Importance of Ocean Heat Uptake Efficacy to Transient Climate Change

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ABSTRACT

This article proposes a modification to the standard forcing–feedback diagnostic energy balance model to account for 1) differences between effective and equilibrium climate sensitivities and 2) the variation of effective sensitivity over time in climate change experiments with coupled atmosphere–ocean climate models. In the spirit of Hansen et al. an efficacy factor is applied to the ocean heat uptake. Comparing the time evolution of the surface warming in high and low efficacy models demonstrates the role of this efficacy in the transient response to CO_2 forcing. Abrupt CO_2 increase experiments show that the large efficacy of the Geophysical Fluid Dynamics Laboratory's Climate Model version 2.1 (CM2.1) sets up in the first two decades following the increase in forcing. The use of an efficacy is necessary to fit this model's global mean temperature evolution in periods with both increasing and stable forcing. The intermodel correlation of transient climate response with ocean heat uptake efficacy is greater than its correlation with equilibrium climate sensitivity in an ensemble of climate models used for the third and fourth Intergovernmental Panel on Climate Change (IPCC) assessments. When computed at the time of doubling in the standard experiment with $1\%\ yr^{-1}$ increase in CO_2, the efficacy is variable amongst the models but is generally greater than 1, averages between 1.3 and 1.4, and is as large as 1.75 in several models.

1. Introduction

The familiar linear zero-dimensional energy balance model is a useful tool for summarizing and analyzing the response of global mean surface temperature to radiative forcing in simulations of forced climate change. Once tuned to a target atmosphere–ocean general circulation model (AOGCM), the hope is that the simple model can be used to predict how the AOGCM would respond to a large range of forcings (e.g., Solomon et al. 2007; Meinshausen et al. 2008).

The equilibrium climate sensitivity of the linear energy balance model is one of the key parameters adjusted to mimic the target AOGCM. However, rather than the equilibrium sensitivity, which is usually estimated using an atmosphere–slab-ocean model, an ''effective sensitivity'' (Murphy 1995) is often used for this exercise, determined from a transient run of the AOGCM (Solomon et al. 2007, Table S8.1), in an attempt to avoid inconsistencies between AOGCM and slab-ocean sensitivities. The effective sensitivity is obtained by scaling up the transient temperature response by the factor $R/(R - N)$, where R is the radiative forcing and N is the top-of-atmosphere heat uptake (we refer to this informally as the ocean heat uptake in the following since the two are nearly the same on the time scales of interest here). The great majority of AOGCMs with available data in the Intergovernmental Panel on Climate Change (IPCC) third and fourth assessment reports (Houghton et al. 2001; Solomon et al. 2007) have effective sensitivities less than their equilibrium sensitivities. However, several researchers have noted an increase in the effective sensitivity over the course of

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long climate change simulations (Senior and Mitchell 2000; Gregory et al. 2004), although this result is not universal (Watterson 2000; Boer and Yu 2003). The increase in effective sensitivity is expected as a model with an effective sensitivity less than its equilibrium sensitivity approaches equilibrium.

Williams et al. (2008) examined the relationship between the top-of-atmosphere (TOA) fluxes and the surface temperature change in the stabilized-CO_2 section of 1% CO_2 increase experiments and defined an ''effective forcing'' by extrapolating this relationship back to a zero temperature change. Six of the eight models they investigated had effective forcings that were less than the traditionally defined radiative forcings. They argue that this is evidence for a direct CO_2 effect on clouds, with fixed ocean temperatures, which modifies the ''forcing'' in analogy to the familiar direct stratospheric response. Gregory and Webb (2008) discuss an analogous analysis of slab-ocean models; however, the time scale of forcing adjustment in Williams et al. (2008) is on the order of decades, implicating oceanic adjustment as an important factor and favoring a feedback interpretation.

In this study, we propose an alternative interpretation of ''effective forcing'' and the time variation of ''effective sensitivity.'' We are inspired by Hansen et al. (2005), who noted that different forcing agents resulting in the same global mean radiative forcing can elicit different global mean temperature responses and accounted for this by introducing an efficacy factor associated with each forcing (see also Solomon et al. 2007, section 2.8.5). In this paper we note that an efficacy can also be applied to ocean heat uptake. It might seem perverse to treat ocean heat uptake as a forcing rather than a feedback when it is clearly internal to the climate system and likely varies with global mean temperature change. One way of rationalizing this approach is to consider a slab-ocean model in which one attempts to mimic the fully coupled system by specifying the heat flux exchanged between the deep ocean and the slab, putting aside the question of how the heat uptake is determined. A linear zero-dimensional model of this system would have as its inputs the heat uptake as well as the radiative forcing, leading one to consider the possibility of nonunitary efficacy of the heat uptake. We argue in the following that nonunitary efficacy of ocean heat uptake is a useful alternative to thinking in terms of effective forcings or the time variation of effective sensitivity.

In section 2, we present a model comparison that motivates the need for an ocean heat uptake efficacy and demonstrate that the feedbacks that apply to ocean forcing can be significantly different from those that apply to CO_2 forcing. In section 3, we define nondimensional quantities that allow us to compare efficacies when radiative forcing and equilibrium sensitivity vary in time and between models. In section 4 we look at the ability of ocean heat uptake efficacy to characterize the time evolution of the climate state in a particular model—the Geophysical Fluid Dynamics Laboratory (GFDL) Climate Model version 2.1 (CM2.1). The distribution of efficacies in the IPCC multimodel ensembles of idealized transient climate change experiments is discussed in section 5. The results are summarized in section 6.

2. The need for ocean heat uptake efficacy

The GFDL CM2.1 and Max-Planck-Institut (MPI) ECHAM5 AOGCMs both report equilibrium sensitivity to CO_2 doubling of 3.4°C. However, in transient experiments, the MPI model appears to be considerably more sensitive to forcing. The transient climate response (TCR), the temperature change at CO_2 doubling in a 1% yr^{-1} CO_2 increase experiment, is nearly 50% greater for the MPI model (Solomon et al. 2007, Table 8.2). The difference in transient sensitivity carries over to scenario forced experiments where the two models bracket the global temperature responses of most of the other models (Solomon et al. 2007, Fig. 10.5). Of 19 models listing transient climate responses in Solomon et al. (2007) only four fall outside the range bounded by these two models. How might the zero-dimensional energy balance model account for the difference in transient response? This model represents the time-varying global-mean net TOA radiative flux N, as the sum of a forcing R and a term proportional to the global mean surface temperature anomaly T:

$$N(t) = R(t) - \lambda T(t). \quad (1)$$

Here N is positive down and λ is the climate feedback parameter. The MPI and GFDL models have similar equilibrium sensitivity, R/λ, and similar radiative forcing R so according to (1), the larger temperature response in the MPI model should be accounted for by a smaller net heat uptake N.

