NOTES AND CORRESPONDENCE

An Assessment of the Primary Sources of Spread of Global Warming Estimates from Coupled Atmosphere–Ocean Models

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(Manuscript received 13 September 2007, in final form 21 February 2008)

ABSTRACT

Climate feedback analysis constitutes a useful framework for comparing the global mean surface temperature responses to an external forcing predicted by general circulation models (GCMs). Nevertheless, the contributions of the different radiative feedbacks to global warming (in equilibrium or transient conditions) and their comparison with the contribution of other processes (e.g., the ocean heat uptake) have not been quantified explicitly. Here these contributions from the classical feedback analysis framework are defined and quantified for an ensemble of 12 third phase of the Coupled Model Intercomparison Project (CMIP3)/Intergovernmental Panel on Climate Change (IPCC) Fourth Assessment Report (AR4) coupled atmosphere–ocean GCMs. In transient simulations, the multimodel mean contributions to global warming associated with the combined water vapor–lapse-rate feedback, cloud feedback, and ocean heat uptake are comparable. However, intermodel differences in cloud feedbacks constitute by far the most primary source of spread of both equilibrium and transient climate responses simulated by GCMs. The spread associated with intermodel differences in cloud feedbacks appears to be roughly 3 times larger than that associated either with the combined water vapor–lapse-rate feedback, the ocean heat uptake, or the radiative forcing.

1. Introduction

The spread of the equilibrium or transient surface temperature response to a CO_2 doubling as predicted by atmosphere–ocean coupled models is still large (Meehl et al. 2007), and an open question is to identify the primary sources of this spread. Global warming estimates depend on radiative forcing, feedback processes that may either amplify or dampen the climate response and, in the transient case, ocean heat uptake. For individual models, it has been suggested that atmospheric processes were the most critical factors for estimating global temperature changes in transient simulations (e.g., Williams et al. 2001; Meehl et al. 2004; Collins et al. 2007). Here our purpose is to investigate whether these results extend to multimodel ensembles, and how much the various feedbacks and the ocean heat uptake contribute to the multimodel mean and spread of global warming estimates. The main radiative feedbacks are associated with changes in water vapor (WV), temperature lapse rate (LR), clouds, and surface albedo. The associated feedback parameters have been diagnosed for some multimodel ensembles (e.g., Colman 2003; Soden and Held 2006; Webb et al. 2006), but they have not been translated into temperature changes. This makes it difficult to compare the temperature change associated with each feedback with that from other processes, such as the ocean heat uptake.

In this paper we show that it is possible to decompose, and thus to compare, the contributions of the different climate feedbacks, and eventually of the ocean heat uptake, to the global temperature response to a specified forcing. After a brief presentation of the feedback analysis framework (section 2), the decomposition methodology is presented (section 3) and, after gathering the required data (feedback parameters, radiative forcing, and ocean heat uptake; section 4), this methodology is applied to an ensemble of models that participated in the World Climate Research Programme's (WCRP's) third phase of the Coupled Model

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DOI: 10.1175/2008JCLI2239.1

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Intercomparison Project (CMIP3) in support of the Intergovernmental Panel on Climate Change (IPCC) Fourth Assessment Report (AR4; section 5). There is very little in this paper that is entirely new. Rather, we propose a new presentation of existing results that allows us to quantify in a more straightforward way the relative contribution of different processes to intermodel differences in global mean temperature changes.

2. The feedback analysis framework

Let us consider a steady-state climate, with a time average value $F_t^o = 0$ of the global mean net flux at the top of the atmosphere (TOA) and a time average value T_s^o of the global mean surface temperature. Let us impose on the climate system a radiative forcing, such as a change in the greenhouse gas concentration or in the TOA incoming solar radiation. In the absence of surface temperature change, this forcing translates into a radiative flux perturbation ΔQ_t at the TOA, called radiative forcing. In response to this disequilibrium, the surface temperature changes. It appears that at any time, the anomalies ΔT_s and ΔF_t of the surface temperature and the TOA flux from their unperturbed initial steady state are approximately related through the following equation:

$$\Delta T_s = \frac{\Delta F_t - \Delta Q_t}{\lambda}, \tag{1}$$

where λ is the "climate feedback parameter," and the fluxes are positive downward. This relationship holds both for transient and equilibrium conditions. If the temperature changes until a new equilibrium is reached, the TOA net flux reaches its steady-state value ($\Delta F_t = 0$) and the equilibrium temperature change is

$$\Delta T_s^{\,e} = \frac{-\Delta Q_t}{\lambda}. \tag{2}$$

The total feedback parameter is commonly split as the sum of five terms,

$$\lambda = \lambda_P + \lambda_w + \lambda_L + \lambda_c + \lambda_\alpha, \tag{3}$$

which are the Planck (P), water vapor (w), lapse-rate (L), cloud (c), and surface albedo (α) feedback parameters. In this approach it is assumed that everything is linear [see, e.g., the appendix of Bony et al. (2006) for more details on this approach and for a discussion of the approximations].

