Understanding Climate Feedback Contributions to Surface Warming: TOA versus Surface Perspective

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Abstract

The global-mean surface temperature has warmed by approximately 1.04°C from 1880-2016, primarily driven by the anthropogenic increase of carbon dioxide (CO_2). Since Earths' temperature is tied to a multitude of physical processes, the increase of CO_2 triggers climate feedbacks that modulate the surface warming response. Thus, to understand the surface warming response to increasing CO_2, we must also understand how the different climate feedbacks it triggers modify the surface temperature. Most climate feedback studies evaluate radiative feedbacks using a top-of-atmosphere perspective, but a few use a surface perspective instead. The effects of radiative feedbacks on surface temperature should be insensitive to the perspective chosen; past studies, however, have shown conflicting results between the TOA and surface perspectives.

A comparison of the two perspectives indicates the largest disparity occurs in the interpretation of the temperature feedback; from a TOA perspective, it is the strongest negative feedback on the surface warming but the strongest positive feedback from a surface perspective. The lapse-rate feedback also displays contradicting effects on the surface warming between the two perspectives, but the contradiction is shown to stem from the contradiction in the temperature feedback. Furthermore, the lapse-rate feedback, as conventionally defined, is shown to be a correction term that adds no additional physical insight. Overall, differences in feedback attribution between the two perspectives are caused by atmospheric absorption. If the radiative feedback is negligibly affected by atmospheric absorption (e.g., albedo feedback), both perspectives will provide the same interpretation.

1. Introduction

The Intergovernmental Panel on Climate Change (IPCC) Assessment Report (AR) 5 states that the warming of the climate system is unequivocal [*IPCC,* 2013]. There is a clear globally averaged combined land and ocean surface warming of ~0.85^{o}C [0.65 to 1.06], calculated by a linear trend over the period 1880-2012 [*Hartmann et al.*, 2013], which if extended to 2016 is ~1.04^{o}C. Over the same time period there has also been a pronounced increase in well-mixed greenhouse gases, particularly carbon dioxide (CO_2), due to human activities [*Hartmann et al.*, 2013]. This anthropogenic increase in CO_2 is the primary driver of the surface warming, which is corroborated by simple climate models to complex coupled global climate models that have demonstrated that increasing the CO_2 concentration leads to a warming of the surface [*Manabe and Wetherald*, 1975; *Ramanathan et al.*, 1979; *Washington and Meehl*, 1984; *Schlesinger and Mitchell*, 1987; *Manabe et al.*, 1991; *Collins et al.,* 2013].

Increased CO_2 triggers not only an increase of surface temperature but also directly and indirectly influences many other climate variables through the complex interactions of the climate system. These perturbed climate variables, including surface temperature, feedback on each other leading to the observed or simulated response of the climate system to an increase of CO_2. A particular emphasis has been placed in the climate literature on understanding the surface warming response to an increase of CO_2, and thus understanding how the different climate feedbacks triggered by the CO_2 increase contribute to the surface temperature response. With this purpose in mind, many climate feedback analysis methods have been developed in an attempt to attribute and understand the contributions of individual climate feedbacks to the surface warming [*Wetherald and*

Manabe, 1988; *Cess et al.*, 1996; *Aires and Rossow*, 2003; *Gregory et al.*, 2004; *Soden et al.*, 2008; *Lu and Cai*, 2009a; *Lahellec and Dufresne*, 2014; *Sejas and Cai*, 2016]. The advantages and disadvantages of these methods have been previously discussed [*Aires and Rossow*, 2003; *Soden et al.*, 2004; *Stephens*, 2005; *Bony et al.*, 2006; *Bates*, 2007; *Cai and Lu*, 2009; *Klocke et al.*, 2013; *Lahellec and Dufresne*, 2013, 2014], but the focus of this study is on the perspective used by these methods to interpret the climate feedback effects on surface temperature.

The most commonly applied methods use a top-of-atmosphere (TOA) energy budget analysis to attribute the different climate feedback effects on surface temperature. The advantage of using a TOA point-of-view is that radiative processes dominate the TOA energy budget, so all that has to be analyzed is the insolation, outgoing solar radiation, and outgoing longwave (LW) radiation (OLR). The simplicity of the TOA perspective is also its limitation, as the feedback effects of sensible and latent heat fluxes, vertical convection, and dynamic transport on surface temperature remain hidden [*Cai and Lu*, 2009]. Therefore, to uncover the effects of dynamical feedbacks on surface temperature, more recent methods make use of the surface energy budget to analyze climate feedback effects on surface temperature (e.g., *Andrews et al.*, 2009; *Lu and Cai*, 2009b; *Sejas and Cai*, 2016). The advantage of the surface perspective is that in addition to radiative feedback effects, dynamical feedback effects on surface temperature can also be evaluated. Due to the extensive spatial coverage and reliability of recent satellite measurements of outgoing solar and LW radiation compared to surface measurements of radiative fluxes, latent and sensible heat fluxes, and dynamic transport, the TOA perspective is the favored method when performing observation-to-model feedback comparisons. The simplicity and utility of the TOA perspective has thus made it the preferred way to evaluate climate feedback contributions to the surface warming.

Regardless of perspective, we contend that the forcing and feedback analysis should give the same conclusions for the CO_2 forcing and all the different radiative feedback contributions to the surface warming. Unfortunately, this is not the case as there are some major discrepancies between the two perspectives. The meridional structure of the CO_2 forcing, for example, demonstrates a stark contrast between the two perspectives, as the surface perspective indicates a larger positive forcing in polar regions than tropics [*Lu and Cai*, 2010; *Cai and Tung*, 2012; *Taylor et al.*, 2013; *Song et al.*, 2014; *Sejas and Cai*, 2016], whereas the TOA perspective indicates a larger positive forcing in the tropics than polar regions [*Colman*, 2002; *Winton*, 2006; *Taylor et al.*, 2011a; *Cai and Tung*, 2012; *Pithan and Mauritsen*, 2014]. Previous studies also indicate the cloud feedback is minimum in polar regions from a TOA perspective [*Colman*, 2002; *Soden et al.*, 2008; *Taylor et al.*, 2011b; *Pithan and Mauritsen*, 2014] but maximum in polar regions from a surface perspective [*Taylor et al.*, 2013; *Pithan and Mauritsen*, 2014; *Song et al.*, 2014; *Sejas and Cai*, 2016]. The most dramatic discrepancy, however, is in relation to the temperature feedback, which is a strong negative feedback on surface temperature from a TOA perspective [*Wetherald and Manabe*, 1988; *Soden et al.*, 2008], but a strong positive feedback from a surface perspective [*Pithan and Mauritsen*, 2014; *Sejas and Cai*, 2016]. The goal of this study is thus to analyze and explain the differences between the results of the two perspectives, so as to reconcile some of the different conclusions given for the forcing and feedbacks in the climate literature.

