The nonlinear and nonlocal nature of climate feedbacks

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ABSTRACT

The climate feedback framework partitions the radiative response to climate forcing into contributions from individual atmospheric processes. The goal of this study is to understand the closure of the energy budget in as much detail and precision as possible, within as clean an experimental set-up as possible. For an aquaplanet simulation under perpetual equinox conditions, we account for rapid tropospheric adjustments to CO$_2$ and diagnose radiative kernels for this precise model set-up. We characterize the meridional structure of feedbacks, energy transport, and nonlinearities in controlling the local climate response. Our results display a combination of strongly positive subtropical feedbacks and polar amplification. These two factors imply a critical role for transport and nonlinear effects, with the latter acting to substantially reduce global climate sensitivity. At the hemispheric scale, a rich picture emerges: heat flux divergence away from positive feedbacks in the tropics; nonlinearities that cool the tropics and warm the high-latitudes; and strong ice-line feedbacks that drive further amplification of polar warming. These results have implications for regional climate predictability, by providing an indication of 1) how spatial patterns in feedbacks combine to affect both the local and nonlocal climate response, and 2) how constraining uncertainty in those feedbacks may constrain the climate response.

1. Introduction

The power of the climate feedback framework lies in its ability to reveal the energy pathways by which the system adjusts to an imposed forcing, such as an increase in atmospheric CO$_2$. These internal adjustments may include changes in physical processes that control the distributions of clouds, water vapor, sea ice, and the vertical structure of temperature, which in turn act to amplify or dampen the surface temperature response to the forcing; these are the climate feedbacks. Further, the system may also adjust by redistributing energy between different latitudes, either by atmospheric or oceanic transport. Understanding the relative importance and effectiveness of these different pathways is crucial for predicting the climate response to a perturbation.

Climate feedbacks are closely related to the change in top-of-atmosphere (TOA) net radiative flux between two equilibrium climate states, $\Delta R$, which can be written as a Taylor series expansion in global-mean surface temperature change, $\Delta T_s$:

$$\Delta R = A + B \Delta T_s + O \Delta T_s^2.$$  \hspace{1cm} (1)

The terms can represent global averages, or be functions of latitude or grid cell. The sign convention is such that a positive radiative flux warms the system. The first term on the right-hand side, $A$, includes the external forcing itself, along with all changes in the radiation balance that are independent of surface temperature change. We refer to $A$ as the climate forcing. The second term, $B\Delta T_s$, reflects radiative flux changes that are linearly dependent on the system response $\Delta T_s$; these are the classical feedback processes. Here, the sign of the feedback term is negative when the system is stable (i.e., a net negative feedback). The third component $O\Delta T_s^2$ represents higher-order terms, which may reflect nonlinearities within individual processes or nonlinear interactions among different processes. In the global mean at equilibrium $\Delta T = 0$, and the temperature-dependent radiative fluxes must balance the forcing. However at any given latitude and longitude, $\Delta R$ is not required to be zero, and can be balanced by changes in atmospheric heat transport and ocean heat uptake. Note that Equation 1 is commonly written in a simplified form, with the nonlinear term $O\Delta T_s^2$ assumed minor and neglected (e.g. Senior and Mitchell 2000; Gregory et al. 2004; Soden and Held 2006).

The goal of this study is to understand the closure of the TOA energy balance in as much detail and precision as possible. Doing so allows us to characterize the relative
importance of local feedbacks, energy transport, and the nonlinear term in controlling the local climate response. We carefully diagnose the climate forcing, taking into account the semi-direct (i.e., temperature independent) response of the atmosphere to CO₂ changes, and we derive the linear part of the response (i.e., the feedbacks) using radiative kernels explicitly calculated for our precise model set-up. In addition, we run our experiment in an idealized aquaplanet model with perpetual equinox conditions and a mixed-layer ocean, which minimizes ambiguities in the results.

In our model, we track the semi-direct (i.e., temperature independent) response of the atmosphere to CO₂ changes, and then integrate the nonlinear term in controlling the local climate response. How important are nonlinearities for the local energy balance, and do they provide insights into understanding the nonlinearity, the forcing, and the influence of meridional energy transport on the TOA energy balance.

2. Analysis

a. Aquaplanet model

We employ the Geophysical Fluid Dynamics Laboratory Atmospheric Model 2 (GFDL AM2) in its aquaplanet configuration, and run to equilibrium. We specify perpetual equinox and daily-mean solar zenith angle. The ocean is represented as a 20-m mixed layer. Sea ice is treated as infinitesimally thin; the ocean albedo is increased to 0.5 where surface temperature drops below 263 K, but no ice thermodynamics are present in the experimental set-up. The critical temperature for sea ice formation was chosen to reproduce a realistic ice-line latitude, when compared to the modern climate. A full description of the AM2 is provided by the GFDL Global Atmospheric Model Development Team (2004). This idealized configuration allows us to cleanly isolate the atmospheric response to CO₂ in the absence of coupled ocean physics, land-sea contrasts and land-surface processes, seasonal and diurnal cycles, and aerosol forcing. Our perturbation is achieved by an instantaneous doubling of CO₂, and then integrating the model out to equilibrium.

