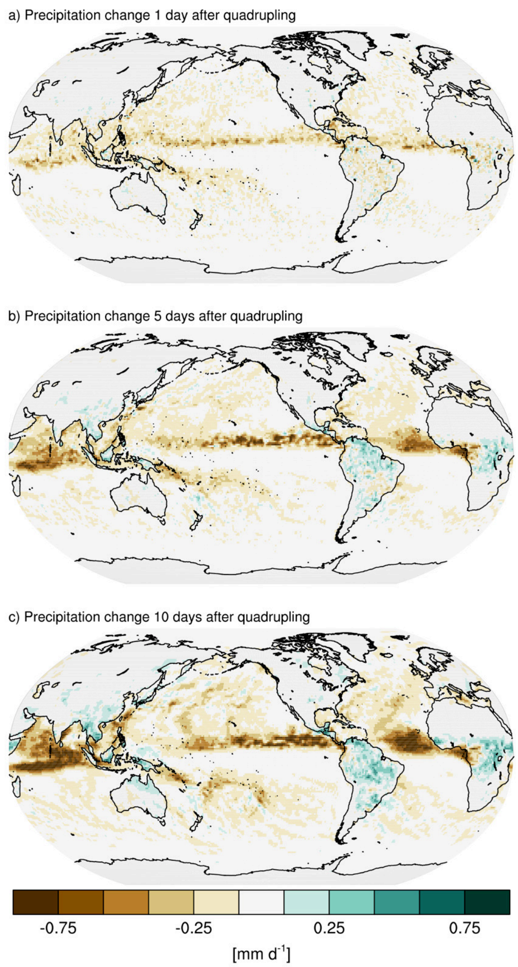

Fig. 5. Rapid development of the CO_2-induced precipitation response (units mm day^{-1}) as simulated by the ECMWF-IFS (Integrated Forecast System) model for October 2011 upon instantaneous quadrupling of CO_2. As CO2 increases, the reduction of the atmospheric radiative cooling warms the troposphere relative to the ocean surface, and warms the surface relative to the atmosphere over land. This adjustment response, which takes place within a few days, reduces precipitation over ocean but enhances it over most land areas. Adapted from Bony et al. (2013).

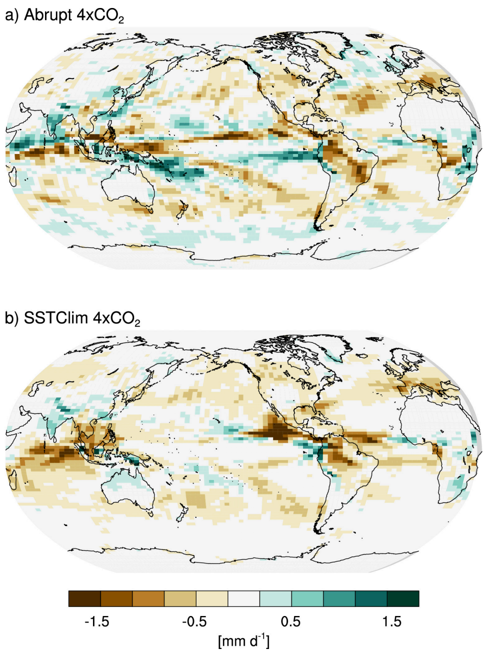

FIG. 6. Adjustment of precipitation to a quadrupling of CO_2 estimated two ways, using a single climate model (IPSL-CM5A). (a) Difference between first year after quadrupling and control climatology, averaged over an ensemble of 12 realizations (this result, which is similar to that obtained by regression method on the ensemble mean time series, shows strong regional influences in the Tropics from SST changes). (b) Change with SST held fixed everywhere, averaged over 12 years. Note scale has double the range of that in Fig. 5.

Response to comments of Reviewer #1.

We have addressed all minor editorial issues raised; below are the more substantive comments with our responses.