2. Feedback Analysis

a. TOA perspective

Feedback analysis methods, such as the partial radiative perturbation [PRP; *Wetherald and Manabe*, 1988] and radiative kernel techniques [*Soden et al.*, 2008] have used a TOA perspective to analyze forcing and feedback contributions to surface warming. The TOA feedback analysis makes use of the perturbation of the TOA energy budget triggered by some external forcing,

$$\Delta \frac{\partial E}{\partial t} = \Delta S_{TOA} - \Delta R_{TOA} + \Delta Dyn\_trans \qquad (1),$$

where the term on the left is the change in the heat content tendency or heat storage rate below the TOA for a given grid point. ΔS_{TOA} is the change in net incoming solar or shortwave (SW) radiative flux, ΔR_{TOA} is the change in net outgoing longwave (LW) radiative flux, and ΔDyn_trans is the change in net heat transport into the column below the TOA by the atmosphere and ocean dynamics. The radiative perturbation is then assumed small enough to linearize,

$$\Delta\left(S_{TOA} - R_{TOA}\right) = \begin{bmatrix} \Delta\left(S_{TOA}^{ext} - R_{TOA}^{ext}\right) + \Delta\left(S_{TOA}^{wv} - R_{TOA}^{wv}\right) + \Delta\left(S_{TOA}^{cld} - R_{TOA}^{cld}\right) \\ + \Delta S_{TOA}^{alb} - \sum_{j=1}^{M} \frac{\partial R_{TOA}}{\partial T_j} \Delta T_j - \frac{\partial R_{TOA}}{\partial T_s} \Delta T_s \end{bmatrix} \qquad (2),$$

where the change in radiative flux has been decomposed into changes in radiative flux caused by the external forcing (*ext*), water vapor changes (*wv*), cloud changes (*cld*), surface albedo changes (*alb*), atmospheric temperature changes over *M* atmospheric layers, and surface temperature changes. Substituting (2) into (1) and rearranging the equation gives

$$\frac{\partial R_{TOA}}{\partial T_s}\Delta T_s = \Delta(S_{TOA}^{ext} - R_{TOA}^{ext}) + \Delta(S_{TOA}^{wv} - R_{TOA}^{wv}) + \Delta(S_{TOA}^{cld} - R_{TOA}^{cld}) + \Delta S_{TOA}^{alb} - \sum_{j=1}^{M}\frac{\partial R_{TOA}}{\partial T_j}\Delta T_j + \Delta Dyn\_trans - \Delta\frac{\partial E}{\partial t} \quad (3).$$

It follows from (3) that if the external forcing causes an increase in net energy flux into the climate system (positive value) the climate will warm. Thus, if any of the changes triggered by the external forcing causes the net energy flux into the climate system to further increase (positive value) the warming will be amplified and that process is said to be a positive feedback. However, if the physical process causes the net energy flux into the climate system to decrease, the warming will be suppressed and the process is said to be a negative feedback. As implied by (3), the warming of the climate system will manifest itself through the surface temperature response.

In the conventional TOA feedback analysis, it is also common for the atmospheric temperature change to be decomposed into an atmospheric temperature response equal to the surface temperature change plus the deviation from vertical uniformity,

$$-\sum_{j=1}^{M}\frac{\partial R_{TOA}}{\partial T_j}\Delta T_j = \left(-\sum_{j=1}^{M}\frac{\partial R_{TOA}}{\partial T_j}\right)\Delta T_s - \sum_{j=1}^{M}\frac{\partial R_{TOA}}{\partial T_j}\left(\Delta T_j - \Delta T_s\right) \quad (4).$$

After substituting (4) into (3) and rearranging, (3) becomes

$$\left(\sum_{j=1}^{M}\frac{\partial R_{TOA}}{\partial T_j} + \frac{\partial R_{TOA}}{\partial T_s}\right)\Delta T_s = \Delta(S_{TOA}^{ext} - R_{TOA}^{ext}) + \Delta(S_{TOA}^{wv} - R_{TOA}^{wv}) + \Delta(S_{TOA}^{cld} - R_{TOA}^{cld}) + \Delta S_{TOA}^{alb} - \sum_{j=1}^{M}\frac{\partial R_{TOA}}{\partial T_j}(\Delta T_j - \Delta T_s) + \Delta Dyn\_trans - \Delta\frac{\partial E}{\partial t} \quad (5),$$

where $-\sum_{j=1}^{M}\frac{\partial R_{TOA}}{\partial T_j}\left(\Delta T_j - \Delta T_s\right)$ is known as the lapse-rate feedback. Considering equilibrium conditions, the heat storage term disappears in (3) and (5), and the non-radiative heat transport term will vanish in (3) and (5) if taking a global-mean.

b. Surface perspective

While much less common in the climate literature, feedback analysis methods, have employed a surface perspective to analyze the contributions of forcing and feedbacks to the surface warming [*Andrews et al.*, 2009; *Lu and Cai*, 2009b; *Pithan and Mauritsen*, 2014; *Sejas and Cai*, 2016]. The surface perspective entails the use of the surface energy budget perturbed by the external forcing and feedbacks,

$$\Delta\left(\frac{\partial E_s}{\partial t}\right) = \Delta S_s - \Delta R_s + \Delta Q_s^{non\_rad} \quad (6),$$

where the terms are similar to their counterparts in (1), except everything is restricted to the surface layer. Following a linearization similar to that in (2) and after some rearrangement, (6) becomes

$$\frac{\partial R_s}{\partial T_s}\Delta T_s = \begin{bmatrix} \Delta\left(S_s^{ext} - R_s^{ext}\right) + \Delta\left(S_s^{wv} - R_s^{wv}\right) + \Delta\left(S_s^{cld} - R_s^{cld}\right) + \Delta S_s^{alb} \\ -\sum_{j=1}^{M} \frac{\partial R_s}{\partial T_j}\Delta T_j + \Delta Q_s^{non\_rad} - \Delta\left(\frac{\partial E_s}{\partial t}\right) \end{bmatrix} \quad (7).$$

