Consistent Differences in Climate Feedbacks between Atmosphere–Ocean GCMs and Atmospheric GCMs with Slab–Ocean Models*

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(Manuscript received 30 July 2012, in final form 12 December 2012)

ABSTRACT

Climate sensitivity is generally studied using two types of models. Atmosphere–ocean general circulation models (AOGCMs) include interactive ocean dynamics and detailed heat uptake. Atmospheric GCMs (AGCMs) with slab ocean models (SOMs) cannot fully simulate the ocean's response to and influence on climate. However, AGCMs are computationally cheaper and thus are often used to quantify and understand climate feedbacks and sensitivity. Here, physical climate feedbacks are compared between AOGCMs and SOM-AGCMs from the Coupled Model Intercomparison Project phase 3 (CMIP3) using the radiative kernel technique. Both the global-average (positive) water vapor and (negative) lapse-rate feedbacks are consistently stronger in AOGCMs. Water vapor feedback differences result from an essentially constant relative humidity and peak in the tropics, where temperature changes are larger for AOGCMs. Differences in lapse-rate feedbacks extend to midlatitudes and correspond to a larger ratio of tropical- to global-average temperature changes. Global-average surface albedo feedbacks are similar between models types because of a near cancellation of Arctic and Antarctic differences. In AOGCMs, the northern high latitudes warm faster than the southern latitudes, resulting in interhemispheric differences in albedo, water vapor, and lapse-rate feedbacks lacking in the SOM-AGCMs. Meridional heat transport changes also depend on the model type, although there is a large intermodel spread. However, there are no consistent global or zonal differences in cloud feedbacks. Effects of the forcing scenario [Special Report on Emissions Scenarios A1B (SRESa1b) or the 1% CO₂ increase per year to doubling (1%to2x) experiments] on feedbacks are model dependent and generally of lesser importance than the model type. Care should be taken when using SOM-AGCMs to understand AOGCM feedback behavior.

1. Introduction

The climate's sensitivity to changes in atmospheric constituents (such as CO₂) is a fundamental quantity for predicting climate change. However, state-of-the-science general circulation models (GCMs) show a large range of sensitivities. The Intergovernmental Panel on Climate Change (IPCC) Fourth Assessment Report (AR4) gives an equilibrium global-average surface air temperature sensitivity of $3.26^{\circ} \pm 0.69^{\circ}\text{C}$ for doubled CO₂ in

atmospheric GCMs (AGCMs) coupled to slab (or mixed layer) ocean models (SOMs) (Meehl et al. 2007). Much of the uncertainty in climate projections is due to different radiative feedbacks in GCMs (NRC 2003). These feedbacks amplify or damp climate changes and thus influence the magnitude of the climate's response, or sensitivity, to an imposed forcing.

Climate sensitivity is generally studied using two types of models. Atmosphere–ocean general circulation models (AOGCMs) include interactive ocean dynamics and detailed heat uptake, but they are computationally expensive and require thousands of simulation years to reach a new statistical steady state in response to a forcing. SOM-AGCM experiments require less computer time per model year and reach equilibrium within decades; however, they cannot fully simulate the ocean's response to and influence on atmospheric changes. Improvements in computational resources and scientific computing, as well as a focus on transient climate behavior, have

* Supplemental information related to this paper is available at the Journals Online website: <http://dx.doi.org/10.1175/JCLI-D-12-00519.s1>.

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reduced the need for SOM-based studies, and the new Coupled Model Intercomparison Project phase 5 (CMIP5) experiments are more focused on fully coupled models. However, SOM-AGCMs are still used in cases where a quick exploration of climate feedbacks is needed (e.g., Gettelman et al. 2012; Bitz et al. 2012). Additionally, an understanding of differences between the AOGCM and SOM-AGCM responses is required to compare current AOGCM behavior with previous SOM-AGCM studies.

The “equilibrium climate sensitivity” is the surface air temperature change required to reach a new equilibrium in response to a constant imposed climate forcing. Because fully coupled state-of-the-art AOGCMs are computationally expensive, the equilibrium climate sensitivity is generally determined using atmospheric GCMs with mixed-layer oceans. The assumption is often made that feedbacks in SOM-AGCMs behave similarly to the feedbacks in AOGCMs. However, oceanic processes affect the climate response, so we cannot assume that these feedbacks are the same. For example, zonal mean atmospheric energy transport differs between the two types of models (Held and Soden 2006), and changes in poleward energy transport are linked to climate feedbacks (Hwang and Frierson 2010; Zelinka and Hartmann 2012). It is important to determine the extent to which SOM-AGCMs can simulate AOGCM feedbacks because biases would limit the applicability of the cheaper SOM-AGCM as a tool to study climate sensitivity. Additionally, a comparison of the behaviors of the two model types provides insight into the effects of ocean dynamics on climate sensitivity.