In climate feedback studies, temperature responses are often compared to the basic equilibrium temperature response $\Delta T_{s,P}$ that would be obtained if the temperature change was horizontally and vertically uniform and was only modifying the infrared emission through a change in the Planck function (e.g., Hansen et al. 1984),

$$\Delta T_{s,P} = -\frac{\Delta Q_t}{\lambda_P}. \tag{4}$$

Because the total feedback parameter may be decomposed as $\lambda = \lambda_P + \Sigma_{\boldsymbol{x}\neq P}\lambda_{\boldsymbol{x}}$ [cf. Eq. (3)], at equilibrium ($\Delta F_t = 0$) Eq. (1) reads

$$\Delta T_s = \frac{1}{1 - \sum_{\boldsymbol{x}\neq P} g_{\boldsymbol{x}}}\Delta T_{s,P}, \tag{5}$$

where $g_{\boldsymbol{x}} = -(\lambda_{\boldsymbol{x}}/\lambda_P)$ is called the feedback gain for the variable $\boldsymbol{x}$. If the total feedback gain

$$g = \sum_{\boldsymbol{x}\neq P} g_{\boldsymbol{x}} \tag{6}$$

is positive (negative), the temperature change ΔT_s is larger (smaller) than the temperature change $\Delta T_{s,P}$ associated with the Planck response.

3. Relative contribution of each feedback to the global temperature change

a. Equilibrium temperature change

When only one feedback loop x is active in addition to the Planck response, the equilibrium temperature change resulting from this feedback is simply and uniquely defined from Eq. (5) as the difference $\delta_1 T_{s,x}$ between the temperature change with and without this feedback x,

$$\delta_1 T_{s,x} = \frac{1}{1 - g_x}\Delta T_{s,P} - \Delta T_{s,P}. \tag{7}$$

When several feedbacks are active, various approaches may be used. The first approach is to quantify, as previously, the effect of each feedback as the difference between the temperature change with and without this feedback x [Eq. (7)]. A second possibility is to quantify this effect as the difference $\delta_2 T_{s,x}$ between the temperature change when all the feedbacks are active and when all the feedbacks but x are active,

$$\delta_2 T_{s,x} = \left(\frac{1}{1 - g} - \frac{1}{1 - (g - g_{\boldsymbol{x}})}\right)\Delta T_{s,P}. \tag{8}$$

In this definition, the effect of a feedback loop x on the temperature change depends both on its gain g_x and on the gain g of all feedbacks (e.g., Hansen et al. 1984; Hall and Manabe 1999); thus, it cannot be defined independently of the rest of the system. The temperature change obtained with these two definitions may be very different.

Because there is no unique way to define the effect of individual feedbacks on the temperature change, we reformulate the question as follows: knowing the global temperature change, what is the part of this temperature change that is due to each feedback? In other words, we enforce that the sum of the different temperature changes $\Delta T_{s,x}$ associated with each feedback x plus the temperature change $\Delta T_{s,P}$ associated with the Planck response equals the total temperature change ΔT_s,

$$\Delta T_s = \Delta T_{s,P} + \sum_{x \neq P} \Delta T_{s,x}. \tag{9}$$

From Eq. (5), it follows that

$$\Delta T_{s,x} = \frac{g_x}{1 - g} \Delta T_{s,P} = g_x \Delta T_s \quad \text{for} \quad x \neq P. \tag{10}$$

This expression can also be directly obtained by noting that ΔT_s [Eq. (5)] cannot be directly decomposed into additive contributions associated with each feedback, whereas the difference $\Delta T_s - \Delta T_{s,P}$ can. This new definition leads to partial temperature changes that have some interesting properties. If the feedback parameter λ_x of a feedback x is multiplied by a factor α and the total gain g is unchanged (in this case, other feedback parameters have also to be modified), the temperature change $\Delta T_{s,x}$ associated with this feedback x is multiplied by α. If the feedback parameters of two feedbacks x and y are both multiplied by a factor α, the ratio $(\Delta T_{s,x}/\Delta T_{s,y})$ is not modified. If the feedback parameters of all the feedbacks are multiplied by a same factor α, the ratio $(\Delta T_{s,x}/\Sigma_{y \neq P} \Delta T_{s,y})$, that is, the relative fraction of the temperature change resulting from each feedback x is not modified. Therefore, this definition of the partial temperature change allows us to compare and to add the contribution of the various feedbacks to the temperature response.