Figure 1 shows climatological surface temperature and outgoing longwave radiation (OLR) for control and perturbation experiments, as well as the differences. For this model set-up, doubling CO₂ results in a global-mean temperature increase of 4.69 K, a climate sensitivity that sits slightly above the upper end of the IPCC AR4 “likely” range (Solomon et al. 2007). The shape of the temperature response as a function of latitude is characterized by strong polar amplification; warming peaks at 11.5 K in high northern latitudes, more than twice the global-mean. For comparison, Hwang et al. (2011) find that Arctic warming ranges from 2 to 3 times the global mean for CMIP3 simulations. Maxima in OLR occur over the dry subtropics, and the global-mean OLR for the control run is 235 W m⁻², which is about 10% larger than April climatology provided by NOAA-CIRES Climate Diagnostics Center¹. In response to CO₂ doubling, there is a strong equatorial peak in ΔOLR associated with a 17% decrease in cloud

¹available online at http://www.cdc.noaa.gov/
fraction in the tropical upper-troposphere. In nature, as in more complex models, the meridional structure of annual-mean OLR is blurred by seasonal variations in the position of the intertropical convergence zone (ITCZ), and by zonal asymmetries due to land-sea contrast. The choice of perpetual equinox conditions, which produces a permanent equatorial ITCZ, leads to a focusing of many of the climate fields, which will also become apparent when we examine the patterns of water vapor and cloud feedbacks. This is a trade-off: We gain a clear picture of the feedback patterns and their dynamical causes, but must be more cautious about a direct application of the results to nature.

b. Determination of radiative forcing

Previous feedback studies have commonly assumed a spatially uniform radiative forcing based on estimates of the global-mean (e.g. Soden et al. 2008). However the pattern of radiative forcing can be quite dramatic, as we will show. Since our goal in this study is to close the energy balance as nearly as possible, an updated approach is required that accounts for this spatial variability, and is exact to our experimental set-up. Alternative definitions of radiative forcing are discussed in Hansen et al. (2005). We consider two methods: stratosphere-adjusted, in which the stratosphere is allowed to adjust radiatively to the presence of the forcing agent; and fixed-SST forcing, in which the troposphere is allowed to adjust as well. Both are shown in Figure 2. For a feedback analysis, the latter is to be strongly preferred since it accounts for all changes in forcing that are independent of surface temperature change. In other words, it is closest to the definition of $A$ in Equation 1. We describe each forcing approach in more detail below.

The first method, stratosphere-adjusted radiative forcing, is calculated from the GFDL radiative transfer code, following definitions provided in the IPCC Third Assessment Report (Appendix 6.1 of Ramaswamy et al. 2001; Hansen et al. 2005). Under this classical “fixed dynamical heating” framework, the stratosphere is allowed to adjust to the forcing prior to calculating the TOA flux change. In other words, changes in the downward flux from the stratosphere, as a result of stratospheric temperature change, are assumed to be part of the forcing. The resulting quantity is sometimes called the “adjusted” radiative forcing, and is relevant for CO$_2$ perturbation experiments because the adjustment of the stratosphere is argued to be fast compared to both the tropospheric response and the lifetime of the forcing agents. Once the stratosphere has adjusted
to its new radiative-dynamical equilibrium, the change in flux at the tropopause and at the TOA are identical. The solid gray line in Figure 2 shows the stratosphere-adjusted radiative forcing. It has a global mean value of $3.4 \text{ W m}^{-2}$ and, notably, varies by about a factor of two as a function of latitude. The spatial pattern of the forcing is strongly dependent on cloud masking: Adding CO$_2$ beneath a region of extensive cloud cover has little impact on TOA radiative fluxes. Highest values are found in the cloud-free subtropics.

The second method, fixed-SST forcing, focuses on $\Delta \bar{R}_f$ as the climate forcing applied to the system independent of and prior to a surface temperature response. This definition is spurred by recent modeling results that have demonstrated semi-direct, tropospheric adjustments in response to CO$_2$ (in addition to the direct radiative effect of the greenhouse gas itself), which precede substantial surface warming and affect the TOA radiation balance. In particular, several studies (e.g., Andrews et al. 2011) have emphasized the importance of the cloud response operating over timescales less than one month. This rapid cloud adjustment manifests primarily as a shortwave effect of $<1 \text{ W m}^{-2}$, which Colman and McAvaney (2011) suggest is driven by a decrease in relative humidity and cloud fraction due to enhanced heating at mid-to-lower levels in the troposphere. Other hypotheses involve shoaling of the planetary boundary layer due to suppressed surface heat fluxes (Watanabe et al. 2011) or reductions in entrainment (Wyant et al. 2012). Since it does not constitute a response to surface temperature change, any effect of rapid tropospheric adjustment is more properly combined with the forcing term. Failure to take this rapid adjustment into account as a forcing may bias the cloud feedback calculation.

We therefore perform a fixed-SST experiment, which is able to incorporate the rapid tropospheric adjustment to CO$_2$ prior to surface temperature change—in essence, turning off the feedbacks. A general critique of fixed-SST experiments in standard GCM configurations is that warming still occurs over land surfaces and sea ice, undermining the goal of having no surface response. However the aquaplanet integrations do not suffer from this inconsistency. We easily fix surface temperature everywhere, and in effect equate the fixed-SST forcing of Hansen et al. (2005) with the “adjusted troposphere and stratosphere forcing” of Shine et al. (2003). The fixed-SST experiment is integrated for 40 years with zonally-symmetric and symmetric-about-the-equator specified SSTs (taken from the final year of our long control run). It is otherwise identical to our model set-up for the feedback analysis. The forcing is then simply the change in net TOA radiative flux between $1\times$CO$_2$ and $2\times$CO$_2$ scenarios, with the first year discarded.

