Radiative feedbacks on global precipitation

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Received 3 December 2009
Accepted for publication 26 February 2010
Published 12 May 2010
Online at stacks.iop.org/ERL/5/025211

Abstract
The radiative kernel technique is employed to quantify twenty-first century changes to the tropospheric energy budget in the Intergovernmental Panel on Climate Change (IPCC) Fourth Assessment Report (AR4) models in order to better understand changes in global-mean precipitation. The strongest feedbacks on the tropospheric radiative cooling are found to be associated with increases in temperature and water vapor, with the water vapor feedback offsetting a significant portion (∼39%) of the increase in radiative cooling due to higher temperatures. Cloud and surface sensible heat flux feedbacks, though not as large in magnitude as the temperature and water vapor feedbacks, are important contributors to the intermodel difference in the global precipitation response to warming, or hydrological sensitivity. The direct effects of radiative forcing agents on the tropospheric energy budget are also important. Rising CO₂ levels reduce tropospheric radiative cooling and hence limit the increase in global rainfall. Additionally, in some of the models, further reductions in radiative cooling occur due to increases in absorbing aerosol, suggesting that differences in aerosol forcing can explain part of the difference in hydrological sensitivity between models.

Keywords: hydrological cycle, climate models, climate change, radiative feedbacks, radiative kernels

1. Introduction
Understanding the response of the planet’s hydrological cycle to anthropogenic forcing presently represents one of the greatest challenges in climate science. It is generally expected that global-mean precipitation (P) will increase with global warming, although the magnitude of this increase is uncertain. Analyses of observations for the period 1987–2006 suggest an increase in global-mean P of about 6% per degree of surface warming (Wentz et al 2007, Adler et al 2008), which is close to the observed (Wentz and Schabel 2000, Trenberth et al 2005) and modeled (Held and Soden 2006) rate of increase of atmospheric water vapor with temperature. Estimates of P changes for this period, however, are associated with large uncertainty (Lambert et al 2008, Liepert and Previdi 2009). In addition, as shown by Previdi and Liepert (2008) and Liepert and Previdi (2009), the global P response to warming can vary significantly on 20 year timescales, both as a result of internal climate variability and changes in radiative forcing. The implication of this is that recently observed P changes may be a poor indicator of longer-term forced changes that will occur during the remainder of the current century.

The latest generation of coupled atmosphere–ocean general circulation models (GCMs) that participated in the Intergovernmental Panel on Climate Change (IPCC) Fourth Assessment Report (AR4) predict a range of global-mean P responses to warming over the 21st century. This is shown in table 1, which lists the global, annual-mean changes in P and surface air temperature (T{s}) between 2001–2010 and 2101–2110, as well as the ratio of P and T{s} changes, or hydrological sensitivity (HS), for 16 of the AR4 models under the A1B emissions scenario. (This scenario corresponds approximately to a doubling in equivalent CO₂ concentration between 2000 and 2100, after which time the radiative forcings are kept constant.) The multimodel mean HS is 1.67% K⁻¹, but varies between 0.71% K⁻¹ and 2.37% K⁻¹. Very similar values
have been found in many previous studies, both for the AR4 models (e.g., Held and Soden 2006) and for older models (e.g., Mitchell et al. 1987, Allen and Ingram 2002). For the AR4 models considered here, HS varies by about a factor of three, which is larger than the range of $T_s$ responses from these models (which is about a factor of two). Yet considerable attention has been devoted to dissecting the reasons for different model temperature responses, comparatively little focus has been placed on understanding intermodel differences in HS.

Any discussion of global-mean $P$ changes should begin with the energetic constraints on these changes. Specifically, with global warming the atmosphere is expected to lose more radiative energy, primarily as a result of increased longwave (LW) emission due to higher temperatures, but also influenced by other factors. To maintain energy balance, there must then be a compensating increase in atmospheric latent heating associated with greater $P$ (e.g., Mitchell et al. 1987, Stephens et al. 1994, Allen and Ingram 2002, Liepert and Previdi 2009). (Changes in sensible heating of the atmosphere are smaller in magnitude, and, as will be shown, tend to act in the same sense as the radiative heating changes, thus requiring a somewhat larger $P$ increase to re-establish energy balance; note also that the heat capacity of the atmosphere is negligible on climate change timescales.) Increases in global-mean $P$ with warming are therefore constrained by the radiative energy budget of the atmosphere, and identifying the different factors that perturb this energy budget is thus central to understanding changes in $P$.