Few equilibrium AOGCM experiments are available, but those that are suggest that AOGCMs and SOM-AGCMs may have different equilibrium sensitivities and feedbacks. Watterson (2003) obtains a smaller climate sensitivity when comparing the Commonwealth Scientific and Industrial Research Organization (CSIRO) Mark version 2 AOGCM with its SOM version, related to reduced poleward heat transport (see Table A1 for an expansion of all model names). On the other hand, Stouffer and Manabe (1999) find a lower climate sensitivity (3.7 versus 4.5 K) for a SOM-AGCM, although it is partially attributed to a too-cool starting temperature. Although Danabasoglu and Gent (2009) find a difference in climate sensitivity of only 0.15 K for the low-resolution version of the NCAR CCSM3, the magnitudes of the individual feedbacks differ between the two model versions (Jonko et al. 2013). In particular, for a doubling of CO₂, the lapse rate and water vapor feedbacks differ, while, for further increases in CO₂, the albedo and cloud feedbacks also differ. These results suggest

that the ability of a SOM-AGCM to replicate the equilibrium climate sensitivity of an AOGCM is model dependent.

To circumvent the long integrations required to calculate the equilibrium climate sensitivity using coupled GCMs, the “effective climate sensitivity” (Murphy 1995) is often used (e.g., Gregory and Mitchell 1997; Senior and Mitchell 2000; Raper et al. 2002). The effective climate sensitivity is an estimate of the equilibrium climate sensitivity as determined from the temperature change at the time of CO₂ doubling (or other point in a transient run) and the fraction of the imposed radiative forcing that the climate has already adjusted to (i.e., the imposed radiative forcing minus the instantaneous heat uptake by the climate system). The use of an effective climate sensitivity assumes that feedback strengths remain constant as the climate changes. However, climate feedbacks may change with time due to a dependence on the climate state. Watterson (2000) finds that the effective climate sensitivity in the CSIRO GCM is approximately constant with time (and in some cases lower than the corresponding AGCM equilibrium response). On the other hand, Raper et al. (2001) show that the effective climate sensitivity in HadCM2 increases from 2.0 to 3.8 K over 900 yr. (The corresponding mixed-layer AGCM equilibrium sensitivity is 4.1 K.) Senior and Mitchell (2000) propose that this time dependence arises from changing cloud feedbacks in the Southern Hemisphere. Using a 1000-yr Canadian Centre for Climate Modeling and Analysis (CCCma) simulation, Boer and Yu (2003a) obtain decreased sensitivity (10%–20%) with time. Thus, the results are highly model dependent, as confirmed by Forster and Taylor (2006) for 220-yr integrations of IPCC AR4 models. Nonetheless, the lack of availability of many equilibrium AOGCM simulations necessitates the use of transient AOGCM simulations. In fact, Andrews et al. (2012) highlight that a focus on equilibrium climate sensitivity and feedbacks is less useful than understanding the transient behavior when models display time-dependent feedbacks. In this study, we compare feedbacks between the two model types (SOM-AGCMs and AOGCMs), keeping in mind the potential of time-dependent feedbacks for the AOGCMs.

In addition to Watterson (2000) and Raper et al. (2001), some studies have found differences in sensitivities and feedbacks from transient AOGCM simulations compared to those from equilibrium SOM-AGCM experiments. Using the CCCma, Boer and Yu (2003c) attribute differences in feedbacks, in particular in high latitudes and the tropical Pacific, to the lack of dynamical processes in SOM-AGCMs. Forster and Taylor (2006)