The solid black line in Figure 2 shows the climate forcing $\Delta \bar{R}_f$, including both external forcing and rapid tropospheric adjustments. It has a global-mean value of $3.8 \text{ W m}^{-2}$, close to that of the uniform forcing (Myhre et al. 1998). The fixed-SST and stratosphere-adjusted forcings share some similarities, particularly in the southern hemisphere, with maxima in the low- to mid-latitudes. However the fixed-SST case is characterized by notable, and perhaps surprising, hemispheric asymmetries. The locations of these asymmetries coincides with peaks in the standard deviation in OLR, indicating large variability in these regions. The lower panel of Figure 2 shows the standard deviation of annual-mean OLR for the 40-year integration, which maximizes at $3.8 \text{ W m}^{-2}$. We find no interannual persistence in OLR, and our global-mean standard error of $0.2 \text{ W m}^{-2}$ from monthly data is comparable to values cited by previous studies (e.g., $0.3 \text{ W m}^{-2}$ in Shine et al. 2003). Some of the hemispheric asymmetry may be due to the perpetual equinox conditions that limit interaction between the hemispheres.

In the analysis that follows we predominantly use this fixed-SST forcing because it is nearest to our definition of a temperature-independent forcing, as presumed by the feedback framework, and because we believe it represents genuine variability in the forcing. The influence of this asymmetry on our results can be seen in later figures; the differences serve as a rough indication of how uncertainty in forcing influences the meridional structure of feedbacks.

c. Kernels and feedbacks

We apply the radiative kernel method of calculating climate feedbacks, following Soden and Held (2006) and Soden et al. (2008). The kernel represents the TOA radiative adjustment due to a differential nudge in the climate fields, and is calculated separately for changes in temperature, water vapor, and surface albedo. A strength of our analysis is that we explicitly calculate radiative kernels for our precise experimental set-up, thus removing one of the most commonly-cited ambiguities associated with this method.

Radiative kernels are not the only approach for calculating feedbacks, and a comparison of various techniques can be found in Yoshimori et al. (2011). Briefly, kernels are a popular choice for intermodel comparisons because the calculation is based on a small and arguably non-model-specific perturbation (Soden and Held 2006), though they break down for sufficiently different mean states, such as under CO$_2$ octupling (Jonko et al. 2012). Non-kernel feedback calculations include partial radiative perturbation (PRP) and regression. The PRP method (Wetherald and Manabe 1988; Colman 2003) suffers from computational expense and field decorrelation effects. Another approach is the regression method of Gregory et al. (2004), but it is complicated by cloud-masking, ambiguities associated with transient adjustments that result in a poorly-constrained forcing estimate, and the inability to separately evaluate temperature, water vapor, and surface changes. Finally,
recent studies have also proposed to reformulate the kernel framework around relative humidity, rather than specific humidity, thus removing the correlation between water vapor and lapse rate changes (Held and Shell 2012; Ingram 2012). However this rearrangement of energy flux changes into different individual feedbacks does not affect the total linear feedback nor the characterization of the nonlinear term, which is the focus of the present study.

Kernels show particular promise where nonlinear effects are of interest. All feedback methods seek to characterize the linear decomposition of TOA radiative flux changes into the relative contributions from different physical processes. The PRP method is arguably the most exact decomposition of the differences between two climate states because the total (i.e., discrete) changes are used in the radiative calculations. However given our goal to estimate the linearity of climate feedbacks, the kernel method, in its use of small differential changes, is actually closer to the “tangent linear” approximation that is the formal basis for the Taylor series expansion in Equation 1.

Hence, following Soden and Held (2006) and Soden et al. (2008), we compute all feedbacks (with the exception of clouds) as products of two factors. The first is the change in TOA flux due to a small perturbation in variable $x$, and the second is the change in $x$ between the two equilibrium climate states, divided by the global-mean surface temperature response.

$$\lambda_x = \frac{\partial R}{\partial x} \cdot \frac{dx}{dT_s}$$

where $\partial R/\partial x = K_x$, the radiative kernel for $x$. To create the kernels, instantaneous temperatures $T$, including the surface temperature $T_s$, are perturbed by 1 K; surface albedo $\alpha$ is perturbed by 1%; and specific humidity $q$ is perturbed to match the change in saturation specific humidity that would occur from a 1 K warming, assuming fixed relative humidity. We perturb $T$, $\alpha$, and $q$ for each latitude, longitude, time, and pressure level. The kernels are calculated from one year of instantaneous eight-times daily model output. We make computations for clear skies (i.e., clouds instantaneously set to zero) as well as for all-sky conditions simulated by the model. The kernels we derive broadly resemble the kernels calculated from more realistic climate models (i.e., with land, seasonal cycles, etc.), as presented for instance in Soden et al. (2008). However the simplicity of our aquaplanet set-up means the spatial patterns of the kernels are sharper, and can be very clearly related to individual aspects of the atmospheric response. The kernels are presented and described in detail in Appendix B.

Feedbacks are calculated by convolving 10 years of equilibrated monthly anomalies with the 12-month kernels, in the case of temperature, water vapor, and albedo (Eq. 3). In particular, the two parts comprising the temperature feedback are calculated from the surface temperature response applied throughout the troposphere (in the case of the Planck feedback), and the departure at each level from that uniform change (for the lapse rate feedback). We then integrate from the surface to the tropopause, defined as 100 mb at the equator and decreasing linearly to 300 mb at the poles.