Equation (7) states that the change in surface thermal emission is equal to the sum of the net radiative flux change due to the external forcing (*ext*), water vapor changes (*wv*), cloud changes (*cld*), surface albedo changes (*alb*), and atmospheric temperature changes plus the surface energy flux change due to non-radiative processes (e.g., surface latent heat flux) minus the change in heat storage rate. Therefore, if the external forcing causes an increase in net energy flux into the surface layer, (7) implies the surface will warm. If a feedback also increases the net energy flux into the surface layer (positive value) it will amplify the surface warming (i.e., positive feedback), but if it decreases the net energy flux into the surface layer (negative value) it will suppress the surface warming (i.e., negative feedback).

Unlike the conventional TOA feedback analysis, climate studies using the surface perspective do not decompose the atmospheric temperature change into that equal to the surface temperature response plus the deviation from vertical uniformity. However, for comparison sake, a decomposition similar to (4) can also be implemented in (7) and after rearrangement we obtain

$$\left(\sum_{j=1}^{M}\frac{\partial R_s}{\partial T_j}+\frac{\partial R_s}{\partial T_s}\right)\Delta T_s=\begin{bmatrix}\Delta\left(S_s^{ext}-R_s^{ext}\right)+\Delta\left(S_s^{wv}-R_s^{wv}\right)+\Delta\left(S_s^{cld}-R_s^{cld}\right)+\Delta S_s^{alb}\\-\sum_{j=1}^{M}\frac{\partial R_s}{\partial T_j}\left(\Delta T_j-\Delta T_s\right)+\Delta Q_s^{non-rad}-\Delta\left(\frac{\partial E_s}{\partial t}\right)\end{bmatrix} \quad (8),$$

where $-\sum_{j=1}^{M}\frac{\partial R_s}{\partial T_j}\left(\Delta T_j-\Delta T_s\right)$ can be considered the lapse-rate feedback from the surface perspective. Notice that while the heat storage term would disappear for equilibrium conditions, the non-radiative term in the surface perspective would not vanish in the global-mean.

3. Data and Analysis Procedures

Data derived from climate simulations of the NCAR CCSM4 are used in this study. The content of this section follows *Sejas et al.*, 2014, which used the same model simulations, and readers are urged to review that manuscript for additional details. Details important for this study are provided here.

a. Model description and simulations

The atmospheric component of the NCAR CCSM4 is the Community Atmospheric Model version 4 (CAM4) with a finite volume dynamic core, 1^{o} horizontal resolution, and 26 vertical levels. The ocean model is the Parallel Ocean Program version 2 (POP2) with 1^{o} horizontal resolution enhanced to 0.27^{o} in the equatorial region and 60 levels vertically. The CCSM4 also includes the Community Land Model version 4

(CLM4), and the Community Sea Ice Code version 4 (CICE4). Please see *Gent et al.* (2011) for more details.

In this study two model simulations are analyzed: (1) A pre-industrial control simulation and (2) a simulation with a 1% yr^{-1} increase in the CO_2 concentration. The CCSM4 pre-industrial control simulation runs for 1300 years holding all forcings constant at year 1850 levels, with a CO_2 concentration of 284.7 ppm. After year 200, the pre-industrial run reaches a quasi-equilibrium state as indicated by the small global mean temperature trend afterwards. Therefore, the 20-year mean between years 311 and 330 in the industrial control simulation is used to define the climatological annual cycle of the control climate simulation. The 1% yr^{-1} CO_2 increase simulation branches out at year 251 of the pre-industrial control simulation. In this transient simulation, the CO_2 increases 1% per year until the CO_2 concentration quadruples. The difference between the 20-year mean annual cycle centered at the time of CO_2 doubling, which corresponds to years 61-80 of the transient simulation (corresponding to the same 20-year span as the control run), and the climatological annual cycle of the control simulation is defined as the transient climate response to the CO_2 forcing (hereafter known as the transient response).

b. Analysis Procedures

The comparison between the TOA and surface perspective will concentrate only on the radiative feedbacks, since conventional TOA feedback studies have exclusively focused on radiative feedbacks. To obtain and isolate the radiative effects of the forcing and feedbacks on the TOA and surface energy budgets the Fu-Liou radiative transfer model [*Fu and Liou*, 1992, 1993] is used for all offline radiative flux calculations at each longitude-latitude grid point of the model using the 20-year monthly mean outputs from the control and transient climate simulations. The radiative flux change at the TOA and surface due to a specific process (e.g. water vapor change) is calculated by taking the perturbed 20-year monthly mean field of the process in question from the model output, with all other variables being held at their unperturbed 20-year monthly mean fields, and using these fields as input in our offline radiative flux calculations; then the unperturbed radiative flux is subtracted from the perturbed offline radiative flux giving the radiative flux change due to that process alone, consistent with the PRP approach. As an extension of the PRP approach similar to the radiative kernel technique, the partial derivatives in the above equations are obtained with the offline radiative transfer model by individually perturbing the temperature in each layer '*j*' by 1 K and calculating the perturbed radiative flux at the TOA and surface due to the 1 K increase of that specific layer alone; then as before the unperturbed radiative flux is subtracted from the perturbed offline radiative flux giving the approximate value of the partial derivative. The monthly-mean calculations are then zonally and annually averaged, from which the analysis follows.