A portion of the atmospheric radiative heating change that occurs in model simulations and in the real world is due to the direct effect of radiative forcing agents (i.e., the effect that occurs independently of any changes in climate). For example, doubling the CO$_2$ concentration increases the LW heating of the troposphere (or, equivalently, decreases LW cooling) by about 2–3 W m$^{-2}$ (Allen and Ingram 2002). It will be shown in the current study that additional forcing agents, most notably absorbing aerosol such as black carbon, are responsible for further reductions in tropospheric radiative cooling in AR4 model simulations for the 21st century, and that differences in aerosol forcing likely contribute to the intermodel spread in HS.

Remaining changes to the atmospheric radiation budget occur as a result of changes in climate, and these will collectively be referred to as radiative feedbacks, since they either reinforce (for a positive feedback) or counteract (for a negative feedback) the initial heating perturbation induced by a radiative forcing agent. The present study will evaluate feedbacks resulting from changes in temperature ($T$), water vapor ($q$), clouds ($c$), and surface albedo ($a$) in the 16 AR4 models listed in table 1. Feedbacks are calculated using the ‘radiative kernel’ technique (Soden et al. 2008). (The exception to this is the cloud feedback, which is calculated in a different manner as discussed in section 2.) Radiative kernels describe quantitatively the radiation changes resulting from incremental changes in the feedback variables ($T$, $q$ and $a$). Kernels are computed in the current work using an offline version of the radiation code from the Max Planck Institute for Meteorology (MPI-M) ECHAM5 GCM. These radiative kernels (adjoints) are then scaled by the predicted 21st century changes in $T$, $q$ and $a$ from the AR4 models to yield estimates of the radiative feedbacks. Soden and Held (2006), Shell et al. (2008), and Soden et al. (2008) recently employed the kernel method to assess top-of-atmosphere (TOA) radiative feedbacks in the AR4 models. The method provides a computationally efficient and consistent way of evaluating feedbacks in these models, and is utilized for the first time here to assess changes in the atmospheric column radiation with the goal of better understanding modeled changes in $P$. Formulation of the

### Table 1. IPCC AR4 models analyzed in the present study. $\Delta P$ is the change in global, annual-mean precipitation between 2001–2010 and 2011–2100 expressed in units of mm year$^{-1}$. Values in parentheses are the precipitation change in % relative to 2001–2010. $\Delta T_s$ is the change in global, annual-mean surface air temperature in K. Changes are calculated from a single realization of the A1B experiment from each model. Listed for $\Delta P$ and $\Delta T_s$ are the plus and minus 2-sigma uncertainty limits based on analysis of 100 year changes in $P$ and $T_s$ in each model’s pre-industrial control experiment. These limits therefore represent the uncertainty in the forced $P$ and $T_s$ changes over the 21st century. The hydrological sensitivity (in % K$^{-1}$) is the ratio of $\Delta P$ to $\Delta T_s$, with the uncertainty range in parentheses based on the $\pm 2$-sigma limits for these changes.