Figure 1 shows the global warming and net TOA flux anomalies for the two models forced with 1% yr^{-1} CO_2 increase to doubling. Counter to (1), the MPI model has more heat uptake than the GFDL model. The heat uptake difference between the two models is evidently responding to the temperature change difference between the two models rather than forcing it, since the warmer model has more heat uptake. The twenty-first century simulations of the two models under Special Report on Emission Scenarios (SRES) B1, A1B, and A2 forcing scenarios show similar relationships (not shown). Raper et al. (2002) noted this tendency for models with

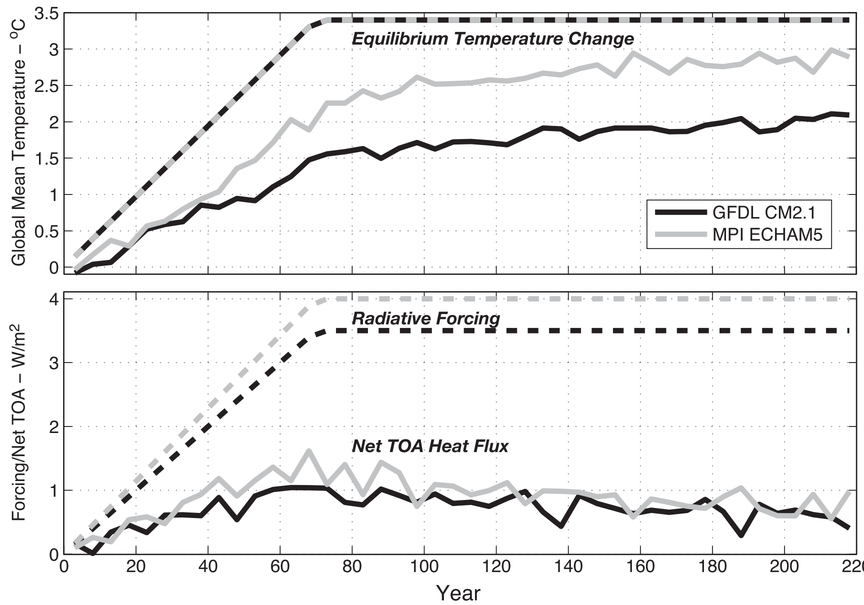

FIG. 1. (top) Global mean temperature anomaly and (bottom) net TOA radiation anomaly for the GFDL CM2.1 and MPI ECHAM5 AOGCMs forced with the 1% yr^{-1} CO_2 increase to doubling. Anomalies are taken relative to the mean of the first century of the preindustrial control runs. The equilibrium temperature change and radiative forcing are taken from Solomon et al. (2007).

larger transient warming to simulate larger heat uptake, relating the two quantities linearly with a constant of proportionality, which they term "ocean heat uptake efficiency."

In this paper, we take the view that the difference in transient response between these two models arises because the models respond differently to a given quantity of heat uptake. We do not propose a model for the ocean heat uptake itself as in Raper et al. (2002), but are concerned instead with its impact on climate change, which, as the previous example shows, varies between models. To evaluate this impact we break the transient temperature change into the sum of an equilibrium temperature change T_{EQ} and disequilibrium temperature difference $T_{EQ} - T$ driven by ocean heat uptake but with a feedback parameter that is smaller than that for CO_2 forcing by a factor of ε, the ocean heat uptake efficacy. We write the two equations for the different responses to radiative and heat uptake forcing as

$$T_{EQ}(t) = R(t)/\lambda \quad \text{and} \tag{2}$$

$$T_{EQ}(t) - T(t) = \varepsilon N(t)/\lambda. \tag{3}$$

One can think of these two equations as respectively representing the responses of an atmosphere–slab-ocean model, on longer time scales compared to the equilibrium time scale of the slab, to the CO_2 perturbation and to the heat uptake. Subtracting (3) from (2) gives

$$T(t) = [R(t) - \varepsilon N(t)]/\lambda. \tag{4}$$

While R and λ are similar for the GFDL and MPI AOGCMs, ε is larger for the GFDL model causing it to have a smaller transient response T. A large efficacy magnifies the effect of the heat uptake in the GFDL model and so its response lags that of the MPI model in time (Fig. 1).

Equation (4) is a generalization of (2), which is its $\varepsilon = 1$ special case. By applying a factor to the ocean heat uptake in (4) we have not sacrificed conservation of energy. As (3) shows, ε modifies the feedback operating on ocean heat uptake. It is simply a matter convenience to attach it as a factor to N.

Hansen et al. (1997) show that the geographical structure of a radiative forcing is an important source of nonunitary efficacy. They show that forcings focused at the surface at high latitudes have the greatest impact on temperature and therefore the larger efficacy. The ocean heat uptake occurs at the surface, of course, and it is largest in the subpolar oceans. Figure 2 compares the doubled CO_2 radiative forcing with the ocean heat uptake at doubling in the 1% yr^{-1} CO_2 increase experiment with the GFDL model. It is clear that the ocean

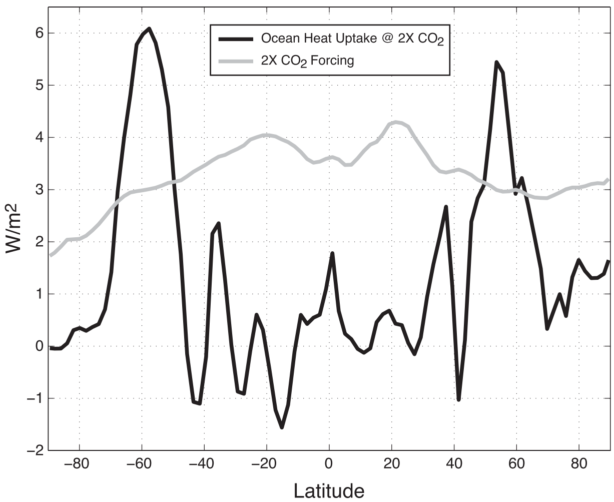

FIG. 2. Zonal mean doubled CO_2 radiative forcing and ocean heat uptake at doubling in the 1% yr^{-1} CO_2 increase experiment of GFDL CM2.1.

heat uptake is enhanced at high latitudes while CO_2 forcing is somewhat larger in the tropics. Therefore, the expectation is that ocean surface heat flux will tend to have an efficacy greater than 1.