Clouds are handled differently, because the radiative effect of vertically-overlapping cloud fields is too nonlinear for the kernel method. Following Soden et al. (2008), the cloud feedback is calculated from the change in cloud radiative forcing ($\Delta CRF$), with adjustments for cloud masking:

$$\lambda_c \Delta T_s = \Delta CRF + (K_{c0}^o - K_{cT})dT + (K_{q0}^o - K_q) dq$$

$$+ (K_{\alpha 0}^o - K_{\alpha})d\alpha + (\Delta \tilde{R}_j^o - \Delta \tilde{R}_j)$$

where $K^0$ terms are the clear-sky kernels, $\Delta \tilde{R}_j^o$ is the clear-sky forcing, and $\Delta CRF$ is defined as the difference between net downward radiative fluxes in all-sky (i.e., the observed meteorological conditions, including clouds if present) and clear-sky (i.e., assuming no cloud) conditions. A discussion of the effect of clouds on clear-sky feedbacks can be found in Soden et al. (2004). Neglecting to account for the cloud-masking adjustments (e.g. Cess et al. 1990; Gregory and Webb 2008) may lead to misdiagnosis of the cloud feedback, as pointed out by Colman (2003).

3. Results

Global-mean feedbacks are shown in Figure 3. We first focus on the solid black bars, which are the feedbacks calculated using the fixed-SST climate forcing. The temperature feedback is strongly negative (i.e., stabilizing the climate): A warmer planet emits more radiation to space (Planck feedback), and the weakened lapse rate, which is a consequence of moist adiabatic stratification, leads to emission from an even warmer atmosphere than if lapse rate were fixed (lapse rate feedback). The individual global-mean values for the Planck and lapse rate feedbacks are -3.03 and -0.69 W m$^{-2}$ K$^{-1}$, respectively. The water vapor feedback is strongly positive (1.62 W m$^{-2}$ K$^{-1}$) because humidity is highly sensitive to warming, and because moistening the atmosphere increases infrared opacity and downwelling radiation. The surface albedo feedback is positive and, as expected, controlled by sea-ice processes. The net cloud feedback is driven by changes in cloud fraction: The longwave cloud feedback is positive due to the insulating effect of widespread increases in high cloud fraction, and the shortwave cloud feedback is positive due to widespread decreases in reflective low cloud fraction (Fig. 5b). These global-mean feedbacks are broadly consistent with coupled-model studies, though our shortwave cloud feedback is on the high end of the range (e.g., Randall et al. 2007). Preliminary results indicate the absence of a Walker circulation in
the sum of the linear feedback terms, which we call "total feedback" for convenience, is small and negative (-0.49 W m$^{-2}$ K$^{-1}$). If the assumption of linearity were correct, then the global climate sensitivity would be $\Delta T_s = (\Delta R - \Delta R_f) / \sum \lambda_x = 7.84$ K, rather than the actual value of 4.69K. This points, then, to a substantial role for the nonlinear term. While it is smaller in magnitude than any individual feedback, comparison of the left- and right-most columns of Figure 3 shows that the nonlinear term is 67% of the total feedback. Thus nonlinearities are of comparable importance to the linear feedbacks in affecting the TOA energy balance, at least in a global-mean sense. Moreover this term tends to have a stabilizing role, in that it reduces global climate sensitivity. The importance of the nonlinearity in the global mean is further motivation to analyze the spatial pattern of the nonlinearity and feedbacks.

How does the magnitude of our nonlinearity compare to previous work? Though reporting conventions vary for the validity of the linear approximation, we can estimate the equivalent nonlinear term based on cited values of feedbacks, forcing, and climate sensitivity. For example, the magnitude of our nonlinear term is -0.33 W m$^{-2}$ K$^{-1}$, and falls between estimates 0.39 W m$^{-2}$ K$^{-1}$ (combining Soden and Held 2006; Soden and Vecchi 2011, for GFDL CM2.1) and 0.13 W m$^{-2}$ K$^{-1}$ (Shell et al. 2008, for CAM3), though our sign is different. Further, if we assume the nonlinear term can be expressed in the form $c\Delta T_s^2$, then it can also be compared to the review by Roe and Armour (2011) of a dozen different estimates. For our present study we find $c \sim -0.07$ W m$^{-2}$ K$^{-2}$. Roe and Armour (2011) report $|c| \leq 0.06$ W m$^{-2}$ K$^{-2}$, with no consensus on sign. Thus the magnitude of our nonlinear term is roughly comparable to previous studies, though on the high end. This may reflect our high climate sensitivity, or be a reflection of the idealized framework. That the nonlinear term is such a large percentage of the total linear feedback is a consequence of the total feedback being small.

Figure 3 shows the spatial pattern of the nonlinear term. For the sake of comparison, Figure 3 also shows global-mean feedbacks for the stratosphere-adjusted radiative forcings. Temperature, water vapor, and albedo feedbacks are unchanged as a function of forcing. The "total feedback" is the sum of the linear feedback terms. We interpret the residual as the nonlinear term. The terms in Equation 2 are normalized by the global-mean surface temperature change, such that units are given in W m$^{-2}$ K$^{-1}$.
The water vapor feedback is positive at all latitudes. However, moistening is most effective—that is, the water vapor feedback is strongest—where humidity is most sensitive to warming (c.f. Fig. B1b in Appendix B) and where high clouds are minimal. These conditions occur in the subtropics. In contrast, the water vapor feedback is weaker along the equator, due to high cloud masking of the tropical moistening at the ITCZ. A key point here is that the water vapor feedback is not independent of the cloud fields, and this interaction between feedbacks hints at the presence of nonlinearity. In other words, water vapor changes under clouds have little effect on the TOA fluxes. The water vapor feedback pattern is particularly sharp due to our perpetual equinox conditions (i.e., lack of seasonality) and aquaplanet configuration. We anticipate that the annual average over seasons would be smoother than the annual average over twelve months of a stationary ITCZ.