compare IPCC AOGCM climate feedbacks determined using the regression technique of Gregory et al. (2004) with equilibrium SOM-AGCM feedbacks and find an approximate correspondence, but the values differ by up to 25%. Williams et al. (2008) find that AOGCMs can only estimate the equilibrium SOM-AGCM climate sensitivity within 0.5 K, even taking into account an effective forcing that incorporates fast processes (see their Fig. 2). Yokohata et al. (2008) use the approximate partial radiative perturbation technique (APRP; Taylor et al. 2007) for the shortwave (SW) variables and the cloud radiative forcing technique (Δ CRF; Cess and Potter 1988) for the longwave (LW) feedbacks to compare feedbacks in eight SOM-AGCM–AOGCM pairs of the World Climate Research Programme’s (WCRP’s) Coupled Model Intercomparison Project phase 3 (CMIP3) multimodel dataset. They find that the equilibrium SOM-AGCM sensitivity to a doubling of CO_2 differs from the corresponding effective sensitivity by -1.3 to 1.6 K. They largely attribute this to ice–albedo and SW–cloud feedbacks. Hwang and Frierson (2010) also use the APRP and the LW Δ CRF methodology and find differences in CMIP3 albedo and cloud feedbacks, as well as corresponding meridional heat transport changes.

The current work compares transient AOGCM feedbacks to those from equilibrium AGCM–SOMs using the CMIP3 multimodel dataset. Most of the previous work has focused on combined (i.e., cloud radiative forcing or clear sky) feedbacks or lacks detail for the LW feedbacks. For example, the APRP technique only works for SW feedbacks, and the Δ CRF technique and regression technique of Gregory et al. (2004) cannot resolve the separate effects of individual constituents (water vapor, lapse rate, surface albedo, and clouds). Here we use the radiative kernel technique (Soden and Held 2006; Soden et al. 2008; Shell et al. 2008) to separate feedbacks into individual constituents. In particular, we focus on the LW noncloud feedbacks (water vapor, lapse rate) and find meridional structures not resolved by these other techniques. Additionally, we expand the analysis to all available models and ensemble members, as well as an additional scenario, to study feedback behavior in as many simulations as possible. The goal is to separate model-independent behavior, which provides information about general structural differences between AOGCMs compared with SOM-AGCMs, from model-dependent differences, which require a closer examination of specific model parameterizations. The kernel technique is particularly useful for this objective, since it allows easy comparison of effective and equilibrium feedbacks.

2. Methodology

The radiative feedback parameter (λ) is defined as the change in net (absorbed minus emitted) top-of-the-atmosphere (TOA) radiation (ΔR) caused by the climate response to an imposed forcing, divided by the global average surface air temperature change ($\Delta \bar{T}_{\text{as}}$). A negative feedback parameter indicates that the feedback stabilizes the climate (decreases R for a temperature increase). The climate sensitivity is inversely proportional to λ .

Following Zhang et al. (1994), we separate the feedback parameter into terms for the different climate components:

$$\lambda = \frac{\Delta R}{\Delta \bar{T}_{\text{as}}} = \lambda_{T_a} + \lambda_{T_s} + \lambda_Q + \lambda_\alpha + \lambda_C + \epsilon. \quad (1)$$

The feedback parameter is thus the sum of the feedback parameters related to surface temperature T_s , atmospheric temperature T_a , water vapor Q , surface albedo α , and clouds C , plus a residual ϵ , which is generally small (about 10%; Shell et al. 2008; Jonko et al. 2012). The normalization by $\Delta \bar{T}_{\text{as}}$ allows comparison of feedbacks between transient (AOGCM) experiments and equilibrium (SOM-AGCM) experiments. Individual feedback parameters (e.g., water vapor) for the two types of models should be equal if the individual feedback strengths are the same, even if other feedbacks vary or the model is not in equilibrium.

The radiative kernel technique (Soden and Held 2006; Soden et al. 2008; Shell et al. 2008) decomposes each feedback parameter λ_X into two parts. The “radiative kernel” ($\partial R / \partial X$) describes the change in TOA fluxes for a standard change in property X (temperature, water vapor, or surface albedo) for every grid point and level (for the multilevel variables), for each month of the year. The kernel is calculated using an offline radiative transfer model and depends on the radiative properties and base state of the model. The second component is the monthly average climate response of the feedback variable to the $\bar{T}_{\text{as}}$ change ($dX/d\bar{T}_{\text{as}}$, calculated as $\Delta X / \Delta \bar{T}_{\text{as}}$)—for example, the difference in 20-yr X averages between the doubled CO_2 and present-day GCM simulations. For the case of water vapor, we use the natural log of the specific humidity (SH) as the variable X , since changes in radiation scale more like the natural log. The feedback parameter λ_X is the product of the kernel and the climate response. For the three-dimensional variables, we sum the results up to the tropopause, defined as 100 hPa at the equator and linearly decreasing to 300 hPa at the poles (Soden et al. 2008).