It is important to note that the temperature change associated with the Planck response [Eq. (4)] and the one associated with each feedback x [Eq. (10)] are of different natures because of the very specific role of the Planck response (the "basic" response on which the others are feedbacks). Equation (10) may also be written as follows:

$$\lambda_P \Delta T_{s,x} = -\lambda_x \Delta T_s \quad \text{for} \quad x \neq P. \tag{11}$$

In this equation, the left-hand side is the change of the TOA flux resulting from the partial temperature change $\Delta T_{s,x}$ if the temperature change was uniform and only affected the thermal emission. The right-hand side is the change of the TOA flux $\Delta F_{t,x}$ resulting from the total temperature change ΔT_s through the feedback x. The temperature change $\Delta T_{s,x}$ associated with a feedback x is the temperature change that would be necessary to produce the same perturbation $\Delta F_{t,x}$ of the TOA flux through thermal emission. This illustrates how the Planck response compensates for the flux disequilibrium associated with each feedback.

b. *Transient temperature change*

Without dealing with the complexity of the feedback analysis under transient conditions (e.g., Hallegatte et al. 2006), we now consider the ocean response in a very simple way in order to quantify the feedback processes in transient runs using the same feedback framework as above. Following Gregory and Mitchell (1997), we assume that in transient experiments in which the forcing increases regularly with time, the disequilibrium ΔF_t of the net flux at the TOA is equal to the ocean heat uptake and is related to the surface temperature change ΔT_s by

$$\Delta F_t = -\kappa \Delta T_s, \tag{12}$$

where κ is the ocean heat uptake efficiency (<0). This assumption is common and useful despite its limited validity. For instance, it is valid neither when the climate tends toward equilibrium (ΔT_s increases slowly, whereas ΔF_t decreases to zero) nor immediately after applying an abrupt forcing ($\Delta T_s \approx 0$, whereas $\Delta F_t \approx \Delta Q_t$). Using Eqs. (1) and (12), the transient temperature change [also called the transient climate response (TCR)] can be expressed as

$$\Delta T_s^t = -\frac{\Delta Q_t}{\lambda + \kappa}. \tag{13}$$

Although the ocean heat uptake is not a feedback, the only difference between the expression of the equilibrium [Eq. (2)] and the transient [Eq. (13)] temperature changes is that in the later one, the ocean heat uptake efficiency κ is added to the total feedback parameter λ. Using the same approach as that for the equilibrium temperature, we thus require the total temperature change ΔT_s to be the sum of the temperature change resulting from the Planck response, climate feedbacks, and ocean heat uptake. We obtain the same equation as for the equilibrium temperature, except that the ocean uptake efficiency has to be added to the sum over x in Eqs. (6) and (9). The contribution $\Delta T_{s,x}$ of a feedback x

TABLE 1. The 2 × CO_2 radiative forcing ΔQ_t, total feedback parameter λ, and ocean heat uptake efficiency κ estimates of the 12 CMIP3/AR4 models used in this paper, and their multimodel mean and standard deviation.

Model	ΔQ_t ($W\ m^{-2}$)	λ ($W\ m^{-2}\ K^{-1}$)	κ ($W\ m^{-2}\ K^{-1}$)
Centre National de Recherches Météorologiques Coupled Global Climate Model, version 3	3.71	−1.17	−0.80
Geophysical Fluid Dynamics Laboratory Climate Model, version 2.0	3.50	−1.18	−0.53
Geophysical Fluid Dynamics Laboratory Climate Model version 2.1	3.50	−1.37	−0.81
Goddard Institute for Space Studies Model E-R	4.06	−1.64	−0.92
Institute of Numerical Mathematics Coupled Model, version 3.0	3.71	−1.46	−0.56
L'Institut Pierre-Simon Laplace Coupled Model, version 4	3.48	−0.98	−0.79
Model for Interdisciplinary Research on Climate 3.2, medium-resolution version	3.60	−0.91	−0.77
Meteorological Research Institute Coupled General Circulation Model, version 2.3.2	3.47	−1.50	−0.61
ECHAM5/Max Planck Institute Ocean Model	4.01	−0.88	−0.57
Community Climate System Model, version 3	3.95	−1.62	−0.70
Parallel Climate Model	3.71	−1.53	−0.62
Met Office Third Hadley Centre Coupled Ocean–Atmosphere General Circulation Model	3.81	−0.97	−0.59
Multimodel mean	3.71	−1.27	−0.69
Intermodel RMS	0.20	−0.27	0.12

to the global temperature change is then given by Eq. (10), where the gain g is replaced by $g' = g + g_o$ with $g_o = -(\kappa/\lambda p)$ and the contribution of the ocean heat uptake is given by $\Delta T_{s,o} = (g_o/1 - g')\Delta T_{s,P}$.