| Model | $\Delta P$ | $\Delta T_s$ | Hydrological sensitivity |
|-------|------------|-------------|------------------------|
| (1) CCCMA T47 | 47.27 ± 2.60 (4.74 ± 0.27) | 2.30 ± 0.11 | 0.60 (1.85–2.29) |
| (2) CNRM | 44.74 ± 6.99 (3.74 ± 0.59) | 2.51 ± 0.29 | 1.49 (1.13–1.95) |
| (3) GFDL CM2.0 | 23.02 ± 7.44 (2.27 ± 0.74) | 2.67 ± 0.29 | 0.85 (0.52–1.26) |
| (4) GFDL CM2.1 | 16.93 ± 4.41 (1.60 ± 0.42) | 2.26 ± 0.22 | 0.71 (0.48–0.99) |
| (5) GISS AOM | 32.58 ± 2.67 (3.14 ± 0.26) | 1.79 ± 0.14 | 1.75 (1.49–2.06) |
| (6) GISS EH | 36.60 ± 11.17 (3.33 ± 0.98) | 1.84 ± 0.30 | 1.81 (1.10–2.80) |
| (7) GISS ER | 37.72 ± 3.09 (3.48 ± 0.28) | 1.73 ± 0.10 | 2.01 (1.75–2.31) |
| (8) INMCM3 | 42.56 ± 8.55 (4.09 ± 0.84) | 2.35 ± 0.27 | 1.74 (1.24–2.37) |
| (9) IPSL | 57.39 ± 4.37 (5.93 ± 0.46) | 2.74 ± 0.18 | 2.16 (1.87–2.50) |
| (10) MIROC MEDRES | 55.21 ± 2.63 (5.62 ± 0.27) | 3.34 ± 0.16 | 1.68 (1.53–1.85) |
| (11) MPI ECHAM5 | 67.64 ± 6.19 (6.33 ± 0.58) | 3.18 ± 0.22 | 1.99 (1.69–2.33) |
| (12) MRI | 45.95 ± 5.06 (4.90 ± 0.55) | 2.07 ± 0.11 | 2.37 (2.00–2.78) |
| (13) NCAR CCSM3 | 52.25 ± 4.46 (5.06 ± 0.44) | 2.15 ± 0.20 | 2.35 (1.97–2.82) |
| (14) NCAR PCM1 | 30.25 ± 3.48 (2.66 ± 0.31) | 1.56 ± 0.16 | 1.71 (1.37–2.12) |
| (15) UKMO HADCM3 | 27.56 ± 4.54 (2.58 ± 0.43) | 2.68 ± 0.15 | 0.96 (0.75–1.19) |
| (16) UKMO HADGEM1 | 35.99 ± 5.21 (3.23 ± 0.47) | 3.08 ± 0.19 | 1.05 (0.84–1.28) |
radiative kernels for the present study is described in more detail next.

2. Methodology

2.1. Radiative kernels

The atmospheric radiative heating, $R$, can be defined as the difference between the net downward radiation at the TOA and at the surface. $R$ is calculated separately for the LW and shortwave (SW) portions of the spectrum, and these two components are summed to yield the net radiative heating (which is almost always negative). Consider now a change in $R$ between two climate states, which can be expressed as

$$dR = G + \Delta T^s (f_T + f_q + f_c + f_a)$$

(1)

where $f$ are radiative feedbacks resulting from changes in $T$, $q$, $c$ and $a$, and an overbar indicates global averaging. If it is assumed that there are no changes to the dynamical heating of the stratosphere, then $dR$ is equivalent to the change in tropospheric radiative heating. $G$ can then be thought of as the tropospheric forcing, or change in tropospheric radiative heating resulting directly from the presence of external forcing agents. The feedback on $R$ due to $T$ changes is denoted by

$$f_T = \frac{\partial R}{\partial T} \Delta T^s \equiv K_T \frac{dT}{dT^s}$$

(2)

with $f_T$ and $K_T$ defined in an analogous manner. (Computation of $f_T$ is described in section 2.3.) It is important to note that forcings/feedbacks have been defined here such that positive values correspond to a gain of radiative energy for the atmosphere and negative values correspond to a radiative energy loss. Thus, a positive forcing/feedback on $R$ will contribute to a decrease in global-mean $P$, and vice versa for a negative forcing/feedback.