The feedbacks parameters corresponding to water vapor and temperature can be combined in multiple ways (Held and Shell 2012):

$$\lambda_{T_a} + \lambda_{T_s} + \lambda_Q = \lambda_T + \lambda_L + \lambda_Q = \tilde{\lambda}_T + \tilde{\lambda}_L + \tilde{\lambda}_H. \quad (2)$$

The Planck feedback (λ_T) is calculated by assuming the surface air temperature change for each grid point extends from the surface through the troposphere, while the lapse rate feedback parameter (λ_L) is the product of the temperature kernels and the departure of ΔT_a and ΔT_s from ΔT_{as} at each grid point and tropospheric level. The quantities λ_{T_a} , λ_{T_s} , λ_T , and λ_L are all calculated assuming a constant specific humidity. If, instead, we calculate the Planck ($\tilde{\lambda}_T$) and lapse rate ($\tilde{\lambda}_L$) feedback parameters by changing the specific humidity to maintain a constant relative humidity while the temperature changes, we are then left with a feedback parameter ($\tilde{\lambda}_H$) describing the effects of changes in relative humidity (RH). This alternative feedback perspective using relative humidity as the state variable, while not affecting the overall climate sensitivity, simplifies analysis by greatly reducing the cancellation between large positive water vapor and large negative lapse rate feedbacks (Held and Shell 2012). We will use these alternate separations to explore various aspects of the climate response.

Differences in cloud feedbacks account for the largest portion of the spread in GCM sensitivities (Randall et al. 2007). Because cloud feedbacks are nonlinear, we cannot simply apply the kernel technique to, for example, the cloud fraction. Instead, we used the adjusted cloud radiative forcing (CRF) technique (Shell et al. 2008). CRF is the difference between the all-sky TOA flux and the clear-sky TOA flux, the TOA flux if clouds were removed but all other variables remained the same. The difference in CRF between two model simulations (ΔCRF) is often used as a measure of the cloud feedback (e.g., Cess and Potter 1988). However, contamination of ΔCRF by contributions from noncloud terms introduces errors (Zhang et al. 1994; Colman 2003; Soden et al. 2004), so we “correct” ΔCRF using kernel-derived contributions to ΔCRF from noncloud variables as well as the CO_2 forcing. Note that this methodology does not distinguish between the cloud response to $\bar{T}_{as}$ and the direct response of clouds to the atmospheric forcing (Gregory and Webb 2008).

We use the radiative kernels based on the Community Atmosphere Model, version 3 (CAM3; Shell et al. 2008). Differences in kernels based on other models contribute less to the spread of estimated feedbacks in IPCC AR4 models than differences in feedback variables (Soden

et al. 2008). Additionally, since AOGCMs and their corresponding AGCMs have the same radiative transfer code, the use of a single kernel should not contribute significantly to errors in feedback differences between the two model types. This technique does, however, assume that changes in TOA fluxes are linear with respect to climate variable anomalies. We partially quantify the error resulting from deviations from this linearity using the clear sky test (Shell et al. 2008) (see Fig. S1 in the supplemental material).

The 12 WCRP CMIP3 models have the needed monthly average output for both the AOGCM and the SOM-AGCM simulations (see Table S1 of the supplemental material). For the Special Report on Emissions Scenarios (SRES) A1B (720 ppm CO_2 stabilization; SRESa1b) experiments, we use the differences between 2080–99 in the SRES A1B simulations and 1980–99 in the twentieth century (20c3m) simulations (i.e., we subtracted the 20c3m values from the SRESa1b values). For the “1% CO_2 increase per year to doubling” (1% to2x) experiments, we compared years 60–80 (50–70 for shorter simulations) of with the first 20 years of each simulation. For the slab models, we took the difference between the last 20 years of the doubled CO_2 (2x CO_2) and control (slabcntl) runs. The use of two AOGCM scenarios allows us to examine the dependence of our conclusions on the “type” of forcing and thus explore the robustness of our conclusions.