4. Comparison of Radiative Feedbacks

a. Radiative feedbacks excluding the temperature or lapse-rate feedback

First, we analyze the changes in radiative flux at the TOA and surface due to the CO_2 forcing and feedbacks, excluding the temperature or lapse-rate feedback, which will be focused upon in the next subsection. The SW and LW effects of the forcing and feedbacks are separated to allow for a thorough comparison between the two perspectives. Focusing on the SW component first, it is clear the CO_2 forcing has no impact on the SW radiative flux (Fig. 1a, 2a). The surface albedo and SW cloud feedbacks have almost identical effects on the TOA and surface SW radiative flux (Figs.

1b-c, 2b-c; respectively). The SW water vapor feedback, however, is a positive feedback from the TOA perspective but a negative feedback from the surface perspective (Figs. 1d, 2d; respectively).

The contrast in effects between the surface and TOA perspectives for the SW water vapor feedback is due to the change in atmospheric SW absorption. There is a small increase in atmospheric SW absorption due to the projected increase in water vapor. The increase in SW absorption reduces both the SW radiation reaching the surface and the reflected SW radiation reaching the TOA. The upward SW flux reduction at the TOA implies surface warming due to the SW water vapor feedback, while the decrease of downward SW flux reaching the surface implies surface cooling.

The negligibly small change in SW absorption by the atmosphere is the reason the surface albedo and SW cloud feedbacks are nearly the same for both perspectives. The reduction of ice increases the SW absorption at the surface and reduces the reflected SW flux reaching the TOA, which implies a warming of the surface (positive feedbacks) from either perspective. A cloud increase (decrease) reduces (augments) the SW flux reaching the surface while enhancing (decreasing) the SW flux reflected back to the TOA; both of which imply a cooling (warming) of the surface.

Focusing on the LW component, the differences between the surface and TOA perspectives become more apparent. Though qualitatively the decrease of OLR and increase in downward LW flux at the surface both indicate surface warming due to the CO_2 forcing, the meridional differences between the TOA and surface perspectives are apparent (Figs. 3a, 4a). Consistent with other studies, the TOA radiative forcing is greater in the tropics than in polar regions implying the CO_2 forcing warms the tropics more than the poles, while the opposite is true from the surface perspective. The LW cloud feedback not only shows a meridional discrepancy between the two perspectives, but also has the opposite sign in parts of the tropics and mid-latitudes (Fig. 3c, 4c). The LW water vapor feedback is qualitatively more similar between the two perspectives, but meridional differences still exist. From the TOA perspective, the minimum is at the poles, while the surface perspective indicates the minimum water vapor feedback is located in mid-latitudes (Fig. 3d, 4d; respectively). The differences between the perspectives arise due to LW atmospheric absorption. Atmospheric absorption is thus the key reason why there are differences between the two perspectives. When atmospheric absorption is negligible both perspectives give very similar results.

b. Focus on Temperature and lapse-rate feedbacks

A special focus is placed on the temperature and lapse-rate feedbacks because they display the largest disparity between the surface and TOA perspectives (Fig. 5). From a surface perspective, the temperature feedback is a large positive feedback that substantially contributes to the surface warming (red solid line in Fig. 5a). On the other hand, from a TOA perspective the temperature feedback is the strongest negative feedback, representing the largest suppressor of the surface warming (blue solid line in Fig. 5a). This stark difference is problematic, as the effects of the temperature feedback on surface temperature should not depend on perspective.

The contradiction is a consequence of the effects of atmospheric temperature on the upward and downward LW fluxes. The general warming of the atmosphere in response to a CO_2 increase causes the LW emission to increase in both the upward and downward direction. This causes an increase in the both the OLR and downward LW radiation reaching the surface. The OLR increase implies a loss of energy to space (i.e., negative feedback), and thus a cooling of the surface from the TOA perspective. On the contrary, the downward LW flux increase implies greater LW absorption by the surface, and thus a warming of the surface (i.e., a positive feedback).

The large disparity in the lapse-rate feedback actually stems from the aforementioned contradiction in the temperature feedback. Consistent with previous studies [*Colman*, 2002; *Taylor et al.*, 2011a; *Pithan and Mauritsen*, 2014], the lapse-rate feedback is a negative feedback in the tropics and a positive feedback in polar regions from a TOA perspective (Fig. 5b; blue curve). However, from a surface perspective, the opposite is true as the lapse-rate feedback is positive in the tropics and negative in polar regions (Fig. 5b; red curve). As indicated by (4), the lapse-rate feedback arises as a result of the subjective decomposition of the temperature feedback into a uniform temperature response equaling the surface temperature change and the deviation from vertical uniformity (i.e., the lapse-rate feedback). The lapse-rate feedback is therefore just the difference between the temperature feedback and uniform temperature response and thus a correction term to the assumption that the atmospheric temperature change is equal to the surface temperature change. Since the surface warms, a vertically uniform temperature response implies the atmosphere warms as well, which increases the OLR and downward LW flux at the surface (Fig. 5a; dashed curves). By assuming a vertically uniform temperature response equal to the surface temperature change, the increase in atmospheric emission has been underestimated in the tropics since the surface warming is less than the atmospheric warming in the tropics, but overestimated in polar regions since the polar surface warming is greater than the atmospheric warming. The lapse-rate feedback therefore corrects for this by "causing" an increase in OLR and surface downward LW flux in the tropics and a decrease in OLR and surface downward LW flux in polar regions. This demonstrates that the contradiction in interpretation between the two perspectives originates from the contradiction in the temperature feedback, namely that an increase in atmospheric temperature causes an increase in both the OLR at the TOA (i.e., a negative feedback) and the downward thermal radiative flux at the surface (i.e., a positive feedback).