$K_T$ in (2), and $K_q$ and $K_a$ in the corresponding equations for the water vapor and albedo feedbacks, are the radiative kernels, which describe the sensitivity of the atmospheric radiative heating to incremental changes in $T$, $q$ and $a$. To compute the kernels, $R$ is calculated at 3 hourly intervals for one year using the ECHAM5 radiation code and output from a present-day control experiment with an atmosphere-mixed layer ocean version of the model. Values of $R$ for both total- and clear-sky conditions are saved at each timestep. The $T$, $q$ and $a$ fields that are input to the radiative transfer calculations are then perturbed in turn and the radiation code is re-run in order to examine the effect on $R$. For each of the model’s 19 vertical layers, and at the surface, instantaneous temperatures are increased by 1 K at every grid point and time. Instantaneous specific humidities are similarly perturbed separately in each model layer by an amount corresponding to a 1 K warming and fixed relative humidity (RH). Surface albedo values are increased by 1% at all grid points and times. The resulting kernels are then monthly averaged before calculating the feedbacks.

Feedbacks are computed by multiplying the radiative kernels by the AR4 models’ predicted climate response between 2001–2010 and 2101–2110, and then normalizing by the global-mean $T$ change. (A single realization of the A1B experiment is used from each model.) For $f_a$, the independent variable is taken to be $\ln(q)$, since absorption of radiation by water vapor is roughly proportional to the natural logarithm of its concentration.

It should be noted that the method employed here assumes that radiative feedbacks on $R$ are additive (see equation (1)), and that individual feedbacks behave linearly with respect to the magnitude of climate change considered. Shell et al (2008) found these assumptions to generally be valid for clear-sky TOA feedbacks in a CO$_2$ doubling experiment with the NCAR Community Atmospheric Model. Additionally, it is assumed that the radiative kernels will be relatively insensitive to the choices of radiation code and base climate state used to compute them. Soden et al (2008) computed kernels using radiation codes from three different GCMs and found that the global-mean, vertically integrated kernels agreed to within about 5% or less, suggesting that intermodel differences in feedbacks arise principally as a result of differences in climate response and not differences in radiative transfer algorithms. Finally, although the interest here is in the tropospheric radiation budget, it is actually the total column (i.e., TOA minus surface) radiation changes that are quantified by the radiative kernels, thus assuming that the effects of tropospheric $T$, $q$ and $a$ changes on stratospheric temperatures can be ignored.

Figure 1 shows the zonal, annual-mean $T$ kernels for total- and clear-sky conditions. The atmospheric component of the kernels (figures 1(a) and (b)) is negative everywhere, indicating that increases in atmospheric $T$ decrease atmospheric radiative heating (increase radiative cooling) as expected. The largest magnitude values occur where temperatures are highest in the tropical lower troposphere, where increases in $T$ strongly enhance the radiative cooling of the atmosphere to the surface. The surface component of the $T$ kernels is depicted in figure 1(c). Increases in surface $T$ reduce the radiative cooling of the atmosphere by enhancing atmospheric absorption of LW energy. This effect is strongest in the tropics where surface temperatures are highest and the atmosphere is opaque to LW radiation.

The LW and SW components of the $q$ kernels are shown in figures 2 and 3. Following Soden et al (2008), the $q$ kernels have been scaled by the factor

$$\xi = \frac{q}{q_s} \frac{dq_s}{dT}$$

(3)

where $q_s$ is the saturation specific humidity calculated from the monthly mean temperature and pressure at each point. Figures 2 and 3 therefore depict the effect that water vapor increases would have on $R$ assuming that the atmosphere warms uniformly by 1 K while maintaining constant RH. The LW kernels (figure 2) are characterized by a dipole structure between the lower and middle/upper troposphere, with $q$ increases in the former portion of the atmosphere contributing to decreases in $R$ and $q$ increases in the latter leading to $R$ increases. This occurs because a moistening of the lower troposphere enhances the LW cooling to the surface, whereas middle and upper tropospheric moistening diminishes the LW
cooling to space of the warmer atmosphere below. It will be shown in section 3 that $q$ increases in the middle and upper troposphere dominate the LW water vapor feedback on $R$ in the AR4 models during the 21st century, which is also true for the TOA water vapor feedback (Soden et al. 2008). The SW $q$ kernels are plotted in figure 3. Values are positive everywhere since increasing water vapor concentrations will lead to greater atmospheric absorption of SW energy. This effect is strongest in the tropical lower troposphere where water vapor mixing ratios are at a maximum and the annually averaged insolation is large.