Because of the ocean heat uptake, g' differs from g, and the transient temperature change $\Delta T^t_{s,x}$ associated with a feedback x differs from that at equilibrium $\Delta T^e_{s,x}$. The transient temperature change ΔT^t_s also differs from that at equilibrium ΔT^e_s, and a direct consequence of Eq. (10) is that the contribution of a feedback x to the global temperature change is the same in both equilibrium and transient conditions,

$$\frac{\Delta T^e_{s,x}}{\Delta T^e_s} = \frac{\Delta T^t_{s,x}}{\Delta T^t_s} \quad \text{for} \quad x \neq P. \tag{14}$$

4. CMIP3/AR4 atmosphere–ocean GCMs

We now apply the above decomposition to the global surface temperature response to a CO_2 doubling predicted by an ensemble of 12 coupled atmosphere–ocean GCMs (AOGCMs) participating in the CMIP3/AR4 (Meehl et al. 2005; Randall et al. 2007a). For this purpose, for each model we need the global mean values of the radiative forcing, the climate feedback parameters, and the ocean heat uptake efficiency.

a. 2 × CO_2 radiative forcing

In this study we use the radiative forcing for a CO_2 doubling reported by Forster and Taylor (2006) and Randall et al. (2007b). These forcings have been computed after stratospheric adjustment in all-sky conditions and are averaged over the globe for a year (Table 1). For the 12 GCMs considered here, the multimodel average of the net radiative forcing (3.71W m^{-2}) is very close to previous Myhre et al. (1998) results, and the relative intermodel standard deviation is about 6% (Table 2).

In another intercomparison study, for 16 GCMs Collins et al. (2006) obtained an intermodel spread of the net radiative forcing as large as 15% (Randall et al. 2007a). These forcing have been computed at 200 hPa for a unique atmospheric profile (midlatitude summer climatological conditions) in clear-sky conditions, without any stratospheric adjustment. When compared with Forster and Taylor (2006) results, the relative values of

TABLE 2. Multimodel mean and intermodel standard deviation of the LW, SW, and net radiative forcing ($W\ m^{-2}$) for a CO_2 doubling computed by GCMs in two intercomparison studies, with two different numerical setups (see text). In parenthesis, the standard deviation is computed relative to the mean.

	LW		SW		Net	
	Mean	Std dev	Mean	Std dev	Mean	Std dev
Forster and Taylor (2006)	3.85	0.31 (8%)	−0.12	0.13 (100%)	3.75	0.23 (6%)
Collins et al. (2006)	5.07	0.43 (8%)	−0.79	0.28 (35%)	4.28	0.66 (15%)

TABLE 3. Multimodel mean and intermodel standard deviation of total feedback parameter λ and its components λ_x (W m^{-2} K^{-1}), the ocean heat uptake efficiency κ (W m^{-2} K^{-1}), the 2 × CO_2 radiative forcing ΔQ_t (W m^{-2}), and their associated equilibrium and transient temperature changes (°C). The multimodel mean and standard deviation of the equilibrium (ΔT_s^e) and transient (ΔT_s^t) temperature changes (°C) are also given.

Variables			Associated equilibrium temperature change		Associated transient temperature change	
	Mean	Std dev	Mean	Std dev	Mean	Std dev
Feedback parameter						
All λ	−1.27	0.30	3.1	0.7	2.4	0.4
Planck $_P$	−3.2	0.05	1.2	0.0	1.2	0.0
Water vapor λ_W	1.80	0.18	1.7	0.4	1.1	0.2
Lapse rate λ_l	−0.84	0.26	−0.8	0.3	−0.5	0.2
WV + LR λ_{WL}	0.96	0.11	0.9	0.2	0.6	0.1
Albedo λ_a	0.26	0.08	0.3	0.1	0.2	0.05
Clouds λ_c	0.69	0.38	0.7	0.5	0.4	0.3
OHU efficiency κ	−0.67	0.12	—	—	−0.4	0.1
Radiative forcing ΔQ_t	3.71	0.20	0	0.2	0.0	0.1
$\Delta T_s^e = \Delta Q_t/\lambda$	3.1	0.7				
$\Delta T_s^t = \Delta Q_t/(\lambda + \kappa)$	2.0	0.3				

the intermodel standard deviation of the longwave (LW) forcing are similar in both studies (8%, see Table 1). This is not the case in the shortwave (SW) domain, and the difference is even larger for the net radiative forcing. In the results of Collins et al. (2006), as reported by Randall et al. (2007a), the standard deviation of the net forcing is larger than the quadratic sum of the standard deviation of the SW and LW forcings, which indicates that the SW and LW intermodel differences are positively correlated. The opposite is found in Forster and Taylor (2006), which indicates that the error in the SW and LW domains are anticorrelated, and that stratospheric adjustment can explain part of it. We believe that the intermodel spread of the forcing reported by Forster and Taylor (2006) is the most relevant for our study because the global warming estimates are derived from global simulations, including clouds and a stratospheric temperature response.