The net cloud feedback is positive everywhere except at high latitudes. The breakdown into shortwave and longwave components is shown in Figure 5, along with the change in cloud fraction, which explains much of the meridional structure. As a reminder, warming associated with a positive cloud feedback can occur by decreases in bright clouds (i.e., the SW effect) or increases in high, insulating clouds (i.e., the LW effect). The first thing to note from Figure 5a is that the shortwave component dominates the sign of the net response observed in Figure 4. Hence the net peak in the tropics is a consequence of a decrease in cloud fraction at all levels, but especially in the upper troposphere (with some compensation between a positive shortwave and negative longwave cloud feedback); these cloud fraction changes are consistent with a weakening of the Hadley Cell. The negative net cloud feedback in the high latitudes is due to an increase in low, bright clouds, and a poleward shift of the storm track. The positive net cloud feedback at intermediate, extratropical latitudes is due to widespread decreases in low cloud fraction and increases in high cloud fraction. A more detailed analysis of the shortwave and longwave components of the cloud feedback can be found in Zelinka et al. (2012) as part of the Cloud Feedback Model Intercomparison Project.

The surface albedo feedback has the largest magnitude of any feedback, though it’s confined to the vicinity of the ice line (Fig. 4). Consistent with expectations, reduction of sea-ice cover and the corresponding decrease in surface albedo in a warmer world lead to an increase in absorbed solar radiation, and further warming. Note that the compensation between positive albedo and negative shortwave cloud feedback is observed in Figures 4 and 5. This is a robust result across intermodel comparisons (Zelinka and Hartmann 2012; Crook et al. 2011), though the extent to which clouds are modified by increases in water vapor and evaporation over newly-open water is not easily constrained in a linear feedback framework (Bony et al. 2006; Stephens 2005). Previous studies have also pointed to an increase in high-latitude cloud optical depth due to increases in cloud water content, as well as phase changes (Senior and Mitchell 1993; Tsushima et al. 2006; Zelinka et al. 2012).

The meridional structure of the total feedback is the
sum of the individual feedbacks, and is shown in Figure 4. Overall, the feedback is negative and stabilizing at high-latitudes (with the exception of the ice-line, where the albedo feedback is strong enough to result in a total feedback approaching zero). This locally negative total feedback might lead one to expect a weak surface temperature response, yet Figure 1 shows strong polar amplification. Further, the total feedback is generally positive in the subtropics, which would imply a locally unstable climate--and an infinite response. Clearly then, either substantial redistribution of energy by meridional transport must occur, or else nonlinear interactions must arise. This finding is reminiscent of the work of Pierrehumbert (1995), in which circulation acts to shunt energy from unstable to stable latitudes, which are likened to “radiator fins.” The general tendency of the total feedback to become more negative towards higher latitudes can also be seen in previous studies: Although a large spread exists among models, Zelinka and Hartmann (2012) find that the zonal-mean total feedback parameter averaged over 12 CMIP3 models exhibits a tropical peak. It is not clear if their tropical peak (rather than our subtropical peak) is an artifact of the ensemble average, or if the absence of seasonality in our framework accounts for the difference in location of the unstable regime. In any case, the combination of strong polar amplification and positive subtropical feedbacks implies critical roles for meridional transport and/or nonlinearities, to which we now turn.

The trade-off between the meridional transport and the local demands of linear feedbacks is reflected in the three-way energy balance of Equation 2. In a perfectly linear world, the changes in transport would exactly balance the combined feedbacks and forcing. However in a nonlinear world, that adjustment is incomplete, and the remainder of the energy balance is accommodated by the nonlinear, or residual, term. Our ability to assess this contribution is a key strength of our approach.

The solid grey line in Figure 6 shows the meridional structure of the combined feedback and forcing term. The positive values equatorward of approximately 40° represent a local warming tendency, which must be balanced by the transport and nonlinear terms in Equation 2. The dashed line in Figure 6 shows that there are indeed increased meridional fluxes out of the subtropics, and that the shape of this term closely mirrors that of the feedback-plus-forcing. In other words, in the subtropics, the system attempts to diverge heat away from the region of strong positive feedback, but isn’t entirely successful. The remainder is taken up by the nonlinear term, which provides a cooling tendency in the low latitudes (equatorward of 50°) and a warming tendency elsewhere. Hence the nonlinear term plays an important, compensating role at many latitudes: It opposes the positive feedback in the tropics, and likewise offsets the negative feedback at high latitudes.

Further, the nonlinear term is minimized (i.e., the assumption of linearity works best) in the midlatitudes; a negative total feedback is balanced by anomalous heat convergence at 45°.

a. Polar amplification

Polar amplification is a striking feature of all climate models predictions, and is also observed in global temperature trends (Solomon et al. 2007). In our simulations we see two scales to the polar amplification: an enhancement of the temperature response polewards of about 30°, and a much larger enhancement polewards of 60°. Polewards of 60°, the average warming is 2.2 times the global-mean response; this degree of amplification is consistent with other studies (Hwang et al. 2011; Holland and Bitz 2003). We can apply the feedback framework toward understanding polar amplification in terms of the spatial patterns of climate feedbacks, forcing, heat transport, and nonlinearities. We are also interested in pursuing the apparently contradictory result that the temperature response is largest in regions where the feedback is most stabilizing (compare Figs. 1 and 4).