While the first 70 years of the 1% yr^{-1} CO_2 increase to doubling experiment contains responses to changes in both radiative forcing and heat uptake, the subsequent stabilization period gives us an opportunity to look at the response to changing ocean heat uptake in isolation. This experiment has been run for 600 years with the GFDL model and its global mean warming over the 530-year stabilization period is about the same as in the initial CO_2 increasing period. The efficacy in the stabilization period is about 2—the model is twice as sensitive to ocean heat uptake as it is to CO_2 forcing, implying that that the feedback parameter for ocean heat uptake, λ/ε, is one-half that for CO_2 forcing, λ [Eq. (3)]. To determine the sources of this difference, we evaluate feedbacks for the transient run stabilization period and the atmosphere–slab-ocean doubled CO_2 experiment using the kernel method of Soden and Held (2006) and the GFDL model radiative kernel. The results are shown in Table 1. The total feedback is about -1 W m^{-2} K^{-1} for CO_2 forcing and -0.5 W m^{-2} K^{-1} for ocean heat uptake. The 0.5 W m^{-2} K^{-1} difference comes from the increased positive cloud and albedo feedbacks and a decreased negative temperature feedbacks in response to ocean forcing. Several studies show that water vapor and temperature feedbacks are tightly coupled through the maintenance of constant relative humidity (Zhang et al. 1994; Soden and Held 2006). This motivates combining of the temperature and water vapor feedback in Table 1 to avoid cancellation of large terms of

TABLE 1. GFDL CM2.1 radiative feedbacks (W m^{-2} K^{-1}).

Feedback (W m^{-2} K^{-1})	CO_2 forced	Ocean heat uptake forced	CO_2 – OHU
Temperature + water vapor	−2.14	−1.96	−0.18
Albedo	0.31	0.39	−0.08
Cloud	0.81	1.04	−0.23
Total	−1.03	−0.53	−0.50

opposite sign. The sum of these two increases the ocean heat uptake efficacy somewhat more than does the albedo feedback difference but less than the cloud feedback difference. Thus, the reasons for ocean heat uptake efficacy are distributed among the individual feedbacks with cloud feedback making the largest contribution, about 50% of the total difference in feedback, after combining temperature and water vapor feedbacks.

Having compared CO_2 and ocean heat uptake forcings and feedbacks, we turn now to their temperature responses. We can use the long stabilization period to estimate the contribution of ocean heat uptake to the SST changes at the time of CO_2 doubling. The first step is to estimate the equilibrium response by extrapolating SST from the time of doubling to equilibrium ($N = 0$) using the change with N over the stabilization period as the slope:

$$\mathrm{SST}(\infty) \approx \mathrm{SST}(70) + [\mathrm{SST}(590) - \mathrm{SST}(70)] \times \frac{N(70)}{N(70) - N(590)}, \tag{5}$$

where all fields represent 20-yr averages centered at the given times and differences from the preindustrial control climate values. The estimated SST equilibrium response is shown in the top panel of Fig. 3. It is in general agreement with the equilibrium response of the atmosphere–slab-ocean model to CO_2 doubling (not shown) although the coupled model pattern has more fine structure due to changes in currents.

Having obtained the equilibrium SST response in this way we can use it to obtain the response forced by ocean heat uptake. Consistent with Eq. (3), this is simply the transient response (Fig. 3, bottom panel) minus the equilibrium response and is shown in the middle panel of Fig. 3. Both ocean and CO_2 forcing induce SST responses that are amplified in the subpolar regions. However the ocean forcing induces a stronger subpolar response so that the total response has a minimum of warming in these regions where the ocean forcing is dominant. This pattern is a common feature of transient simulations