All contributions to the global warming ΔT_s are proportional to $\Delta T_{s,P}$ [Eq. (10)], and therefore to the forcing ΔQ_t [Eq. (4)]. Part of intermodel differences in these contributions may thus arise from intermodel differences in the radiative forcing. To quantify this part, for each model we compute $\Delta T_{s,P}$ for a reference forcing value ΔQ_t^r (set to the multimodel mean forcing estimate, viz., 3.71W m^{-2}; see Table 1) and we add a term that represents the impact of the discrepancy δQ_t on ΔT_s between the actual forcing of each model and the reference value,

$$\Delta T_s = \frac{1}{1-g}\left(\frac{-\Delta Q_t^r}{\lambda p}\right) + \frac{1}{1-g}\left(\frac{-\delta Q_t}{\lambda p}\right). \quad (15)$$

b. Feedback parameters

As reviewed by different authors (e.g., Soden et al. 2004; Stephens 2005; Bony et al. 2006), several approaches have been followed to decompose the total feedback parameter into its several components (water vapor, clouds, surface albedo, etc.), with each method having its own strengths and weaknesses. Soden and Held (2006) computed these feedback parameters for 12 CMIP3/AR4 models (Table 1), using the Special Report on Emissions Scenarios (SRES) A1B simulations, and their results are fairly consistent with previous results obtained by Colman (2003) with older GCMs (cf. Bony et al. 2006). The multimodel mean and standard deviation of the total feedback parameters ($\overline{\lambda}$ = −1.3 W m^{-2}, σ_λ = 0.3 W m^{-2}, Table 3) are consistent with the values obtained by Forster and Taylor (2006) for a larger set of CMIP3/AR4 models and for different ensembles of runs. When analyzing the 1% yr^{-1} increase of CO_2 simulations performed by 20 AOGCMs, they found a multimodel mean value of the total feedback parameter $\overline{\lambda}$ = −1.4 W m^{-2} and a standard deviation σ_λ = 0.3 W m^{-2}. When considering another set of experiments, namely, doubled-CO_2 equilibrium runs from 11 atmospheric GCMs coupled to slab oceans, they found a mean value $\overline{\lambda}$ = −1.2 W m^{-2} and a standard deviation σ_λ = 0.3 W m^{-2}.

c. Ocean heat uptake efficiency

We computed the ocean heat uptake efficiency κ using Eq. (12). For each model, the TOA flux F_t and the surface air temperature T_s were averaged over the 20-yr period centered at the time of CO_2 doubling, that is,

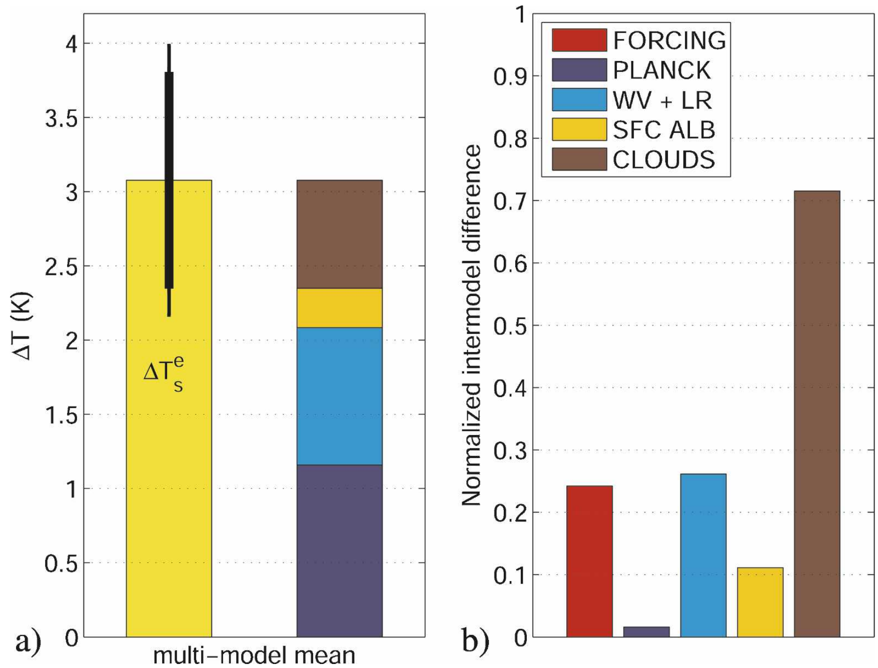

FIG. 1. For a CO_2 doubling, (a) multimodel mean ±1 standard deviation (thick line) and 5%–95% interval (thin line) of the equilibrium temperature change (ΔT_s^e), and contributions to this temperature change associated with the Planck response, combined water vapor and lapse-rate (WV + LR) feedback, surface albedo feedback, and cloud feedback. (b) Intermodel standard deviation of the temperature change estimates associated with the radiative forcing, the Planck response, and the various feedbacks normalized by the intermodel standard deviation of the equilibrium temperature change ΔT_s^e reported in (a).