5. Discussion

The conventional TOA feedback analysis implicitly assumes that a decrease (increase) of OLR or upward SW radiation at the TOA, due to the CO_2 forcing or feedbacks, will be balanced by an increase (decrease) in OLR caused by an increase (decrease) of surface temperature. This assumption, however, is strictly true only if there were no atmosphere (i.e., the TOA and surface are the same). The presence of an atmosphere means changes in OLR and upward SW radiation at the TOA will be balanced by changes in OLR caused by *both* atmospheric and surface temperature changes. A reduction (increase) in energy loss to space at the TOA therefore does not necessarily imply a surface warming (cooling); instead it implies a warming (cooling) of the atmosphere, surface, or both. This also explains why differences in feedback attribution between the two perspectives only occur for radiative feedbacks influenced by atmospheric absorption. If the radiative feedback does not change the atmospheric absorption, it will not warm or cool the atmosphere, and thus the changes in OLR must be balanced exclusively by changes in surface temperature (matching the surface perspective). However, if the radiative feedback does change the atmospheric absorption, it will warm or cool the atmosphere, and the changes in OLR will not be exclusively balanced by changes in surface temperature. Even if we assume that changes in TOA radiative flux are primarily balanced by changes in OLR due to surface temperature changes, the TOA perspective is at best a first order approximation of the effects of radiative forcing and feedbacks on surface temperature. On the other hand, changes in surface energy flux (radiative or dynamical in nature) will be balanced by changes in upward LW flux due to surface temperature changes.

6. Summary and Conclusions

An increase of CO_2 warms the surface and triggers climate feedbacks that amplify or suppress the surface warming. Evaluating the relative contributions of the forcing and feedbacks in establishing the surface warming response in model projections is important to further our understanding of the projected warming. Traditionally, feedback analyses have employed a TOA perspective to evaluate these contributions, but these analyses exclude an explicit evaluation of non-radiative feedbacks. As a result, more recent studies have used a surface perspective, which include the evaluation of non-radiative feedbacks effects on surface temperature in addition to radiative feedbacks. Radiative feedback effects on surface temperature, however, should be independent of perspective.

Unfortunately, the two perspectives provide conflicting results. The root of the problem is the implicit assumption in the TOA perspective that changes in outgoing energy flux at the TOA are balanced exclusively by changes in surface temperature, when in reality atmospheric temperature changes also play a role. The two perspectives are therefore only in accordance when atmospheric absorption is inert to or affects the radiative feedback being analyzed minimally (e.g., surface albedo feedback). The TOA perspective is therefore at best a first order approximation of the radiative feedback effects on surface temperature. Furthermore, the lapse-rate feedback, commonly cited as the largest suppressor of the surface warming from a TOA perspective [*Bony et al.*, 2006], is a correction term that indicates how much the OLR has been overestimated or underestimated by assuming the atmospheric warming is equal to the surface warming (i.e., the uniform temperature response) and thus provides no additional physical insight as conventionally defined. The surface perspective therefore provides a better and direct representation of the effects of radiative feedbacks on surface temperature with the added benefit of including dynamical feedbacks as well. The TOA perspective can still be used to understand the surface temperature response, but only for radiative feedbacks that are negligibly affected by atmospheric absorption (such as surface albedo and cloud SW feedbacks).

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References

Aires, F., and W. B. Rossow (2003), Inferring instantaneous, multivariate and nonlinear sensitivities for the analysis of feedback processes in a dynamical system: Lorenz model case-study, *Q. J. R. Meteorol. Soc.*, *129*(587), 239–275, doi:10.1256/qj.01.174.

Andrews, T., P. M. Forster, and J. M. Gregory (2009), A Surface Energy Perspective on Climate Change, *J. Clim.*, *22*(10), 2557–2570, doi:10.1175/2008JCLI2759.1.

Bates, J. R. (2007), Some considerations of the concept of climate feedback, *Q. J. R. Meteorol. Soc.*, *133*(624), 545–560, doi:10.1002/qj.62.

Bony, S., R. Colman, V. M. Kattsov, R. P. Allan, C. S. Bretherton, J.-L. Dufresne, A. Hall, S. Hallegatte, M. M. Holland, and W. Ingram (2006), How well do we understand and evaluate climate change feedback processes?, *J. Clim.*, *19*(15), 3445–3482.

Cai, M., and J. Lu (2009), A new framework for isolating individual feedback processes in coupled general circulation climate models. Part II: Method demonstrations and comparisons, *Clim. Dyn.*, *32*(6), 887–900, doi:10.1007/s00382-008-0424-4.

Cai, M., and K.-K. Tung (2012), Robustness of Dynamical Feedbacks from Radiative Forcing: 2% Solar versus 2 × CO_2 Experiments in an Idealized GCM, *J. Atmospheric Sci.*, *69*(7), 2256–2271, doi:10.1175/JAS-D-11-0117.1.

Cess, R. D. et al. (1996), Cloud feedback in atmospheric general circulation models: An update, *J. Geophys. Res. Atmospheres*, *101*(D8), 12791–12794, doi:10.1029/96JD00822.

Colman, R. (2002), Geographical contributions to global climate sensitivity in a general circulation model, *Glob. Planet. Change*, *32*(2), 211–243.

Fu, Q., and K. N. Liou (1992), On the Correlated *k* -Distribution Method for Radiative Transfer in Nonhomogeneous Atmospheres, *J. Atmospheric Sci.*, *49*(22), 2139–2156.

Fu, Q., and K. N. Liou (1993), Parameterization of the Radiative Properties of Cirrus Clouds, *J. Atmospheric Sci.*, *50*(13), 2008–2025.

Gent, P. R. et al. (2011), The Community Climate System Model Version 4, *J. Clim.*, *24*(19), 4973–4991, doi:10.1175/2011JCLI4083.1.

Gregory, J. M., W. J. Ingram, M. A. Palmer, G. S. Jones, P. A. Stott, R. B. Thorpe, J. A. Lowe, T. C. Johns, and K. D. Williams (2004), A new method for diagnosing radiative forcing and climate sensitivity, *Geophys. Res. Lett.*, *31*(3), L03205, doi:10.1029/2003GL018747.

Klocke, D., J. Quaas, and B. Stevens (2013), Assessment of different metrics for physical climate feedbacks, *Clim. Dyn.*, *41*(5–6), 1173–1185, doi:10.1007/s00382-013-1757-1.

Lahellec, A., and J.-L. Dufresne (2013), A Formal Analysis of the Feedback Concept in Climate Models. Part I: Exclusive and Inclusive Feedback Analyses, *J. Atmospheric Sci.*, *70*(12), 3940–3958, doi:10.1175/JAS-D-12-0218.1.