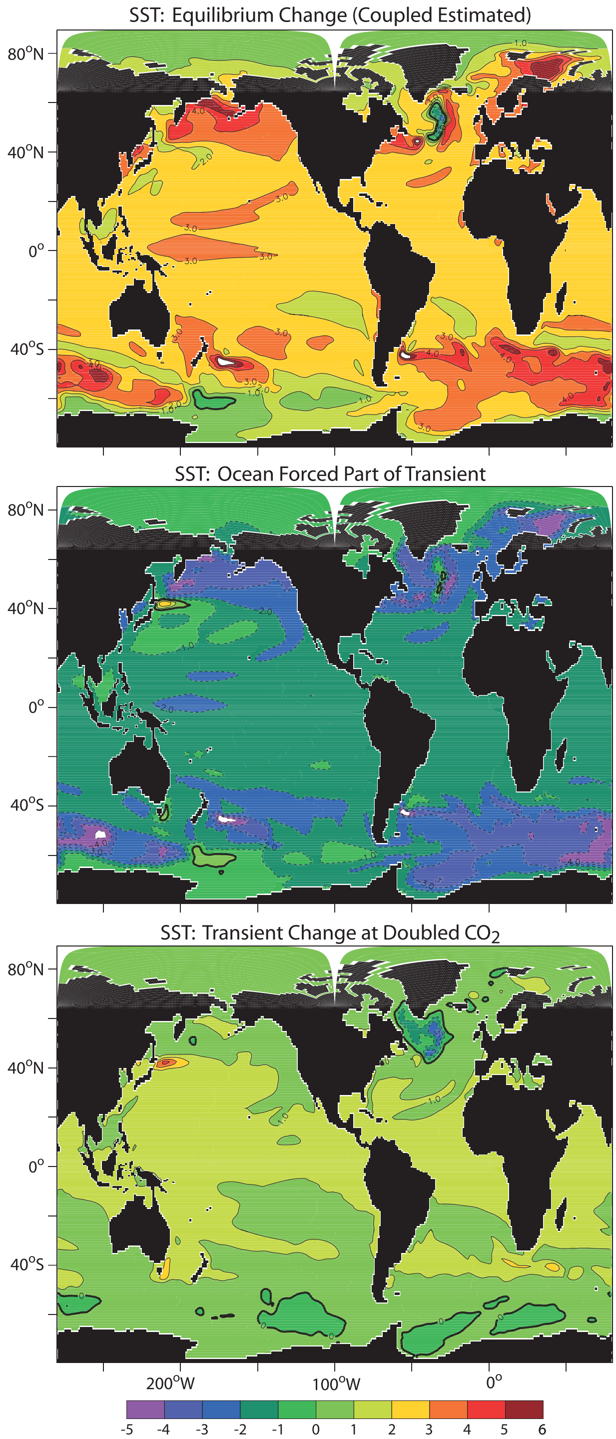

FIG. 3. (top) SST equilibrium response (°C) to a CO_2 doubling estimated from a long coupled model run of a 1% yr^{-1} CO_2 increase to doubling experiment using Eq. (5), (middle) ocean heat uptake forced component of the transient response at CO_2 doubling, and (bottom) the transient response at CO_2 doubling, which is sum of the (top) and (middle).

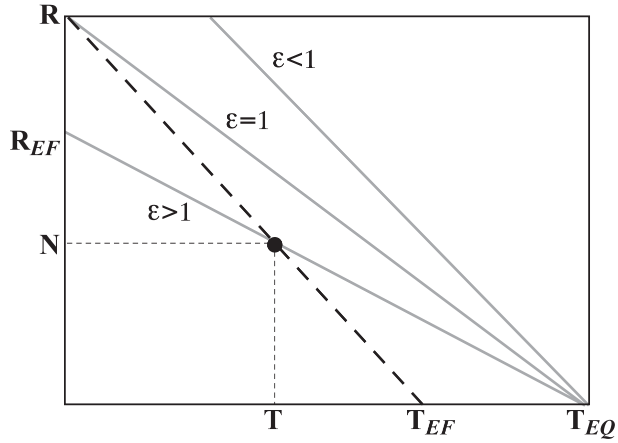

FIG. 4. Schematic relationships between radiative forcing R, equilibrium climate sensitivity T_{EQ}, effective climate sensitivity T_{EF}, effective forcing R_{EF}, and ocean flux efficacy ε, on a plot of global mean temperature T, against net TOA heat flux N, which is nearly equal to the net ocean heat flux over climatological time scales. If the climate state traverses the thick gray line between $[0, R]$ and $[T_{EQ}, 0]$, $R_{EF} = R$, $\varepsilon = 1$, and $T_{EF} = T_{EQ}$; T_{EF}, R_{EF}, and ε are different ways of accounting for deviations of the climate state from this path.

and appears in the Solomon et al. (2007) multimodel mean. The pattern was noted by Manabe et al. (1991) who showed that the deep mixed layers and large isopycnal mixing in the Southern Ocean and in the North Atlantic lead to minima in the ratio of transient to equilibrium response in those regions.

3. Ocean heat uptake efficacy with variable forcing and sensitivity

The difference in efficacy of the two models discussed in the last section was apparent because the models had similar equilibrium climate sensitivities and similar radiative forcings. We would also like to compare models with different sensitivities and also evaluate the efficacy in a single model, over time. For this purpose, we define the climate state, as illustrated schematically in Fig. 4, as consisting of the transient temperature change T relative to the equilibrium value T_{EQ} on the x axis, and the net heat uptake N relative to the radiative forcing R on the y axis.

The transient response to doubling $T(2\times)$ is conventionally evaluated as a 20-year average centered on year 70 in a 1% yr^{-1} increase of CO_2 experiment. At this time there will be a significant net flux $N(2\times)$. It will prove useful to also consider the hypothetical equilibrium response $T_{EQ}(t)$ that would result if the climate system adjusted instantly to the time-varying forcing. Since the equilibrium response to CO_2 is approximately linear in its forcing magnitude (Hansen et al. 2005), a reasonable approximation is

$$T_{EQ}(t) = T_{EQ}(2\times)R(t)/R(2\times). \tag{6}$$

The equilibrium response $T_{EQ}(2\times)$ is assumed to be evaluated by integrating or extrapolating until $N = 0$ (or using a slab-ocean model to approximate this value), so that (6) and (2) imply

$$\lambda = R(2\times)/T_{EQ}(2\times). \tag{7}$$

Employing $R(t)$ and $T_{EQ}(t)$ as scales we can use (6) and (7) to rewrite (4) as

$$\varepsilon N(t)/R(t) = 1 - T(t)/T_{EQ}(t). \tag{8}$$

This relationship is depicted on Fig. 4 by the shaded lines. The case $\varepsilon = 1$ is shown as the straight line between $[0, R]$ and $[T_{EQ}, 0]$. Climate states above this line have $\varepsilon < 1$ and those below have $\varepsilon > 1$. Lines of constant efficacy intersect at $[T_{EQ}, 0]$. The quantities T/T_{EQ} and N/R in (8) will be used in the following two sections to compare climate states in a single model as forcing varies over time (section 4) and compare models with different radiative forcings and climate sensitivities at the same point in time (section 5).

The effective climate sensitivity (Murphy 1995) is the extrapolation of the transient climate state from $[0, R]$ through $[T, N]$ to $N = 0$ (Fig. 4):

$$T_{EF}(t) = T(t)/[1 - N(t)/R(t)]. \tag{9}$$

A difference between $T_{EF}(2\times)$ and $T_{EQ}(2\times)$ implies a change in T_{EF} over time in order for $T_{EF}(t) \rightarrow T_{EQ}(2\times)$ as $N \rightarrow 0$. Effective sensitivity will always vary in time unless the system stays on the line between $[0, R]$ and $[T_{EQ}, 0]$.

The effective forcing, R_{EF} (Williams et al. 2008), is obtained by extrapolating the climate state back to $T = 0$ using the CO_2 stabilized section of an AOGCM experiment. From (8), we can write this in terms of the efficacy as

$$R_{EF} = R(2\times)/\varepsilon. \tag{10}$$

Thus effective forcing and efficacy of ocean heat uptake are closely related quantities, and both can be used to describe the time variation of the effective sensitivity. Our preference is for the concept of efficacy because it clearly ties this differential response to the nature of the forcing. In our view the "tropospheric response" described by Williams et al. is primarily an intrinsically coupled ocean–atmosphere transient phenomenon associated with the geographic pattern of ocean heat uptake rather than an atmosphere-only response analogous to the stratospheric adjustment to increased CO_2. Fast tropospheric responses analogous to stratospheric adjustment are possible, and can be isolated in the switch-on mixed-layer simulations of Gregory and Webb (2008), or in fixed SST experiments with imposed forcings. This fast response is small in CM2.1 and is not related to the efficacy as defined here.