Below are some comments on specific wording and other details. In addition, I hope that the authors go over the paper again and identify the main points, express them more clearly, and summarize them in the conclusions.

We have tried to do this in the revision, thanks for the suggestion.

First of all, the title is misleading. It could easily be interpreted to mean "adjusting the framework". I suggest changing it to : "The role of (rapid) adjustments in the forcing-feedback framework"

We have chosen to keep the current title, which we do not think is misleading and was already the result of much discussion among the authors.

19-21 The last sentence includes concepts that are only marginally part of the paper. As such, I am not sure they merit being included in the abstract.

We have considered this but do not think they are marginal, and hope they come out a bit more strongly in the revision.

14 "which are direct responses to forcings that are not mediated by surface temperature changes" - global mean temperature is in the feedback definition, and used to calculate feedbacks, but arguably, feedbacks in the real climate system are mediated by temperature changes on a veriey of spatial scales.

We have tried to clarify this point in the revision. Clearly temperature changes heterogeneously in the real world; the question is whether changes specifically associated with that heterogeneity (and orthogonal to the mean) count as feedback responses or as direct consequences of the forcing. This Is somewhat ambiguous but there are principles that apply, as articulated in the revision.

35 surface snow cover - Why not mention sea ice? It's a stronger, more straightforward, feedback mechanism because the albedo change is larger.

Agreed, done

37 " .. called the "climate sensitivity". "Climate sensitivity parameter" is not a term used widely in the literature. I believe that in particular for the purpose of making the information contained in this paper accessible to a broad audience, it is best not to introduce terminology that is not standard.

We believe clarity is needed on this. The terminology is standard and our use is consistent with e.g. the IPCC AR5.

41-42 Rewrite this sentence. It sounds out of context, especially since the "new approaches" have not been specifically mentioned yet.
In the preceding sentence, you could reference Figure 1b. Write a sentence or two to describe and reference the other parts of the Figure as well.

We now postpone reference to “new approaches” and provide pointers to this figure later in the text (middle of “Why Adjustments?” and beginning of “Ways of Defining and Calculating Adjustments”).

123 Again, consider replacing global-mean temperature with surface temperature.

No, it is crucial that this be the global-mean temperature for the framework to be logically consistent. This relates to an earlier point that we hope we have clarified better.

131 You provide some specific examples for slow adjustment processes, an example or two of fast feedback processes relevant here would be helpful.

Done.

165 The subtitle "How should rapid adjustments be defined and diagnosed?" leads the reader to expect specific guidelines for calculating these quantities. However, the section does not provide this information, but rather a discussion of methods to perform these calculations and their shortcomings. Consider changing either the title or the contents of this section.

We have changed the subtitle.

171 What is the definition of a "transient simulation" you are using here? I assume that you are referring to a simulation where a forcing was applied *instantaneously in the beginning of the run, and the simulation traces the climate's response over time. However, "transient" is often also used when considering transient forcings that evolve with time. Be specific. It is my understanding, that the Gregory method is best applied to equilibrium, rather than transient (in the second sense of the term) model simulations. But I may be wrong.*

No, it is not essential that the forcing be applied instantaneously here, only that the simulation be transient in the usual sense: one where the system is out of equilibrium. We have ensured that all uses of the word "transient" consistently mean this and that if we specifically mean abrupt forcing application we say so.

238-253 This may be personal preference, but I would refrain from invoking the Gaia hypothesis, even if it is just by referencing Daisy world. Aside from being controversial, it is not a straightforward enough analogy to be useful to a broad audience.

We have eliminated this.

The last sentences of the paper should reiterate your main points. The current ending is rather weak. Try and end on a strong note.

Good advice, we have tried to do this.

Figure 3 - I understand this is taken from a different paper. It would still be useful, if the raw data is available to the authors, to plot the cloud changes using the same scale. This will highlight not only the spatial differences, but also the differences in magnitude. The Figure suggests the cloud changes are of roughly the same magnitude, but different distribution, unless one carefully looks at the scale.