Lahellec, A., and J.-L. Dufresne (2014), A Formal Analysis of the Feedback Concept in Climate Models. Part II: Tangent Linear Systems in GCMs, *J. Atmospheric Sci.*, *71*(9), 3350–3375, doi:10.1175/JAS-D-13-0334.1.

Lu, J., and M. Cai (2009a), A new framework for isolating individual feedback processes in coupled general circulation climate models. Part I: formulation, *Clim. Dyn.*, *32*(6), 873–885, doi:10.1007/s00382-008-0425-3.

Lu, J., and M. Cai (2009b), Seasonality of polar surface warming amplification in climate simulations, *Geophys. Res. Lett.*, *36*(16), L16704, doi:10.1029/2009GL040133.

Lu, J., and M. Cai (2010), Quantifying contributions to polar warming amplification in an idealized coupled general circulation model, *Clim. Dyn.*, *34*(5), 669–687, doi:10.1007/s00382-009-0673-x.

Manabe, S., and R. T. Wetherald (1975), The Effects of Doubling the CO_2 Concentration on the climate of a General Circulation Model, *J. Atmospheric Sci.*, *32*(1), 3–15, doi:10.1175/1520-0469(1975)032<0003:TEODTC>2.0.CO;2.

Manabe, S., R. J. Stouffer, M. J. Spelman, and K. Bryan (1991), Transient Responses of a Coupled Ocean–Atmosphere Model to Gradual Changes of Atmospheric CO2. Part I. Annual Mean Response, *J. Clim.*, *4*(8), 785–818, doi:10.1175/1520-0442(1991)004<0785:TROACO>2.0.CO;2.

Pithan, F., and T. Mauritsen (2014), Arctic amplification dominated by temperature feedbacks in contemporary climate models, *Nat. Geosci.*, *7*(3), 181–184, doi:10.1038/ngeo2071.

Ramanathan, V., M. S. Lian, and R. D. Cess (1979), Increased atmospheric CO2: Zonal and seasonal estimates of the effect on the radiation energy balance and surface temperature, *J. Geophys. Res. Oceans*, *84*(C8), 4949–4958, doi:10.1029/JC084iC08p04949.

Schlesinger, M. E., and J. F. B. Mitchell (1987), Climate model simulations of the equilibrium climatic response to increased carbon dioxide, *Rev. Geophys.*, *25*(4), 760–798, doi:10.1029/RG025i004p00760.

Sejas, S. A., and M. Cai (2016), Isolating the Temperature Feedback Loop and Its Effects on Surface Temperature, *J. Atmospheric Sci.*, *73*(8), 3287–3303, doi:10.1175/JAS-D-15-0287.1.

Sejas, S. A., M. Cai, A. Hu, G. A. Meehl, W. Washington, and P. C. Taylor (2014), Individual Feedback Contributions to the Seasonality of Surface Warming, *J. Clim.*, *27*(14), 5653–5669, doi:10.1175/JCLI-D-13-00658.1.

Soden, B. J., A. J. Broccoli, and R. S. Hemler (2004), On the Use of Cloud Forcing to Estimate Cloud Feedback, *J. Clim.*, *17*(19), 3661–3665, doi:10.1175/1520-0442(2004)017<3661:OTUOCF>2.0.CO;2.

Soden, B. J., I. M. Held, R. Colman, K. M. Shell, J. T. Kiehl, and C. A. Shields (2008), Quantifying Climate Feedbacks Using Radiative Kernels, *J. Clim.*, *21*(14), 3504–3520, doi:10.1175/2007JCLI2110.1.

Song, X., G. J. Zhang, and M. Cai (2014), Characterizing the Climate Feedback Pattern in the NCAR CCSM3-SOM Using Hourly Data, *J. Clim.*, *27*(8), 2912–2930, doi:10.1175/JCLI-D-13-00567.1.

Stephens, G. L. (2005), Cloud feedbacks in the climate system: A critical review., *J. Clim.*, *18*(2), 237–273.

Taylor, P. C., R. G. Ellingson, and M. Cai (2011a), Geographical Distribution of Climate Feedbacks in the NCAR CCSM3.0, *J. Clim.*, *24*(11), 2737–2753, doi:10.1175/2010JCLI3788.1.

Taylor, P. C., R. G. Ellingson, and M. Cai (2011b), Seasonal Variations of Climate Feedbacks in the NCAR CCSM3, *J. Clim.*, *24*(13), 3433–3444, doi:10.1175/2011JCLI3862.1.

Taylor, P. C., M. Cai, A. Hu, J. Meehl, W. Washington, and G. J. Zhang (2013), A Decomposition of Feedback Contributions to Polar Warming Amplification, *J. Clim.*, *26*(18), 7023–7043, doi:10.1175/JCLI-D-12-00696.1.

Washington, W. M., and G. A. Meehl (1984), Seasonal cycle experiment on the climate sensitivity due to a doubling of CO2 with an atmospheric general circulation model coupled to a simple mixed-layer ocean model, *J. Geophys. Res. Atmospheres*, *89*(D6), 9475–9503, doi:10.1029/JD089iD06p09475.

Wetherald, R. T., and S. Manabe (1988), Cloud Feedback Processes in a General Circulation Model, *J. Atmospheric Sci.*, *45*(8), 1397–1416.

Winton, M. (2006), Amplified Arctic climate change: What does surface albedo feedback have to do with it?, *Geophys. Res. Lett.*, *33*(3), doi:10.1029/2005GL025244.

Figure Captions

Figure 1. The net change in solar (SW) radiation ($W*m^{-2}$) at the TOA due exclusively to changes in (a) CO_2, (b) surface albedo, (c) clouds, and (d) water vapor.

Figure 2. The net change in solar (SW) radiation ($W*m^{-2}$) at the surface due exclusively to changes in (a) CO_2, (b) surface albedo, (c) clouds, and (d) water vapor.

Figure 3. The net change in LW radiation ($W*m^{-2}$) at the TOA due exclusively to changes in (a) CO_2, (b) surface albedo, (c) clouds, and (d) water vapor.

Figure 4. The net change in LW radiation ($W*m^{-2}$) at the surface due exclusively to changes in (a) CO_2, (b) surface albedo, (c) clouds, and (d) water vapor.