4. Time evolution of efficacy in the GFDL CM2.1

We now focus on the time variation of the climate state in a single model, the GFDL CM2.1. This model has the largest efficacy at CO_2 doubling of any model in the IPCC ensemble presented in the next section. We employ the time-varying radiative forcing and equilibrium temperature change as scalings for the net TOA flux and transient temperature change, respectively, as in (8), to show the evolution of climate state over two 600-year experiments with a 1% yr^{-1} CO_2 increase to doubling and to quadrupling. The results are shown in Fig. 5. The two experiments pass through a similar arc of states that have increasing efficacy over time, more rapidly at first, and then more gradually. There is a somewhat narrower band of efficacies in this arc in the 1% to 4 times CO_2 experiment, presumably because of the larger signal-to-noise ratio. The scatter is larger in the transient forcing period of the experiments for the same reason. The two experiments are in reasonable agreement in this normalized climate state space, even in the band of states where the 1% to doubling experiment has stabilized forcing but the 1% to quadrupling experiment forcing is still increasing. The efficacy does not appear to be very sensitive to forcing history. The convergence of the model state on the lower right corner of Fig. 5 is an indication of the agreement of the coupled and slab-ocean equilibrium sensitivities since the slab-ocean value has been used to normalize the temperature axis.

The descent from the $\varepsilon = 1$ line in the 1% yr^{-1} CO_2 increase experiments is seen to occur in the early pentads, while the forcing is ramping up. It should be noted that forcing increases alone do not induce efficacy—nonunitary efficacy develops as a climate response to forcing changes. To explore this early adjustment further, we also show the ensemble mean of four instantaneous CO_2 doubling experiments with the same model. The first four pentads of this ensemble mean are denoted on Fig. 5 with the numbers 1 through 4. These show that the efficacy approaches a value near two within the first two decades. During this period a pattern is established of sea surface temperature change with reduced warming, and even some areas of cooling, in the subpolar North

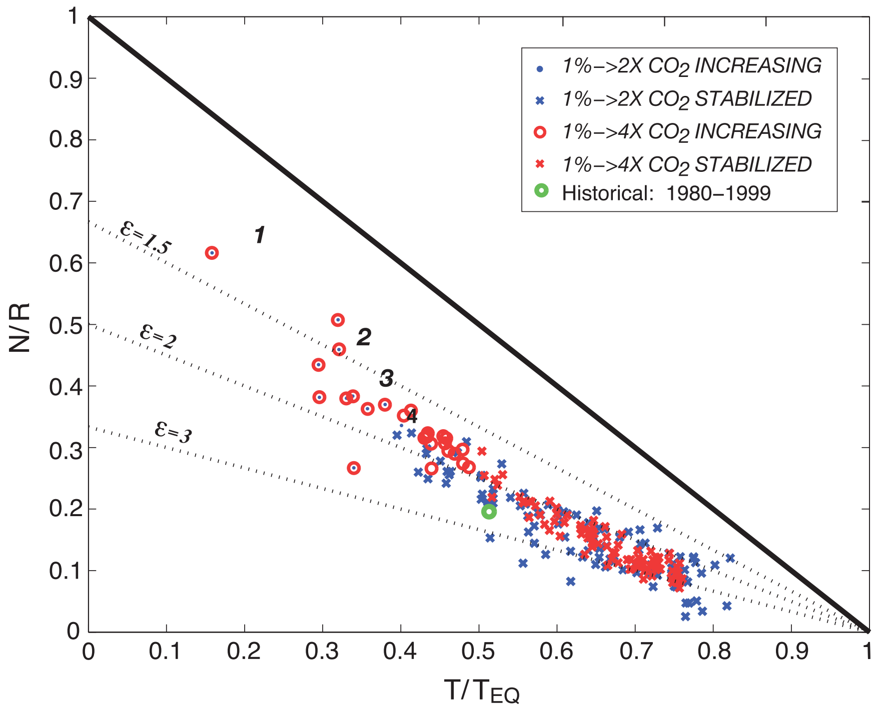

FIG. 5. A scatterplot of scaled global mean temperature T/T_{EQ} against scaled TOA net heat flux N/R for the 1% CO_2 increase to doubling and quadrupling experiments with the GFDL CM2.1 climate model. All points are pentadal averages. The first three pentads of the 1% yr^{-1} experiments fall outside the box due to smallness of the forced response early in the experiments. The numbers represent pentadal and four-member ensemble means from the first 20 years of an instantaneous CO_2 doubling experiment with the same model. The green circle shows the mean state of five-member ensemble of twentieth-century runs between 1980 and 1999.

Atlantic and Southern oceans as the competing cooling effect of ocean heat uptake overcomes the radiatively forced response in these areas. The ocean warming that occurs in this early adjustment period is confined to the mixed layer and nearby regions.

Figure 5 also shows the mean state of a five-member ensemble of twentieth-century runs of CM2.1 averaged over the period 1980 to 1999 relative to an 1860 control run. The forcing is calculated as the change in top-of-atmosphere flux between two ensembles of fixed SST experiments, one with time-varying forcing agents and one with tine-invariant forcing agents. The late-twentieth-century climate state indicates that this model's high efficacy is applicable to the historical period as well as to idealized forcing experiments. An important implication is that accurate measurements of the temperature change, forcing and ocean heat uptake associated with anthropogenic forcing in the current climate will not be sufficient to determine the equilibrium climate sensitivity and the committed warming if the actual heat uptake efficacy is significantly different from unity, as it is in this model.

The 600-year time series of pentadal and global mean temperature changes for the CM2.1 1% yr^{-1} to doubling experiment is shown in Fig. 6. This time series shows that about half of the total warming occurs in the CO_2-stabilized portion of the run. If we use Eq. (8) for the fit, taking the heat uptake from the model itself, with an efficacy of 1, there is too much warming in the CO_2 increasing period and not enough in the CO_2 stabilized period although this fit seems to be approaching the AOGCM's temperature at the end of the experiment. If we use an effective sensitivity in place of the equilibrium sensitivity in the equation, as is commonly done in reduced model fits to AOGCMs, a similar but smaller bias is apparent early in the run. Although this fit works reasonably well in the first two centuries, it has insufficient temperature increase in the final 400 years of the experiment.