year 70 for the 1% yr^{-1} simulation. The differences with the corresponding period of the control simulation were performed and the values of κ reported in Table 1.

d. Representativity of the ensemble of models considered

Using the values reported in Table 1, the equilibrium and transient temperature changes are computed for each of the 12 models as $\Delta T_s^e = -\Delta Q_t/\lambda$ and $\Delta T_s^t = -\Delta Q_t/(\lambda + \kappa)$, respectively. This leads to a multimodel mean ±1 standard deviation of the equilibrium temperature change of 3.1° ± 0.7°C. These numbers are comparable with those of the AR4 equilibrium climate sensitivity estimates derived from 18 atmospheric GCMs coupled to slab oceans (3.3° ± 0.7°C; see Meehl et al. 2007). For the transient temperature change, we obtain 2.0° ± 0.3°C, which is closed to the AR4 values reported on the basis of 19 coupled atmosphere–ocean GCMs (1.8° ± 0.3°C; see Meehl et al. 2007). As far as global temperature change is concerned, the subset of 12 models considered here is therefore representative of the larger set of CMIP3/AR4 models.

5. Results

a. Decomposition of equilibrium temperature changes

The multimodel mean of the equilibrium temperature change and the contributions associated with the Planck response [Eq. (4)] and each feedback [Eq. (10)], computed for a reference radiative forcing, are shown in Fig. 1a and reported in Table 3. On average, for the set of models considered here, the Planck response represents about a third of the total temperature response (1.2° versus 3.1°C), while climate feedbacks account for two-thirds of it. The increase of water vapor with warming enhances the absorption of longwave radiation and enhances the warming by 1.7°C. Lapse-rate changes are associated with a negative feedback, resulting from the moist-adiabatic structure of the tropical atmosphere. Because of the strong anticorrelation between these two feedbacks, it is convenient to consider the sum of both of them (WV + LR; Soden and Held 2006). This combined feedback increases the temperature by 0.9°C, which is slightly less than the Planck response. The cloud feedback's contribution to the warming is, on av-

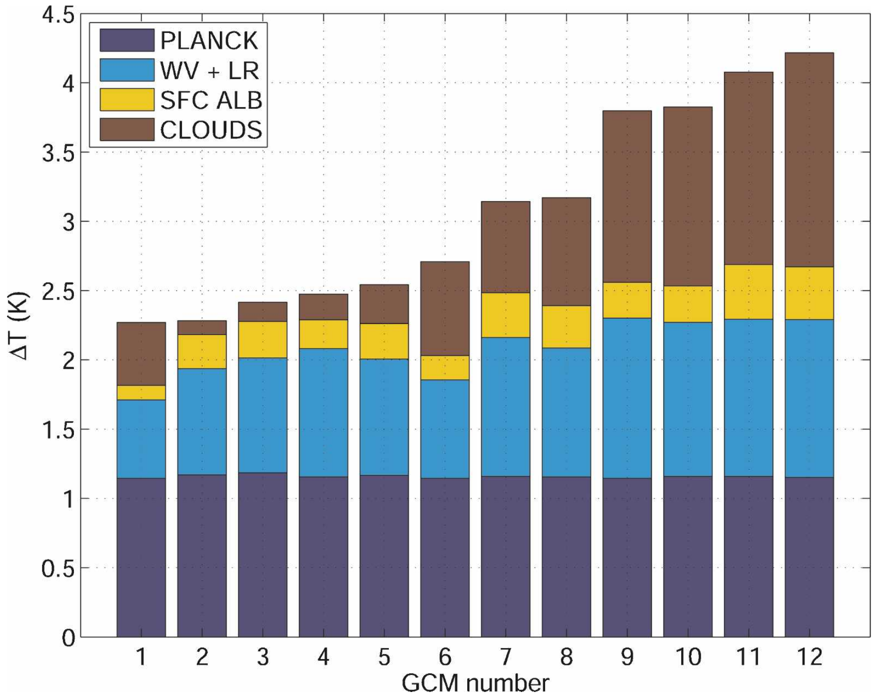

FIG. 2. Equilibrium temperature change associated with the Planck response and the various feedbacks, computed for 12 CMIP3/AR4 AOGCMs for a 2 × CO_2 forcing of reference (3.71 W m^{-2}). The GCMs are sorted according to ΔT_s^e.

erage, slightly weaker than that of the WV + LR feedback, and the surface albedo feedback's contribution is the smallest.