Figure 5. (a) The net change in LW radiation ($W*m^{-2}$) at the TOA (blue lines) and surface (red lines) due exclusively to the temperature feedback (solid lines) and uniform temperature response (dashed lines). (b) The net change in LW radiation ($W*m^{-2}$) at the TOA (blue line) and surface (red line) due exclusively to the lapse-rate feedback. Notice that the difference between solid and dashed lines of the same color in (a) corresponds to the solid line of the same color in (b).

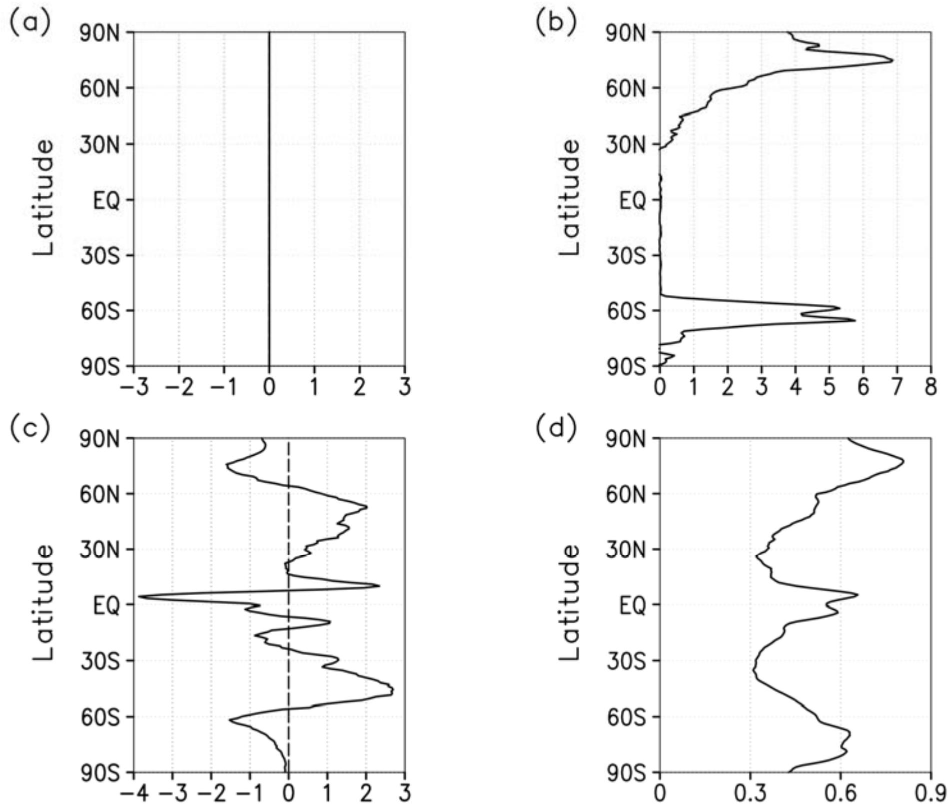


Figure 1. The net change in solar (SW) radiation ($W*m^{-2}$) at the TOA due exclusively to changes in (a) CO_2, (b) surface albedo, (c) clouds, and (d) water vapor.

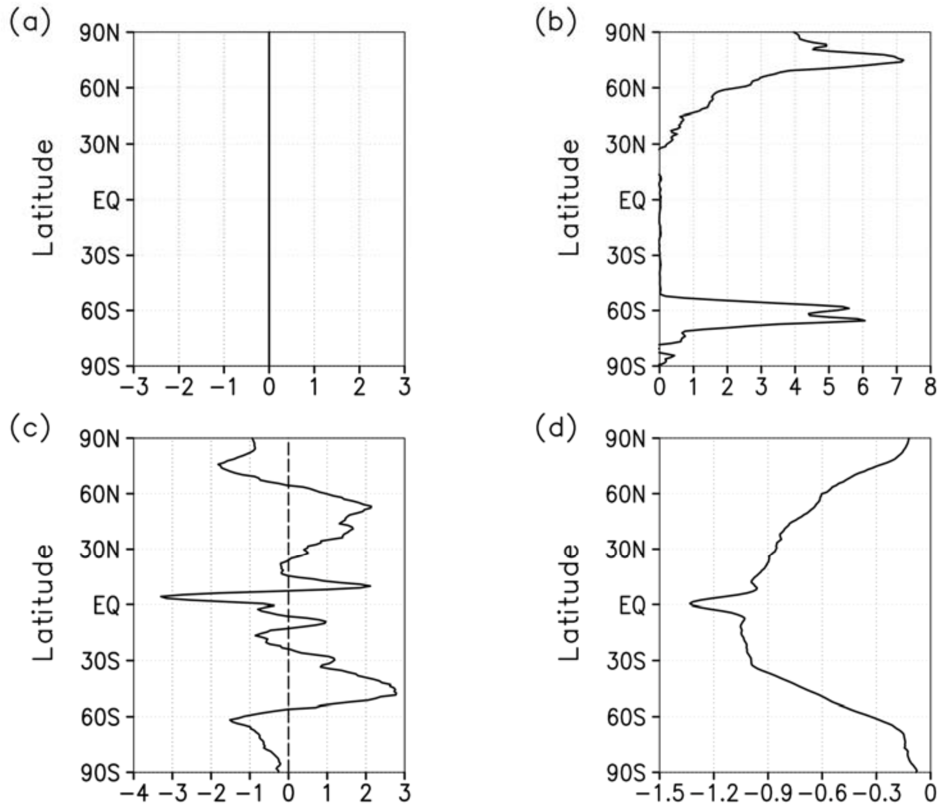


Figure 2. The net change in solar (SW) radiation ($W*m^{-2}$) at the surface due exclusively to changes in (a) CO_2, (b) surface albedo, (c) clouds, and (d) water vapor.