The use of an efficacy allows us to fit both the CO_2 increasing and CO_2 stabilized portions of the time series. However, applying the efficacy naively to all time scales has the effect of increasing the amplitude of short-term temperature variations. This suggests that these short-term variations in N are not subject to the same efficacy as the longer-term variations, as would be plausible if these are not as concentrated in high latitudes as the long-term evolution of N–El Niño Southern Oscillation

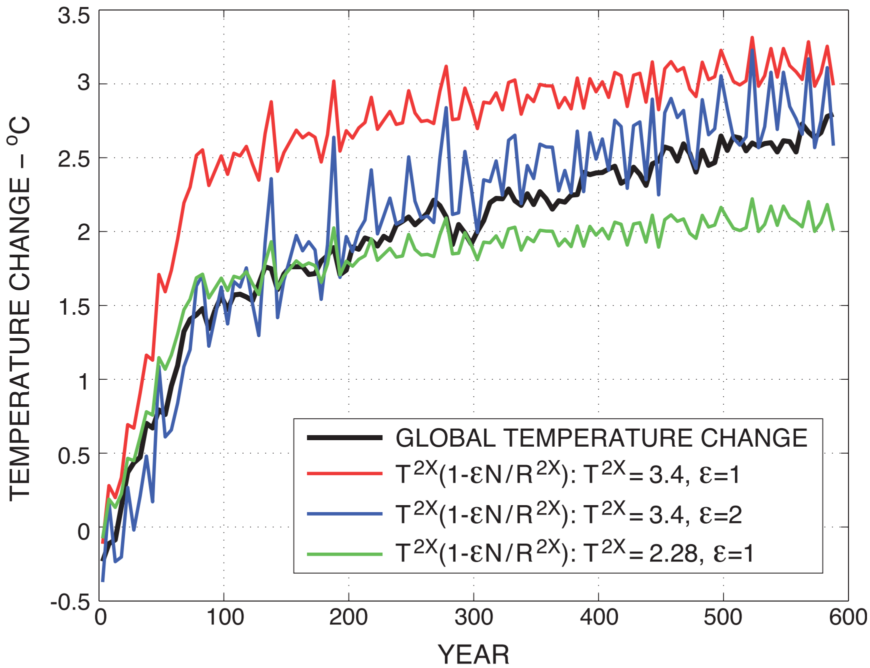

FIG. 6. Time series of pentadally averaged global mean temperature change in the 1% yr^{-1} CO_2 increase to doubling experiment of the GFDL CM2.1 climate model. The plot also shows estimates of the transient temperature change using Eq. (8) with an efficacy of 2 and Eq. (9) with an effective climate sensitivity of 2.28°C.

(ENSO) variability for example. An efficacy parameter should be useful in the simple models that are fit to AOGCMs when it is desirable to capture the long-term behavior of the AOGCM.

5. Multimodel transient efficacies at CO_2 doubling

We turn now from the time evolution of efficacy in a single model to the intermodel variation of efficacy and related parameters in the IPCC AOGCMs, evaluated at CO_2 doubling in 1% yr^{-1} CO_2 increase experiments. The IPCC reports contain doubled CO_2 forcing, transient climate response, equilibrium sensitivity, and effective sensitivity for a large number of AOGCMs. For a group of 14 models used for the Solomon et al. (2007) with doubled CO_2 forcing $R(2\times)$ available in the report, we used the published equilibrium climate sensitivity values in conjunction with $T(2\times)$ and $N(2\times)$ calculated from Coupled Model Intercomparison Project (CMIP3) data. For a group of eight Houghton et al. (2001) models, we used the published values of $T(2\times)$, $T_{EQ}(2\times)$, and $T_{EF}(2\times)$ to calculate $N(2\times)/R(2\times)$ using

$$N(t)/R(t) = 1 - T(t)/T_{EF}(t), \tag{11}$$

obtained by rearranging (9), and $\varepsilon(2\times)$ using

$$\varepsilon(t) = [1 - T(t)/T_{EQ}(t)]/[1 - T(t)/T_{EF}(t)] \tag{12}$$

obtained by combining (8) and (9). To calculate N from N/R we use the doubled CO_2 radiative forcing listed in Houghton et al. (2001) when available and the mean of the other models, when not. The models used and their parameter values are listed in Table 2. There are some small differences between the transient climate responses in Table 2, and those given in the report for the Fourth Assessment Report (AR4) models owing to differences in the treatment of the control. Here we have used averages over the 140-year period of the control run centered on the time of doubling in the perturbation run. The parameters in the table correspond to terms

$$T = T_{EQ}[1 - (\varepsilon N/R)], \tag{13}$$

in which is Eq. (8) rearranged. The intermodel correlations of these parameters are shown in Table 3. Figure 7 shows climate model transient climate responses (normalized by their equilibrium sensitivities) and net top-of-atmosphere fluxes (normalized by their radiative forcings) at CO_2 doubling. Figure 7 and Table 2 show that the great majority the models have an efficacy >1. The mean efficacy is 1.34. The two generations of models have similar distributions of efficacies.

TABLE 2. IPCC AOGCM global parameters [Eq. (13)] at CO_2 doubling in 1% yr^{-1} CO_2 increase experiment.