Equation 2 can be rewritten with local temperature change \( \Delta T_s \) substituted for global-mean \( \Delta T_s \), and the Planck feedback \( \lambda_P \) separated from the non-Planck feedbacks \( \sum \lambda_{NP} \):

\[
\Delta T_s = \frac{1}{\lambda_P} \left[ \Delta R - \left( \sum \lambda_{NP} \right) \Delta T_s - \Delta \bar{R}_f - \mathcal{R} \right]. \tag{5}
\]

In essence, we normalize the terms in the energy balance by the Planck feedback. This weighting avoids an undefined surface temperature response where the total feedback goes
Figure 7: (a) Zonal-mean, annual-mean partial temperature changes. Components are weighted by the Planck feedback, which has meridional structure. (b) Local temperature change $\Delta T_s$ if global-mean weighting $\lambda_P^{\infty}$ were instead applied in Equation 5 (dashed line). Solid line reproduced from upper panel.

to zero. The feedback term in Equation 5 is more similar in form to the conventional definition where feedback factor $f = -\lambda_{NP}/\lambda_P$ (e.g., Roe 2009); this decomposition is also taken by Crook et al. (2011). Thus the pattern of local temperature response is given as the partial temperature change attributed to each term on the right-hand side of Equation 5. These individual contributions, as a function of latitude, are presented in Figure 7, together with the total surface temperature change, shown in grey. As a reminder, the non-Planck feedbacks include lapse rate, water vapor, surface albedo, and cloud feedbacks.

The forcing produces a small and relatively uniform warming of 0.9-1.6 K (red line, Fig. 7). The nonlinear term is also small (green line, $\pm1.6$ K) and, as expected from Figure 6, cools the tropics and warms the high latitudes. The transport term exhibits a similar overall pattern of tropical cooling and high-latitude warming, consistent with a poleward export of heat from the tropics, though its meridional structure and magnitude are more variable. The non-Planck feedbacks provide a warming tendency at all latitudes, and are the major contributor to the more than 10 K warming near the ice-line. In general, non-Planck feedbacks and transport exhibit strong compensation, while the nonlinear term and forcing contribute small perturbations with less meridional variability. Overall then, the enhancement of the average response poleward of 30°, relative to the response equatorward of 30°, may be attributed predominantly to the change in sign of the transport term (and to a lesser degree, the nonlinear term). The pole-to-equator shape of the polar amplification is largely explained by the combined effects of feedbacks and transport.

The further amplification of surface temperature polewards of 60° may be characterized in two parts: non-Planck feedbacks (particularly surface albedo, longwave cloud, and lapse rate feedbacks, see Figs. 4 and 5) from 60-70°, and meridional heat transport of 4.7 K poleward of 70°. The strong warming tendency of the non-Planck feedbacks at the ice-line is partially offset by the transport term (i.e., a cooling tendency due to heat export). Polewards of the ice-line there is anomalous convergence of at least a portion of this exported heat, which maintains the enhanced warming right to the poles. At the poles, none of the terms act as cooling tendencies. Hence we find a consistent picture at both hemispheric and regional scales, in which local temperature change is controlled by heat transport away from regions of strong positive feedbacks (i.e., the ice-line and the subtropics), towards regions of more negative feedbacks (i.e., the midlatitudes and poles).

The influence of the Planck weighting in Equation 5 is demonstrated in the lower panel of Figure 7. The dashed line shows how the predicted surface warming would change if the global-mean weighting $\lambda_P^{\infty}$ had been used in Equation 5, instead of the full spatial field. The meridional structure of the Planck feedback, which increases in magnitude toward the poles (see Fig. 4), contributes an additional 23% warming in the high latitudes (poleward of 60°) and 15% cooling in the subtropics (5-25°). Thus the Planck feedback comes in at tertiary importance, behind the other feedbacks and transport, in explaining polar amplification, though its approximately 2 K high-latitude warming is distributed amongst the other terms and cannot be easily isolated.

Our results have demonstrated the importance of meridional heat fluxes to the system’s response. We next consider the breakdown of the transport term into changes in latent and dry-static energy flux, following Trenberth and Stepaniak (2003) and Hwang and Frierson (2010). As part of the calculation, the surface flux includes contributions from net downward radiation at the surface, sensible heat flux, and latent heat flux due to evaporation and melting snowfall into the ocean. The change in surface flux is smaller than $\pm0.73$ W/m² at all latitudes and negligible in the global mean. The northward latent energy flux is calculated as the integral, with respect to latitude, of evapor-
Changes in northward energy fluxes are shown in Figure 8. Positive slopes in the figure correspond to regions of anomalous flux divergence, and negative slopes to anomalous convergence. The total flux change (grey line) confirms an increase in divergence away from the sub-tropics, and an anomalous divergence away from the ice-line—though in both climate states the latter remains a region of absolute convergence. Relative to the total flux change, the latent and dry-static energy components are large and mostly compensating. In the warmer climate, there is an increase in latent energy flux poleward of approximately 25-30° (solid black line). This is significantly offset by a decrease in dry-static energy flux (dashed line), presumably due to weaker mid-latitude temperature gradients. However the total flux change is still positive, and thus it is the increase in latent energy flux that drives polar amplification polewards of 30°. Interestingly, the dry-static energy gradient weakens considerably polewards of the ice-line. Therefore the contribution of heat transport to polar amplification at the highest latitudes (see also Hwang et al. (2011); Langen et al. (2012)) is driven solely by the latent energy flux convergence, with no compensation from dry-static energy.