3. Global-average feedback differences

Some multimodel mean global- and annual-average feedbacks exhibit differences between the different types of models, summarized in Table 1. The largest feedback is the fixed-SH Planck λ_T (or fixed-RH Planck $\tilde{\lambda}_T$) feedback, but these have lower spreads and standard deviations than the other feedbacks. The shortwave cloud feedback (λ_{CSW}), while not having the largest mean magnitudes, does have the largest standard deviations, in agreement with previous analyses of the CMIP3 models (Randall et al. 2007). The mean values for the AOGCM (SRESa1b and 1%to2x) experiments are in better agreement with each other than with the SOM experiment mean feedback for some feedbacks (fixed-SH Planck λ_T and lapse rate λ_L , specific humidity λ_Q , fixed-RH lapse rate $\tilde{\lambda}_L$, and all noncloud feedbacks). The differences between AOGCMs and SOM-AGCMs for these particular feedbacks are approximately the same as the standard deviations for an individual experiment, but the spreads (maximum minus minimum values) indicate a substantial overlap between the AOGCM and SOM-AGCM feedbacks. The fixed-SH lapse rate feedback (λ_L) has the largest mean difference

TABLE 1. Mean, intermodel spread (difference between maximum and minimum values), and standard deviations of global- and annual-average feedbacks ($\text{W m}^{-2} \text{K}^{-1}$) for the two AOGCM experiments [SRESa1b (A1B) and 1%to2x (1%)] and the 2xCO₂ SOM-AGCM experiment. Shown are the Planck, lapse rate, and specific humidity feedbacks; the fixed-RH Planck and lapse rate feedbacks; the relative humidity feedback; the albedo feedback; the shortwave and longwave cloud feedbacks; the sum of all the feedbacks except cloud (Non-C = $\lambda_T + \lambda_L + \lambda_Q + \alpha$); the sum of all the feedbacks (All); and the global average surface air temperature change (K). The values are determined using the ensemble means for each model.

		λ_T	λ_L	λ_Q	$\tilde{\lambda}_T$	$\tilde{\lambda}_L$	$\tilde{\lambda}_H$	λ_α	λ_{CSW}	λ_{CLW}	Non-C	All	$\Delta\bar{T}_{\text{as}}$
Mean	A1B	-3.11	-0.77	1.90	-1.75	-0.20	-0.03	0.26	0.26	0.38	-1.71	-1.06	2.77
Mean	1%	-3.12	-0.78	1.95	-1.75	-0.20	-0.01	0.24	0.34	0.44	-1.72	-0.94	1.62
Mean	2xCO ₂	-3.07	-0.57	1.75	-1.75	-0.13	-0.02	0.27	0.32	0.36	-1.62	-0.94	3.29
Spread	A1B	0.13	0.63	0.47	0.02	0.28	0.32	0.23	1.27	0.68	0.59	1.03	2.15
Spread	1%	0.10	0.44	0.56	0.02	0.17	0.29	0.14	1.44	0.72	0.35	0.98	1.22
Spread	2xCO ₂	0.08	0.50	0.44	0.02	0.20	0.33	0.16	1.14	0.59	0.47	0.80	2.60
Std dev	A1B	0.04	0.19	0.17	0.01	0.08	0.10	0.06	0.44	0.20	0.17	0.35	0.61
Std dev	1%	0.03	0.16	0.19	0.01	0.06	0.11	0.05	0.49	0.22	0.12	0.40	0.39
Std dev	2xCO ₂	0.03	0.14	0.15	0.01	0.06	0.09	0.05	0.34	0.16	0.13	0.28	0.72

between the AOGCM and SOM-AGCM feedbacks, followed closely by the specific humidity feedback (λ_Q); other differences are less than half the magnitudes of these two. Other feedbacks show either little variation in mean values among the three experiments (fixed-RH Planck $\tilde{\lambda}_T$, relative humidity $\tilde{\lambda}_H$, and surface albedo λ_α) or have mean differences that are much smaller than the standard deviations and the spread (SW and LW cloud feedbacks, λ_{CSW} and λ_{CLW}). Note that the AOGCM simulations are not yet in equilibrium. Therefore, the global average surface air temperature changes $\Delta\bar{T}_{\text{as}}$ differ, although some portion of these differences could represent actual differences in equilibrium climate sensitivity.