Model[a]	TCR K	T_{EQ} K	N W m^{-2}	R W m^{-2}	ε
Canadian Centre for Climate Modelling and Analysis (CCCma) Canadian Global Coupled Model (CGCM1)[b]	2.0	3.5	1.64	3.60	0.97
CSIRO MK2	2.0	4.3	1.59	3.45	1.16
NCAR CSM1	1.4	2.1	0.89	3.60	1.29
GFDL R15a[b]	2.1	3.7	1.76	3.60	0.86
United Kingdom Met Office (UKMO) HADCM2	1.7	4.1	1.11	3.47	1.83
MRI1[b]	1.6	4.8	1.30	3.60	1.85
MRI2[b]	1.1	2.0	0.96	3.60	1.69
Department of Energy (DOE) Parallel Climate Model (PCM)	1.3	2.1	0.91	3.60	1.56
CCCma CGCM3.1	1.9	3.4	1.17	3.32	1.25
CSIRO MK3.0	1.5	3.1	1.32	3.47	1.39
GISS MODEL EH	1.6	2.7	1.27	4.06	1.34
GISS MODEL ER[c]	1.5	2.7	1.50	4.06	1.19
GFDL CM2.0	1.5	2.9	0.92	3.50	1.85
GFDL CM2.1	1.5	3.4	1.00	3.50	1.99
Institut Pierre Simon Laplace (IPSL) CM4	2.1	4.4	1.57	3.48	1.18
MPI ECHAM5	2.1	3.4	1.31	4.01	1.16
Model for Interdisciplinary Research on Climate (MIROC 3.2) MEDRES	2.0	4.0	1.63	3.09	0.94
MIROC 3.2 HIRES	2.7	4.3	1.61	3.14	0.74
MRI CGCM2.3.2A	2.2	3.2	1.21	3.47	0.92
NCAR CCSM3.0	1.5	2.7	1.10	3.95	1.65
UKMO HADCM3[d]	2.1	3.8	1.23	3.81	1.42
UKMO HADGEM1[d]	1.9	3.4	1.34	3.78	1.25
Mean	1.8	3.4	1.29	3.60	1.34
Standard deviation	0.37	0.78	0.27	0.26	0.35
Coefficient of variable	0.21	0.23	0.21	0.07	0.26

[a] TAR model (italicized) TCRs and T_{EQ}s are taken from Table 9.1 of Houghton et al. (2001); AR4 model T_{EQ}s are from Table 8.2 of Solomon et al. (2007) except where noted. The Center for Climate Studies Research (CCSR) NIES2 model was available in Table 9.2 of Houghton et al. (2001) but has been excluded from this study because its effective sensitivity of 11.6°C makes it an outlier in the combined ensemble. The global temperature and net TOA flux changes for the AR4 models were calculated from CMIP3 database data using the differences between 20-year averages taken at CO_2 doubling and a 140-year period, centered on the time of doubling, from the control runs. The doubled CO_2 forcings are taken from Table 10.2 of Solomon et al. (2007).

[b] R was not available. The mean R of the reporting models (3.6 W m^{-2}) was used.

[c] Year 70 in 1% CO_2 increase per year to 4× experiment was used.

[d] T_{EQ}s estimated by extrapolation from long transient experiments taken from Table 2 of Williams et al. (2008).

Table 3 shows that the radiative forcing has little correlation with transient and equilibrium warming. The methodology for computing radiative forcings in not fully standardized, and it is likely that the intermodel spread of forcing values would be smaller with more standardization, so it is encouraging that the mean and standard deviation of efficacy in the models is not altered substantially if one substitutes a uniform value of the forcing for the tabulated values. Some of the lowest values of efficacy are eliminated if one uses a uniform forcing, however.

The quantities T_{EQ}, ε, and N are well correlated with T, the transient climate response and the sign of the correlations is such that T_{EQ} and ε variations enhance the intermodel TCR differences while N variations damp them. Of these, T_{EQ} is the most difficult to diagnose. Because (13) defines ε, it would be possible for ε to capture spurious variance from misdiagnosis of T_{EQ}. The lack of correlation between the two parameters allays this concern. We conclude that efficacy is an important driver of intermodel TCR variance in addition to, and relatively independent of, the equilibrium sensitivity.

The right side of (13), $1 - \varepsilon N/R$, is anticorrelated with T_{EQ} ($\rho = -0.56$) but poorly correlated with TCR ($\rho = 0.20$). It is of interest that ε can have a stronger association with TCR than T_{EQ} in spite of its confinement

TABLE 3. IPCC AOGCM global parameter intermodel correlations [Eq. (13)].

	TCR	T_{EQ}	N	R	ε
TCR	1	0.68	0.71	−0.33	−0.74
T_{EQ}		1	0.62	−0.41	−0.20
N			1	−0.16	−0.74
R				1	0.19
ε					1

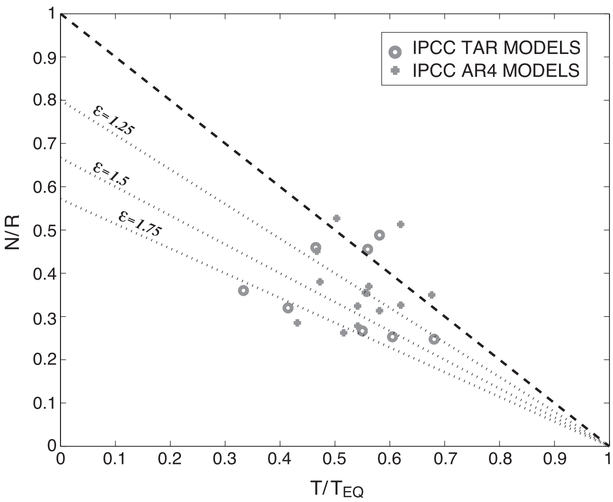

FIG. 7. Scatterplot of global mean temperature at CO_2 doubling scaled by the equilibrium temperature change (T/T_{EQ}) against net TOA heat flux scaled by the doubled CO_2 radiative forcing (N/R) for 22 climate models used in the IPCC third and fourth assessment reports. Twenty-year means centered on year 70 of the 1% yr^{-1} CO_2 increase to doubling experiments are used for the estimates.

within a term that has a poor correlation with TCR. The key is correlation of N with the other parameters. Following Gregory and Mitchell (1997) and Raper et al. (2002) and noting the intermodel correlation of N and TCR, we make use of the ocean heat uptake efficiency, γ, defined by

$$N = \gamma T. \qquad (14)$$

Note that efficiency represents ocean mixing processes while efficacy represents radiative processes in response to ocean heat flux. The heat uptake parameterization (14) essentially treats the ocean as an infinite reservoir, it does not account for impact of ocean warming on reducing N that is evident in the stabilized CO_2 section of the experiments (Fig. 1), for example. Nevertheless, it is useful for comparing models when forcing is rapidly increasing and a dynamic balance is established between radiative forcing of temperature anomalies and the sum of their damping to space and to the deep ocean. Using (14) in (13), along with $\lambda = R/T_{EQ}$, we obtain an alternative expression for the degree of equilibration:

$$T/T_{EQ} = 1 - (\varepsilon N/R) = 1/(1 + \varepsilon\gamma/\lambda). \qquad (15)$$

This expression is a generalization for efficacy of a similar relationship derived in Raper et al. (2002). As was noted by Raper et al. and others, more sensitive models, with smaller λ, have less equilibrated surface temperature responses. The appearance of λ ($=R/T_{EQ}$) in (15) is a source of anticorrelation between T_{EQ} and $1 - \varepsilon N/R$.