Increasing the surface albedo has only a very small impact on $R$, which is expected a priori. $R$ increases by ~0.01–0.02 W m$^{-2}$ per % increase in $a$. This occurs because a higher albedo increases the atmospheric path length for photons, which results in a greater probability that a photon will be absorbed by water vapor or other atmospheric constituents.
2.2. Estimating clear-sky tropospheric forcing

The portion of $dR$ that is independent of climate change is referred to above as the tropospheric forcing and denoted by $G$ in equation (1). While estimates of $G$ for the A1B simulations are not available, it is possible to derive the clear-sky tropospheric forcing, $G^0$, using model-simulated changes in the clear-sky tropospheric radiative heating, $dR^{0}_{TROP}$:

$$G^0 = dR^{0}_{TROP} - \Delta \tilde{T}_s (f^0_T + f^0_q + f^0_a)$$

(4)

where $f^0$ are the clear-sky feedbacks. $dR^{0}_{TROP}$ is defined as the difference between the net downward clear-sky radiation at the tropopause and at the surface. A subset of four of the models listed in table 1 (GFDL CM2.0, GFDL CM2.1, MIROC MEDRES, and MPI ECHAM5) has archived radiative fluxes at the 200 hPa level, and these are used here to compute $dR^{0}_{TROP}$. Feedbacks in (4) are vertically integrated from the surface to the tropopause, defined in this case as 100 hPa at the equator and increasing linearly with latitude to 300 hPa at the poles. It is important to point out that estimates of $G^0$ obtained in this manner will be affected by any errors in the assumed tropopause height and in the calculated clear-sky feedbacks.

For the four models that had the necessary data, the global, annual-mean LW $G^0$ was found to vary between 3.84 and 4.16 W m$^{-2}$. However, while the models generally agree on the LW forcing, this is not the case for the SW component of $G^0$, as illustrated by figure 4. The GFDL CM2.0, GFDL CM2.1, and MIROC models are found to have global-mean
Figure 3. Zonal, annual-mean shortwave water vapor kernel for (a) total-sky and (b) clear-sky conditions.

SW forcings of 1.21, 1.28, and 1.67 W m$^{-2}$, respectively, which are of opposite sign to the inferred SW forcing for ECHAM5 of $-0.73$ W m$^{-2}$. To understand these differences, it is important to realize that the various AR4 models did not include the same set of external forcing agents in their A1B simulations. While all models prescribed the same changes in well-mixed greenhouse gases and sulfate aerosols, it was left to the discretion of individual modeling groups whether or not to include additional forcings. Meehl et al. (2007) list the forcings that were included in each model. The two GFDL models and the MIROC model prescribed time-varying black and organic carbon in their 21st century simulations, with the latter model additionally specifying variations in dust and sea salt. In contrast, concentrations of all of these aerosol species were held fixed in the 21st century in ECHAM5. It thus seems likely that differences in aerosol forcing, in particular forcing associated with absorbing aerosol such as black carbon, can explain the differences in SW $G^0$ amongst the four models examined here. This is supported by the spatial pattern of the forcing in figure 4. In the GFDL and MIROC models, the largest forcing is concentrated mainly in the Northern Hemisphere where aerosol emissions are highest. Maximum values of $G^0$ occur over and downwind of regions such as southeast Asia that are major emitters of black carbon. These results therefore suggest that increases in absorbing aerosol during the 21st century in some of the AR4 models may decrease the global-mean tropospheric radiative cooling by more than 1 W m$^{-2}$ (at least under clear skies), indicating
a negative impact on global-mean $P$. Since these aerosol changes are not included in all of the models, however, they likely contribute to the intermodel spread in HS.