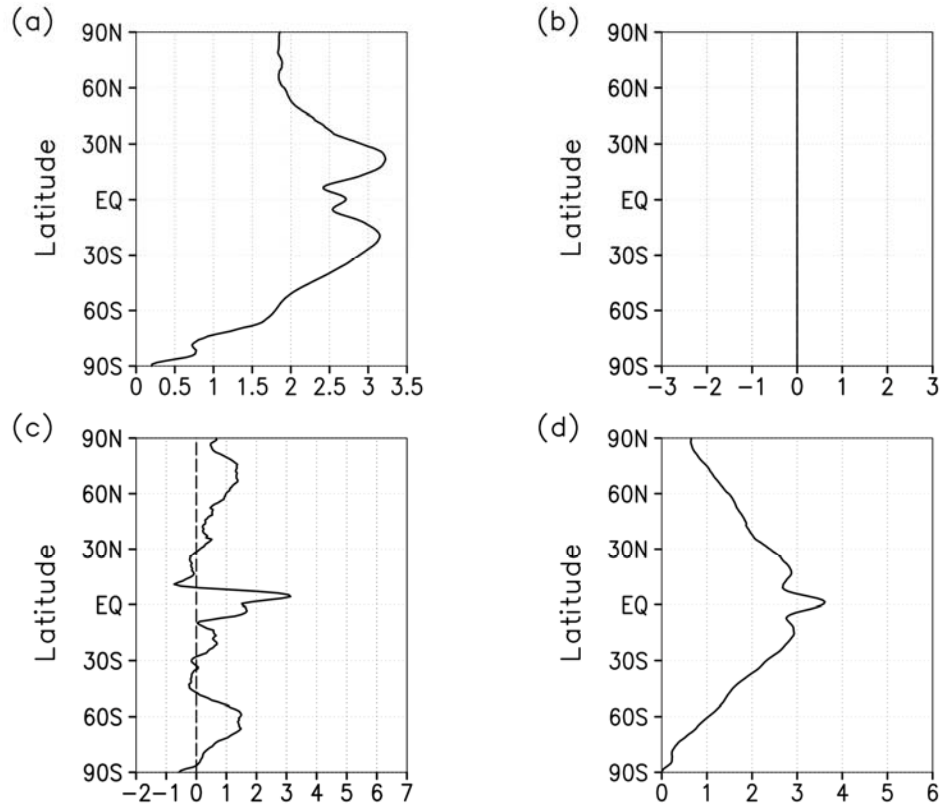
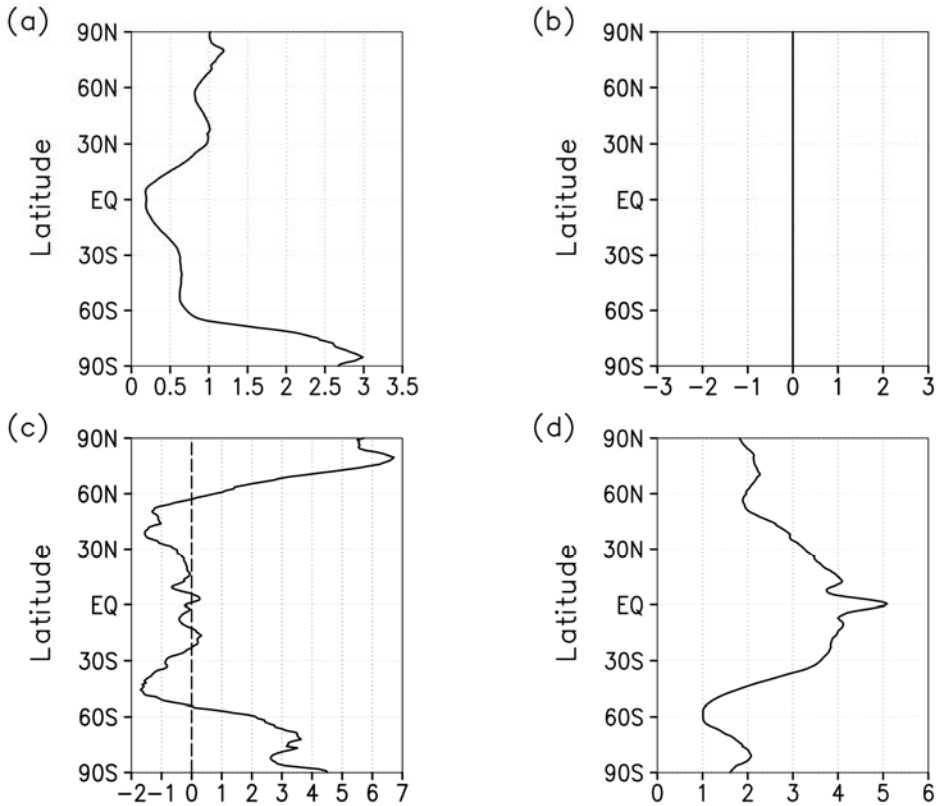


Figure 3. The net change in LW radiation ($W*m^{-2}$) at the TOA due exclusively to changes in (a) CO_2, (b) surface albedo, (c) clouds, and (d) water vapor.



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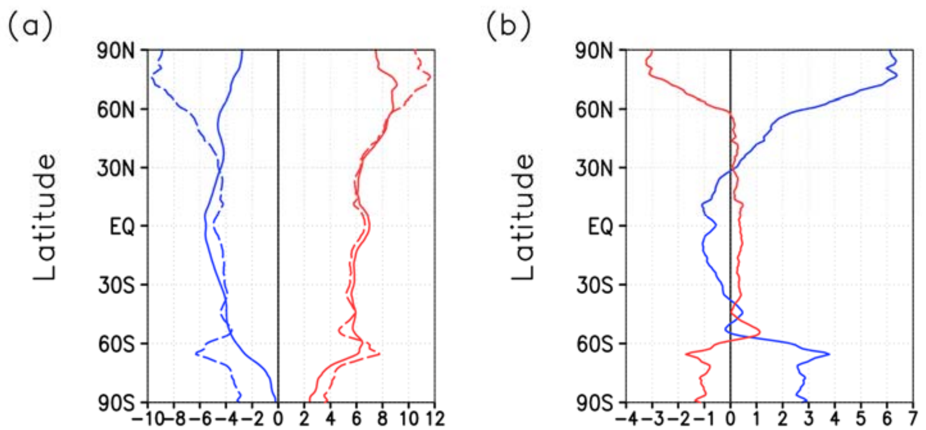

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