TABLE 4. IPCC AOGCM global parameter intermodel correlations [Eq. (15)].

	TCR/T_{EQ}	λ	ε	γ
TCR/T_{EQ}	1	0.50	−0.57	−0.32
λ		1	0.26	0.18
ε			1	−0.03
γ				1

While λ affects TCR through T_{EQ} as well as through the degree of equilibration, ε and γ have their impact on the TCR entirely through the degree of equilibration. Table 4 shows the correlation of these three parameters with T/T_{EQ}. The signs of the correlations are consistent with (15). Efficacy has the largest correlation with TCR/T_{EQ} but little correlation with the efficiency γ. The correlation of ε with N (Table 3) is apparently accounted for by its correlation with TCR after assuming (14). In this view, the anticorrelation between ε and N comes about because efficacy reduces warming by enhancing the cooling effect of heat uptake; the reduced warming, in turn, feeds back to reduce heat uptake.

Equation (15) expresses the simple idea that equilibration is decreased by a large ratio of γ, deep-ocean–surface climate coupling, to λ/ε, the coupling of the resultant anomalies to space. The degree of equilibration in the multimodel global mean is a little >0.5, indicating that this ratio is near 1, the strength of coupling to space and to the deep ocean are about the same. Mathematically, efficacy and efficiency enter as a product in (15) and Fig. 8 shows the impact of their intermodel variation on the product. The variations in efficacy are responsible for most of the variation in the product, so that, as expected from the correlations in Table 4, it has a larger influence on the degree of equilibration.

The implication of our simple model interpretation is that one would be more effective in reducing AOGCM uncertainties in transient climate sensitivity by reducing uncertainty in the radiative response to ocean heat uptake than in the relationship of the uptake magnitude to the surface climate perturbation. Uncertainty in radiative feedbacks substantially impacts not only the simulated equilibrium response but also the trajectory toward equilibrium for which ocean processes might have been thought dominant.

6. Conclusions

We argue that simple energy balance model fits to AOGCMs should make use of the concept of the efficacy of ocean heat uptake. This is equivalent to, but we

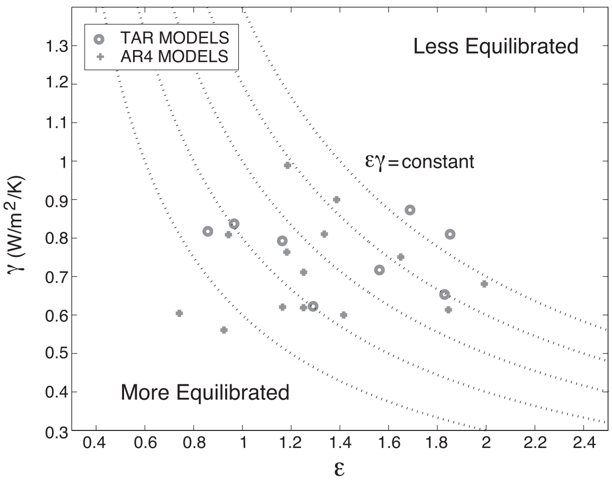

FIG. 8. Scatterplot of the ocean heat flux efficacy against efficiency for 22 climate models used in the IPCC third and fourth assessment reports. The product of the two scattered quantities reduces the equilibration of the surface climate TCR/T_{EQ}.

believe more physically intuitive than, the concept of "effective forcing" since the adjustments that establish efficacy or effective forcing take place on a decadal scale, favoring interpretation as a response rather than a forcing. We also show that efficacy is more parsimonious than "effective sensitivity" since a considerable part of the time dependence of effective sensitivity can be captured with a time-invariant efficacy. The efficacy factor is variable across the AOGCMs used for IPCC assessments but is generally larger than 1 with an average value between 1.3 and 1.4, and can approach 2. Thus for most models the simulated warming is more sensitive to ocean heat uptake than to CO_2 radiative forcing. Amongst the models, the transient climate response is better correlated with the efficacy than it is with the equilibrium climate sensitivity. The efficacy and climate sensitivity have little correlation, indicating that they represent different model characteristics. An understanding of the reasons for the differences in efficacy amongst the models should be useful for resolving the differences in the magnitude of transient climate change simulated in these models.

The use of an efficacy, or its equivalent, is necessary to fit the global mean temperature in both the forcing-increasing and forcing-stabilized sections of a 1% yr^{-1} CO_2 increase experiment with the GFDL CM2.1. The potential significance of high efficacy in slowing the warming is well illustrated by this model and by an analysis of models utilized in the third and fourth IPCC assessments. The stabilized forcing warming commitment inherent in a given level of ocean heat uptake is magnified by the efficacy. High efficacy implies a greater fraction of the equilibrium response will occur after stabilization. Therefore uncertainty about efficacy poses a difficulty for determination of the equilibrium climate sensitivity from observations of forcing, temperature, and ocean heat uptake.

Plattner et al. (2008) and Solomon et al. (2009) have presented the long-term response to CO_2 emissions in intermediate complexity models. In these experiments, there is a near cancellation between the warming effect of reduced ocean heat uptake and the cooling effect of reduced radiative forcing as carbon enters the ocean in the millennium following a cessation of carbon emissions, leading to a global temperature that declines only slightly. Our study indicates that radiative feedbacks play an important role in the impact of the ocean heat uptake reductions and that different AOGCMs may give differing results because of differences in the efficacy of heat uptake. A larger heat uptake efficacy would imply a more durable temperature response to CO_2 emissions as reduction in radiative forcing accompanying oceanic CO_2 uptake experiences a relatively larger warming offset from reduced ocean heat uptake. Our results suggest that the AOGCMs, which contain the most comprehensive simulations of radiative feedbacks and efficacy, should be applied to this long-term emissions commitment problem.

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