2.3. Calculating cloud feedback

Nonlinearities associated with clouds preclude the use of the kernel method to estimate cloud feedback. An alternative approach that has often been used is to examine the change in cloud radiative forcing (CRF). In the present study, interest is in the atmospheric CRF, which represents the effect of clouds on the atmospheric radiative heating. The change in atmospheric CRF, dCRF, is defined simply as the difference between $dR$ and $dR^0$, the latter being the change in clear-sky atmospheric radiative heating. One of the difficulties with employing this method to estimate cloud feedback is that the CRF depends on the clear-sky radiative fluxes. Thus, changes to these clear-sky fluxes resulting from changes to the noncloud variables (i.e., $T$, $q$, and $a$) may lead one to calculate a CRF change and infer a cloud feedback even in the absence of any actual variations in cloud properties. Shell et al (2008) and Soden et al (2008) utilized a technique of adjusting the CRF change to account for these effects of the noncloud variables. This approach is used again here to estimate the cloud feedback, $f_c$. The change in atmospheric radiative heating, $dR$, can be written as the sum of the changes in the clear-sky radiative fluxes and the change in CRF:

$$dR = G^0 + \Delta T_s (f_T^0 + f_q^0 + f_a^0) + dCRF.$$  \hspace{1cm} (5)

Equating (5) with (1) and dividing by the global-mean $T_s$ change yields

$$f_c = \frac{dCRF}{\Delta T_s} + (f_T^0 - f_T) + (f_q^0 - f_q) + (f_a^0 - f_a) + \frac{(G^0 - G)}{\Delta T_s}. \hspace{1cm} (6)$$

$G^0$ was calculated for 4 of the AR4 models in section 2.2. These and an additional 5 models had the necessary data.
available to compute the change in CRF. For the remaining 5 models, the multimodel mean \( G^0 \) derived in section 2.2 is used in equation (6). Based on a doubled CO\(_2\) calculation with the ECHAM5 radiation code, it is estimated that the presence of clouds increases the global-mean clear-sky tropospheric forcing by \( \sim 9\% \). Assuming that this also holds for the A1B simulations, \( G \) is obtained by increasing \( G^0 \) by a spatially uniform 9%.

### 3. Evaluation of feedbacks in the AR4 models

In this section, feedbacks on the atmospheric radiative heating, \( R \), in the A1B simulations are analyzed. Figure 5(a) shows the global-mean, vertically integrated feedbacks resulting from changes in \( T \), \( q \) and \( c. \) (Feedbacks are vertically integrated from the surface to the tropopause, again defined as 100 hPa at the equator and increasing linearly with latitude to 300 hPa at the poles.) The albedo feedback is small in all models (\( -0.01 \) W m\(^{-2}\) K\(^{-1}\) < \( f_a \) < 0) and is not shown. Numbers along the abcissa correspond to the models listed in table 1. The required data were available to compute the temperature feedback for all 16 models, the water vapor feedback for 15 of the models, and the cloud feedback for 9 of the models. Figure 5(a) indicates that the strongest feedback is the \( T \) feedback, with a multimodel mean value of \( -3.23 \) W m\(^{-2}\) K\(^{-1}\). The \( T \) feedback can be thought of as the sum of two components: a Planck feedback, which represents the change in \( R \) due to a vertically uniform warming of magnitude \( \Delta T \); and a lapse rate feedback, which is the additional \( R \) change associated with vertically non-uniform warming. The multimodel mean Planck and lapse rate feedbacks are \(-2.16\) and \(-1.07\) W m\(^{-2}\) K\(^{-1}\), respectively.

The effects of tropospheric warming on \( R \) are opposed by the effects of tropospheric moistening, with the multimodel mean \( q \) feedback found to be \( 1.27 \) W m\(^{-2}\) K\(^{-1}\). In the hypothetical situation in which the atmosphere is assumed to maintain constant RH as it warms, the strength of the water vapor feedback is reduced by \( \sim 18\% \) on average (green bars in figure 5(a)). (Note that the constant RH \( q \) feedback could only be computed in 14 of the models.) The cloud feedback is smaller in magnitude than the \( T \) and \( q \) feedbacks and is positive in seven of the nine models, with multimodel mean of \( 0.15 \) W m\(^{-2}\) K\(^{-1}\).