The two panels have different units so it is not possible to put them on the same scale. The more important effect in most models when appropriately scaled is the temperature-induced one, which does appear bolder as drafted by Zelinka et al., so in this respect the figure is in no way misleading.

Figure 4 - Please remove the negative values of N from this Figure. The y-Axis can just as easily go from 0 to 8 without any loss of information. Removing the dashed 0 line however, will make the figure look cleaner. Also make sure the tick marks on the x-Axis are spaced the same as for the y-Ax.

Thanks, done.

Responses to Reviewer #2

I still wonder, though, if this message could have been formulated simpler. Maybe because I consider myself a more visual person, I like conceptional drawings (like Figure 1). But I struggled at times with complicated content of the text, especially when sentences where seemingly unnecessarily elongated.

We have tried to simplify the language in those places that may have caused problems for the reviewer.

minor comments

Figure 1

Nice conceptional illustration. Sill N is not explained (apparently the net input to the system from the text). I wonder if the direct arrow from the perturbation is needed in a) and c) as N is modified by pl.albedo and IR greenhouse? Should not also the link from N to dT in a) be left out?

No the link is needed to represent solar forcing. This is now explained in the caption, as is N.

Figure 2

Please add "uneven distributed" after a), processes for c) and d) are not that easily visually understood as a) or b) cane we improve?

Done w/respect to wording. No changes made to figure.

Figure 3

Both a) and b) have 'warming effects removed". If this correct, what is the case difference?

No one is warming-removed and one CO2-removed. Wording altered to reduce the chance of confusion here.

The (rapid) adjustment vs feedback is an interesting approach. I am not sure, however, that it makes it easier for aerosol. An adjustment component for aerosol is more difficult to quantify then a direct (radiative) effect (when separating old-fashioned in aerosol direct and the sum of all aerosol indirect effects). In order to include the semi-direct effect complicated modeling is required, which in part depends on less than perfect cloud parameterizations in modeling.

The framework is already being used for aerosols, all we have done is to point out the parallels with CO2 and proposed a clarification of the framework. The semi-direct effect does indeed require complicated modeling, but is automatically part of a GCM experiment with aerosols so it cannot be avoided, and is no more complicated than the adjustments to other forcings which also involve clouds.

I also do not think that the CCN/IN impact is not also affected by the semi-direct effect so any split will be artificially (as also the authors admit that timescales of adjustments and feedbacks can be 'comparable')

We have not split them, semi-direct effects associated with CCN are all part of the adjustment. Not sure what change the reviewer wants here.

Agreed, similar to the direct effect also the Twomey effect is one of many effect, which has been picked, because the ability to observe and easily simulate. I agree that "only the combined effect" matters … but what if such simulated effects are derived with models having deficiencies in cloud-representation and convection? (shouldn't we stick to something that can be understood?)

This, and the above comments indicate a philosophical difference. Such effects are already being examined, and models indicate they are sometimes of significant magnitude. We therefore need a framework for understanding them. The fact that they are uncertain is no reason to ignore them, when they are already present in comprehensive models and (implicitly) in observations.

Gregory's method is based on global averages. Can we apply such an averaging linear concept for uneven distributed responses? (… well the next page talks about this).

It is now stated more explicitly that the regression method may not be optimal for regional responses but can be used with appropriate ensemble averaging.

Can we be sure that adding absorbing aerosol decreases global precipitation? Any reference?

Reference now given.

Why is response in Figure 6 (twice the scale but similar colors) so much larger than in Figure 5?

First it is two different models; second, one is showing the response after only a few days while the other is the total response.

The last sentence (which I like) is a nice summary of what this contribution intends (… although the given examples for aerosol and precipitation come across as complex, so that better detection and attribution will remain an uphill battle).

Thanks we have moved this toward the end.

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