However, Fig. 2 shows that for each feedback there are some intermodel differences, especially for the cloud feedback contribution, and that the amplitude of the equilibrium temperature change is primarily driven by the cloud feedback component. This appears also clearly when considering the intermodel standard deviation of the temperature change resulting from each feedback normalized by the intermodel standard deviation of the total temperature change (Fig. 1b). The standard deviation resulting from cloud feedback represents nearly 70% the standard deviation of the total temperature change. The temperature spread resulting from the radiative forcing is comparable to the spread resulting from the WV + LR feedback and the spread resulting from the surface albedo feedback is the smallest.

b. *Decomposition of transient temperature changes*

The transient temperature changes (or TCR) from individual GCMs, as well as the contribution of the various feeedbacks, are displayed in Fig. 3. The multimodel mean and standard deviation are displayed in Fig. 4 and reported in Table 3. The temperature damping resulting from the ocean heat uptake is about −0.4°C, and its absolute value is comparable to the multimodel contributions of the WV + LR (0.6°C) and cloud (0.4°C) feedback. The mean transient temperature change is nearly 2/3 of that at equilibrium; therefore, the transient temperature changes associated with each feedback scale with it [cf. Eq. (14)]. The intermodel standard deviation of the temperature change resulting from cloud feedback represents nearly 90% of the standard deviation of the total temperature change (Fig. 4b). Similarly for the equilibrium case, cloud feedbacks thus constitute the main source of spread of the transient temperature response among GCMs. The WV + LR feedback, the ocean heat uptake, and the radiative forcing constitute secondary and roughly comparable sources of spread, and the surface albedo feedback constitutes the smallest one.

The intermodel standard deviation of the global temperature change may also be normalized with the multimodel mean global temperature change. This relative standard deviation is comparable in both equilibrium and transient conditions; the spread in equilibrium is slightly larger (23% versus 16%). The same holds for the relative standard deviation of the temperature change associated with each feedback. Therefore, the contribution of the various feedbacks to the total

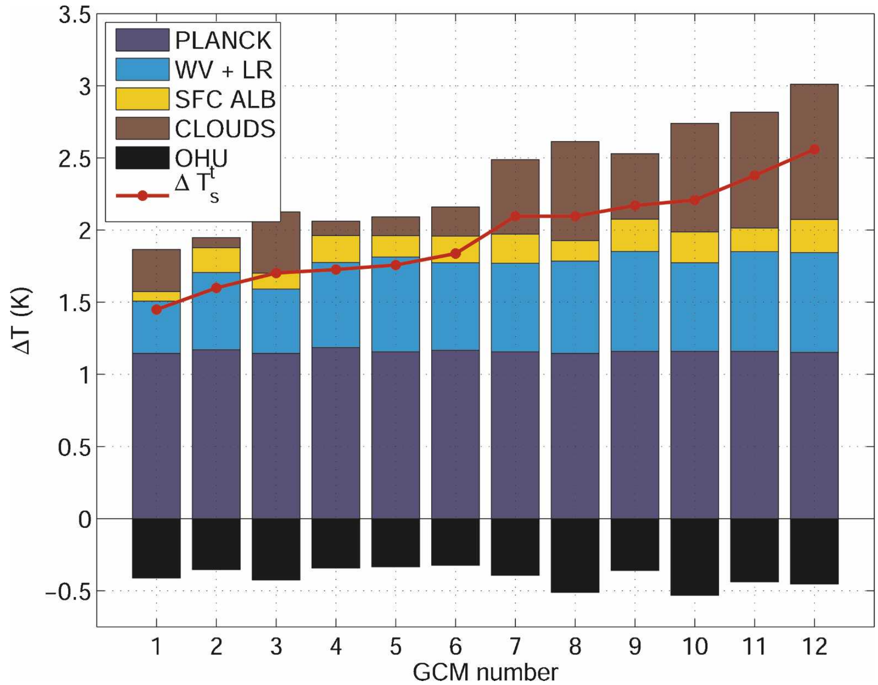

FIG. 3. Transient temperature change (ΔT_s^t or TCR, red line) and contributions to this temperature change associated with the Planck response, the ocean heat uptake (OHU), and the various feedbacks, computed for 12 CMIP3/AR4 AOGCMs for a 2 × CO_2 forcing of reference (3.71 W m^{-2}). The GCMs are sorted according to ΔT_s^t.

spread is, in relative terms, as important in the transient case as in the equilibrium case.