The total water vapor feedback in the AR4 models arises as a result of positive contributions from both the LW and SW (0.29 and 0.98 W m\(^{-2}\) K\(^{-1}\), respectively, in the multimodel mean). This is at odds with the results of Mitchell et al (1987), who found that a positive SW \( q \) feedback on \( R \) was countered by a negative LW \( q \) feedback of about the same magnitude. (The Mitchell et al study examined changes in \( R \) in an atmospheric GCM forced with a doubling of CO\(_2\) and a spatially uniform increase in sea surface temperature (SST) of 2 K.) At first glance, the findings in the current work may also appear to disagree with observational studies showing a positive correlation between the column-integrated water vapor and the (clear-sky) atmospheric LW cooling (e.g., Stephens et al 1994, Allan 2009). However, since water vapor is tightly coupled to temperature, this correlation probably largely reflects the effect of \( T \) changes on the LW cooling. Additionally, changes to the column-integrated water vapor are dominated by changes in the lower troposphere, and \( q \) increases in this portion of the atmosphere are associated with increases in the column LW cooling, in contrast to the effect of \( q \) increases in the middle and upper troposphere (figure 2).

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**Figure 5.** Global, annual-mean (a) radiative and (b) non-radiative (surface sensible heat flux) feedbacks on the atmospheric energy budget between 2001–2010 and 2101–2110. Positive values signify a gain of energy for the atmospheric column. Numbers along the abscissa correspond to the AR4 models listed in table 1.
Figure 6. Annual, multimodel mean radiative feedbacks associated with changes in (a) temperature, (b) water vapor, and (c) clouds. The unadjusted change in atmospheric cloud radiative forcing (normalized by the global-mean surface air temperature change) is shown in (d).

Indeed, Inamdar et al (2004) conclude based on satellite observations that increases in the integrated water vapor above 500 hPa are associated with decreases in the total column LW cooling for SSTs higher than about 297 K. It is unclear why the results of the present study differ from those of Mitchell et al (1987). The latter study did, however, find a significant (∼5–10%) decrease in RH in the tropical upper troposphere, which would explain at least part of the difference from the results presented here.

Although the focus in this paper is mainly on changes to the radiative energy budget of the atmosphere, $dR$, it is actually $dR$ plus the change in surface sensible heat flux (SHF) that is balanced by the change in global-mean $P$. Figure 5(b) shows the global-mean SHF change (normalized by the global-mean $T_s$ change) over the 21st century in the AR4 models. In 14 of the 16 models, there is a decrease in the sensible heating of the atmosphere, with a multimodel mean decrease of $-0.40 \text{ W m}^{-2} \text{ K}^{-1}$. SHF changes therefore represent an additional loss of energy from the atmosphere, thus requiring a somewhat larger $P$ increase to re-establish energy balance (for the same $dR$).

Geographical patterns of the vertically integrated temperature, water vapor and cloud feedbacks are presented in figures 6(a)–(c). The largest magnitude temperature and water vapor feedbacks occur in tropical latitudes where the sensitivity of $R$ to $T$ and $q$ perturbations is highest (figures 1–3). The largest positive cloud feedback is found in a band along the equatorial Pacific and the largest negative cloud feedback is found over the Arctic. In the global, multimodel mean, cloud changes act to decrease the radiative cooling of the atmosphere ($f_c = 0.15 \text{ W m}^{-2} \text{ K}^{-1}$), in agreement with the results of Stephens and Ellis (2008) and Lambert and Webb (2008). However, this cloud effect on $R$ is weaker than would be inferred from the (unadjusted) change in atmospheric CRF.
(figure 6(d)), thus highlighting the potential difficulties in using the CRF change as an indicator of cloud feedback (Soden et al. 2004).

To better understand the spread in HS between models, intermodel standard deviations of the feedbacks are computed. (Note that intermodel spread in HS may also arise due to differences in tropospheric forcing, as discussed above.) Since the lapse rate and water vapor feedbacks are tightly coupled, it is logical to consider the sum of these two feedbacks when comparing different models rather than considering each feedback individually (Soden and Held 2006). The standard deviation of the combined lapse rate plus water vapor feedback is found to be 0.22 W m$^{-2}$ K$^{-1}$. This is much larger than the standard deviation of the Planck feedback (0.03 W m$^{-2}$ K$^{-1}$), but is similar to the standard deviations of the cloud and SHF feedbacks (0.20 and 0.23 W m$^{-2}$ K$^{-1}$, respectively). Thus, just as differences in cloud feedback are the major source of intermodel spread in climate (temperature) sensitivity (e.g., Soden and Held 2006), they are likewise a major source of intermodel spread in hydrological (precipitation) sensitivity. Additionally, for HS, differences in the lapse rate plus water vapor feedback and in the surface SHF change between models are also important considerations.