Flux changes within the tropics are consistent with warmer temperatures. Figure 8 also shows an increase in equatorward latent heat flux and an increase in poleward dry-static energy flux, which have the same sign as the climatological fluxes. The weakening of the Hadley Cell opposes these tendencies but not enough to overcome them. The total flux change in the tropics is near zero, but its sign is uncertain due to small hemispheric asymmetries in the climate response (e.g., Fig. 6).

4. Summary and Discussion

In this study we have sought to understand the spatial structure of climate feedbacks and the relative importance of nonlinearities and meridional energy transport. We have designed a clean experiment, which seeks to remove as many of the common energy-balance approximations as possible. In particular, we employ a simplified aquaplanet model, and explicitly calculate both fixed-SST climate forcing and radiative kernels for this precise set-up. Our high climate sensitivity of 4.69 K is consistent with large subtropical regions of positive water vapor and cloud feedbacks, and this radiative imbalance translates into polar amplification of the surface temperature response via meridional latent heat transport. Nonlinearities reinforce this pattern. Though some of our results are doubtless a consequence of experimental design, we feel the idealized framework provides a unique lens on the radiative interactions that will have implications for more realistic models.

As mentioned in previous sections, strictly speaking our “nonlinear” term is the residual between the energy-flux changes predicted by linear theory and the actual, model-produced flux changes. While we allow that there may be some effects we have neglected to account for, the residual cannot be attributed to the following: model mismatch between the kernel and experiment; use of a discrete (i.e., PRP) rather than differential (i.e., kernel) approximation of feedback; inexact representation of radiative forcing; complications arising in a more realistic model, such as land-surface/atmosphere interactions, land-sea contrast, or aerosols. It is plausible that our residual would be reduced in magnitude had we included a representation of ocean heat transport and a seasonal cycle; future work will address this relationship. However generally speaking, our results caution against the use of methods in which the residual is subsumed into one of the linear feedbacks; inexact representation of radiative forcing; the cloud feedback of Soden and Held 2006).

Nonlinearities reinforce this pattern. Though some of our results are doubtless a consequence of experimental design, we feel the idealized framework provides a unique lens on the radiative interactions that will have implications for more realistic models.
latter possibility, we have already seen evidence (Figs. 4, B1) of the interconnected relationship between clouds and water vapor, and clouds and temperature sensitivity. Other studies that have investigated nonlinear climate feedbacks include Colman et al. (1997) and Langen et al. (2012).

A possible clue for the source of the nonlinearity comes from the meridional structure of the linear feedbacks. We note that the nonlinear term is negative (a cooling tendency) in the tropics and positive (warming) at higher latitudes, enhancing the pattern of polar amplification. From the cast of linear feedbacks, the only one to work in this manner is the lapse rate feedback, while the water vapor feedback operates with the opposite tendency. In principle then, a small enhancement of the lapse rate feedback at the expense of the water vapor feedback might have the same pattern as the nonlinear term. And in fact, changes in relative humidity, assumed to be fixed in calculating the kernels, might accomplish just such a shift.

The sensitivity of TOA radiation fluxes to upper tropospheric humidity is well known (Cess 1975; Spencer and Braswell 1997). Figure 9 shows changes in relative humidity that are consistent with a weakening and expansion of the Hadley Cell (e.g., Held and Soden 2006)—in particular a decrease in relative humidity of the tropical upper troposphere and stratosphere, and an increase at high latitudes aloft. As noted in Bony et al. (2006), “a change in relative humidity alters the radiative compensation between the water vapor and lapse rate variations, so that an increase (decrease) in relative humidity will enhance (lessen) the water vapor feedback relative to the lapse rate feedback.” The pattern of upper-level relative humidity changes shown in Figure 9 would result in the same meridional structure as the nonlinear term. Hence nonlinear feedbacks may reflect dynamical constraints within the system: Changes in atmospheric circulation modulate the degree of compensation between terms in the energy balance, in a way not accounted for by the linear Taylor-series approximation. Thus dynamically-controlled changes in relative humidity are a likely contributor to the nonlinearity. Although this suggestion is not a definitive explanation of the nonlinear term, it offers one plausible physical mechanism.

To evaluate the nonlinear term properly requires further work. Possible avenues of progress include the following: (1) a comparison of the kernels from control and 2×CO₂ climatologies, or extension of the method to calculate second-order terms; (2) the inclusion of greater realism such as a seasonal cycle or ocean heat transport, which would directly affect the energy balance via the transport term and indirectly through the coupling between feedbacks and surface response; and (3) use of a wider range of forcings to address higher-order feedbacks (e.g., Colman and McAvaney 2009). Furthermore, the equilibrated climate change must satisfy both radiative and dynamical constraints. We’ve seen indications of how dynamically-induced changes in relative humidity can introduce changes in TOA fluxes, which would not occur were the feedbacks linearly independent. Sharply honed numerical experiments that address the conditions under which either dynamics or radiation dominates the response would be useful.

In addition to characterizing the nonlinear term, we have sought to understand how local processes affect nonlocal responses. As an example, the feedback pattern is characterized by strongly positive subtropical feedbacks, and the temperature response pattern characterized by polar amplification. Clearly, meridional energy transport matters for resolving this apparent contradiction. Regions of strong positive feedback, such as the subtropics (and ice-line for non-Planck feedbacks), force anomalous divergence of heat flux. Thus we see transport play a role on several spatial scales: from the subtropics to the high-latitudes, and from the high-latitudes, poleward. The midlatitudes also display interesting complexity, with both regions of positive feedbacks contributing to a maximum increase in convergence at 45°, near the latitude where the nonlinearity is minimized.