6. Summary and conclusions

In this note we propose a simple decomposition of the equilibrium and transient global temperature responses to an external forcing into a sum of contributions associated with the Planck response, the different climate feedbacks, and, eventually, the ocean heat uptake. This allows us to quantify how the various processes contribute to the multimodel mean and intermodel spread of the global temperature change. This is illustrated (Figs. 1–4) using published results for the feedback parameters and the radiative forcings (Soden and Held 2006; Forster and Taylor 2006; Randall et al. 2007b), and by diagnosing the ocean heat uptake efficiency from model outputs. In transient simulations, the absolute values of the contributions of the WV + LR feedback, the cloud feedbacks, and the ocean heat uptake to the global temperature response appears to be comparable (Fig. 4a). However, for the ensemble of models considered here, the spread of the transient temperature change resulting from intermodel differences appears to be primarily due to cloud feedback. The spread resulting from WV + LR feedback, ocean heat uptake, or radiative forcing appears to be of the same order of magnitude and roughly one-third of the spread resulting from the cloud feedback (Fig. 4b). Note that the radiative forcing associated with non-CO_2 greenhouse gases and aerosols is more uncertain than that associated with CO_2 (Forster et al. 2007). Therefore, the intermodel spread of radiative forcing estimates might be larger either for twentieth-century simulations or for climate change simulations, based on emission scenarios that include changes in aerosol concentrations, than in this study. This difference is mitigated, however, by the fact that the relative contribution of aerosols versus greenhouse gases is likely to decrease in the future (Dufresne et al. 2005).

Our analysis shows that the contribution of each feedback and of the radiative forcing to intermodel differences in temperature change is roughly similar, in a normalized sense, in equilibrium and transient simulations (Figs. 1b and 4b). In particular, cloud feedbacks appear to be the main source of spread in both cases. Intermodel differences in cloud feedbacks have been shown to arise primarily from the response of low-level clouds (Bony and Dufresne 2005; Webb et al. 2006; Wyant et al. 2006). Understanding and evaluating the physical processes that control these cloud responses thus appears to be of primary importance for better

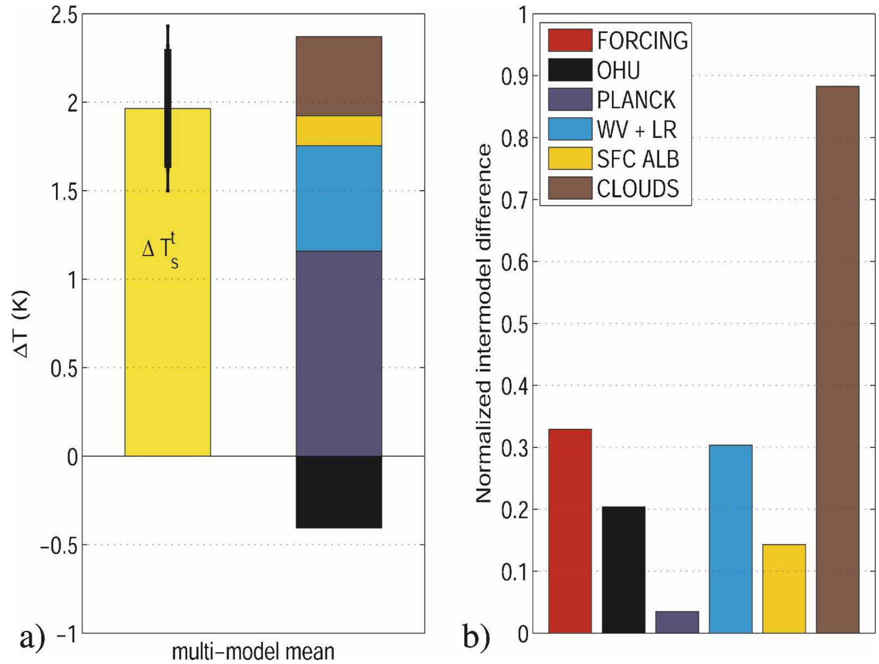

FIG. 4. For a CO_2 doubling, (a) multimodel mean ±1 standard deviation (thick line) and 5%–95% interval (thin line) of the transient temperature change (ΔT_s^t) and contributions to this temperature change associated with the Planck response, OHU, combined water vapor and lapse-rate (WV + LR) feedback, surface albedo feedback, and cloud feedback. (b) Intermodel standard deviation of the transient temperature change estimates associated with intermodel differences in radiative forcing, Planck response, ocean heat uptake, and the various feedbacks normalized by the intermodel standard deviation of the transient temperature change ΔT_s^t.

assessing the relative credibility of climate projections from the different models.

Acknowledgments. We thank Jean-Yves Grandpeix for frequent and useful discussions on feedback. We are grateful to Jonathan Gregory and an anonymous reviewer for useful comments on this study, and we acknowledge the modeling groups for making their simulations available for analysis, the Program for Climate Model Diagnosis and Intercomparison (PCMDI) for collecting and archiving the CMIP3 model output, and the WCRP's Working Group on Coupled Modelling (WGCM) for organizing the model data analysis activity. The WCRP CMIP3 multimodel dataset is supported by the Office of Science, U.S. Department of Energy. This research was supported by CNRS, byINSU-EVE French Program under the project MissTerre, and by the European Commission under the project ENSEMBLES (Contract GOCE-CT-2003-505539).

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