4. Summary and conclusions

This paper has assessed changes to the tropospheric energy budget over the 21st century (i.e., 2101–2110 minus 2001–2010) in the IPCC AR4 models in order to better understand simulated changes in global-mean precipitation. Both forcings and ensuing climate feedbacks on the energy budget have been quantified. The clear-sky tropospheric forcing, $G^0$, was estimated for a subset of four of the models. While these models generally agreed on the LW $G^0$ (range of 3.84–4.16 W m$^{-2}$ in the global, annual-mean), there were large differences in the SW (range of −0.73–1.67 W m$^{-2}$). This is due to the fact that the LW forcing is dominated by increases in well-mixed greenhouse gases (mainly CO$_2$) that are prescribed uniformly across models, whereas the SW forcing is dominated by aerosol changes that can vary significantly from model to model. Thus, in all models rising greenhouse gas levels act to reduce tropospheric LW cooling and hence limit the magnitude of the global-mean $P$ increase. In some of the models, significant increases in the SW heating of the troposphere occur due to increases in absorbing aerosol concentrations, which would tend to further damp the $P$ response. Intermodel differences in HS in the A1B simulations are therefore likely to partly reflect intermodel differences in aerosol forcing.

In addition to being dependent on the tropospheric forcing, model-simulated $P$ changes also depend on the various feedbacks that act to further modulate the tropospheric energy budget. Radiative feedbacks associated with changes in temperature, water vapor, clouds, and surface albedo, as well as the non-radiative feedback due to surface SHF changes, have been evaluated in the present study. The $T$, $q$ and $a$ feedbacks were quantified using the radiative kernel technique, a method employed previously to estimate TOA feedbacks in the AR4 models (Soden and Held 2006, Soden et al. 2008). The $c$ feedback was calculated by adjusting the CRF change to account for changes to the clear-sky radiative fluxes (Shell et al. 2008, Soden et al. 2008). Not surprisingly, the strongest feedback was found to be the $T$ feedback. In the global, annual and multimodel mean, increases in $T$ enhance the radiative cooling of the troposphere by 3.23 W m$^{-2}$ K$^{-1}$. The effects of tropospheric warming on $R$ are opposed by the effects of tropospheric moistening. In particular, increases in water vapor are shown to reduce tropospheric radiative cooling by 1.27 W m$^{-2}$ K$^{-1}$, thus offsetting about 39% of the $T$ effect. Cloud (SHF) feedbacks act to decrease (increase) tropospheric energy loss by 0.15 W m$^{-2}$ K$^{-1}$ (0.40 W m$^{-2}$ K$^{-1}$). While these feedbacks are not as large in magnitude as the temperature and water vapor feedbacks, they explain a significant proportion of the intermodel spread in HS. The $a$ feedback is negligible in all models.

The current work has sought to enhance our understanding of the processes controlling global-mean precipitation changes in state-of-the-art GCMs. There is still more to be done, however. For example, the assumption of linearity of the feedback mechanisms should be further tested. In addition, several of the feedback mechanisms (e.g., associated with clouds) are still poorly understood. Finally, results presented here suggest the need to be able to better quantify radiative forcing in GCM simulations, including forcing at the TOA, surface and throughout the atmospheric column. Future model intercomparisons will benefit immensely if the forcing used to drive the models is routinely and consistently calculated and made available to the scientific community.

Acknowledgments

I thank Daniel Klocke for providing the ECHAM5 radiation code used to compute the radiative kernels. The kernels will be made available to members of the scientific community upon request. I also thank two anonymous reviewers whose comments helped strengthen the manuscript.

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