Our breakdown of the meridional structure of temperature response into individual components (Equation 5) also illustrates some issues for the predictability of regional climate change. Local feedbacks alone do not set the pattern of temperature response: Atmosphere (and ocean) dynamics act to redistribute energy in the system, and so one must constrain the feedbacks everywhere in order to constrain the response anywhere. Figure 7 shows the partial temperature change for feedbacks, transport, forcing, and nonlinearities as a function of latitude in our simulations. It also provides some sense for how the meridional structure of predicted climate change might vary, if improved
understanding resulted in a different pattern of total feedback.

Conventional climate feedback analysis characterizes only the energy balance and is inherently linear by construction. We have extended that perspective in an idealized framework to include nonlinear terms and to consider dynamical effects. These must operate in the real climate system and are an important component of understanding predictability. The meridional structure of individual feedbacks are governed by the classical climatic zones (i.e., the ITCZ, the subtropics, the midlatitudes, the poles), and thus are a consequence of mean-state dynamics. However dynamical changes in the circulation pattern also contribute to substantial nonlinearities, which decrease climate sensitivity from very high to merely high. We have speculated that the Hadley Cell is capable of adjusting the compensation between individual feedbacks. Further, the system tends to allocate energy towards latitudes that can most effectively radiate to space. This means that warming is minimized in the subtropics in spite of strong positive feedbacks. A complete picture of climate sensitivity must unify dynamical and radiative frameworks, and it is our hope that the current study offers some insights into what that may entail.

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APPENDIX A

Why a clear-sky residual?

The clear-sky (rather than all-sky) residual is a consequence of our cloud feedback calculation. Equation 2 can be rearranged to give

\[ \Delta R = \Delta \tilde{R}_f + \left( \sum_n \lambda_n \right) \Delta T_s + \lambda_c \Delta T_s, \]  

(A1)

where the cloud feedback \( \lambda_c \) is split from the other, non-cloud feedbacks (\( n = T, q, \alpha \)). Substituting Equation 4 into Eq. A1 gives

\[ \Delta R = \Delta \tilde{R}_f^0 + \Delta CRF + \left( \sum_n \lambda_n^0 \right) \Delta T_s, \]  

(A2)

where superscripted terms represent clear-sky fluxes. Hence the residual becomes

\[ \mathcal{R} = (\Delta R - \Delta CRF) = \left[ \Delta \tilde{R}_f^0 + \left( \sum_n \lambda_n^0 \right) \Delta T_s \right], \]  

(A3)

or the difference between actual, model-produced clear-sky fluxes (\( \Delta R - \Delta CRF \)) and kernel-approximated clear-sky fluxes (the remaining terms).

APPENDIX B

Radiative kernels

To facilitate comparison with previous studies (Soden and Held 2006; Soden et al. 2008; Shell et al. 2008), we present height-latitude cross sections of our perpetual equinox, aquaplanet kernels. The kernels in Figure B1 represent the contribution of each level and latitude to the change in longwave TOA fluxes. The temperature kernel (Fig. B1a) is strongly negative (i.e., stabilizing the climate) because an increase in temperature increases OLR, following the Stefan-Boltzman Law. Under clear skies (not shown) the sensitivity peaks in the tropics where temperatures are highest. However all-sky TOA fluxes are sensitive to cloud-top temperature, with the largest contributions from regions of high convective clouds and subtropical and midlatitude boundary layer clouds. The surface component of the temperature kernel (Fig. B1d) exhibits cloud masking, with decreased sensitivity aligned beneath regions of high cloudiness. Cloud-masking effects are also apparent in the surface albedo kernel (Figure B1c), though this kernel obviously only matters near the climatological ice-line.

The water vapor kernel (Fig. B1b) shows the TOA radiative flux response to atmospheric moistening. In calculating the kernel, specific humidity \( q \) was perturbed to match the change in saturation specific humidity that would occur from a 1 K warming, assuming fixed relative humidity (Soden and Held 2006). Positive values indicate that an increase in atmospheric water vapor leads to an increase in infrared opacity and downwelling radiation (decreasing OLR), consistent with the role of water vapor as a greenhouse gas. High sensitivity in the tropics is also influenced by self-broadening of water vapor absorption spectra (Shine and Sinha 1991). At high latitudes and low levels, the water vapor kernel is negative (an anti-greenhouse effect); the effect of humidifying the atmosphere is to raise the emission level (Cess 1975; Held and Soden 2000), leading to an increase in OLR in regions of temperature inversions. The water vapor kernel peaks strongly in the climatologically dry upper troposphere because of the high sensitivity of saturation vapor pressure at very cold temperatures and low pressures (via the Clausius-Clapeyron relationship). For instance, for fixed relative humidity at 200 K, specific humidity changes by 15%/K (Held and Soden 2000). Hence the pattern of this kernel is tied to the assumption of fixed relative humidity. If relative humidity were instead allowed to decrease, then warming would not require moistening, and it would be possible to imagine a
weakened water vapor feedback in the subtropics—though other feedbacks would adjust accordingly to compensate this effect.

Figure B1: Zonal-mean, annual-mean radiative kernels for the GFDL aquaplanet model.

REFERENCES