Figures 1 and 2 compare the global- and annual-average feedbacks of the individual SOM-AGCMs with those of their corresponding SRESa1b (circles) or 1%to2x (stars) AOGCMs. Each AOGCM ensemble member is plotted separately, so that the range of circles or stars of a given color can be used to roughly estimate the spread in that model's climate response. The diagonal lines show the one-to-one values. If feedbacks were the same in SOM-AGCMs and AOGCMs, all of the points would lie along the lines. To the extent that points lie off the lines, the ocean model configuration (including the sea ice behavior) is influencing the feedbacks. Since the kernel technique holds $\partial R/\partial X$ constant, this implies that $\Delta X/\Delta\bar{T}_{\text{as}}$ is different. The climate response could be dependent on the control state, since SOM-AGCMs generally do not replicate the sea surface temperature and sea ice patterns of their corresponding AOGCMs (Bitz et al. 2012). A dependence on the surface state would indicate that, to some extent, transient feedbacks diverge from the equilibrium feedbacks (i.e., the effective sensitivity changes with time). Alternately (or additionally), $\Delta X/\Delta\bar{T}_{\text{as}}$ may vary because

differences in ocean dynamics and heat transport alter the climate response. In this case, the equilibrium feedbacks may differ between the two model types. Table 2 lists the ensemble mean differences between the SRESa1b or 1%to2x and the SOM-AGCM experiments for each model. (Feedback parameters for each ensemble member of each experiment are provided in Table S2 of the supplemental material, and Table S3 gives the differences in feedbacks between the AOGCM-AGCM and SOM-AGCM ensemble members.)

We start by examining the standard fixed-specific-humidity reference state framework [i.e., the common separation of feedback parameters into the Planck (λ_T), lapse rate (λ_L), and specific humidity (λ_Q) feedbacks in Eq. (2)]. While the traditional Planck feedback (Fig. 1a) shows little variation among models and between AOGCMs and SOM-AGCMs, the (negative) lapse rate feedback (Fig. 1b) is, with the exception of a single experiment, larger (more negative) in the AOGCM experiments, by up to $0.46 \text{ W m}^{-2} \text{K}^{-1}$. For about half of the experiments, the magnitude of the lapse rate feedback difference between the two model types is larger than the other feedback difference magnitudes (Table 2). Note, however, that it is not always the same sign as the sum of all the feedback parameters (All), implying cancellation by some other feedback (e.g., water vapor). The (positive) water vapor feedback (Fig. 1c) is consistently larger, by up to $0.39 \text{ W m}^{-2} \text{K}^{-1}$, in the AOGCMs than in the SOM-AGCMs. Multimodel mean AOGCM water vapor feedbacks are larger (more positive) by $0.15 \pm 0.08 \text{ W m}^{-2} \text{K}^{-1}$ for SRESa1b and $0.21 \pm 0.12 \text{ W m}^{-2} \text{K}^{-1}$ for 1%to2x experiments (where the uncertainty indicates one standard deviation), and the lapse rate feedbacks more negative by 0.19 ± 0.13 (SRESa1b) and 0.22 ± 0.12 (1%to2x) $\text{W m}^{-2} \text{K}^{-1}$ (Table 2). The water vapor feedback is dominated by

FIG. 1. Comparison of global- and annual-average feedbacks ($\text{W m}^{-2} \text{K}^{-1}$) from AOGCM and SOM-AGCM experiments. Circles correspond to SRESa1b AOGCM experiments; stars indicate 1%to2x experiments. When multiple ensemble members are available for an AOGCM experiment, additional circles or stars are plotted. The diagonal line indicates a one-to-one relationship. While individual plots are centered on the average value for that variable, the distance between major tick marks (0.3) is constant.

the longwave component (see Figs. S2a,b in the supplemental material). The sums of the water vapor and lapse rate feedbacks show better agreement (Fig. 1g), since anomalies in these feedbacks tend to compensate for a given simulation (Colman 2004; Soden and Held 2006). While this compensation results in a smaller

combined water vapor and lapse rate feedback effect from a net TOA energy budget framework (a mean of $-0.04 \text{ W m}^{-2} \text{K}^{-1}$ for SRESa1b and $-0.01 \text{ W m}^{-2} \text{K}^{-1}$ for 1%to2x experiments), the consistency of the signs of the differences for each feedback is indicative of fundamental processes. Additionally, as will be shown

FIG. 2. As in Fig. 1 with the SW sum including the SW water vapor, surface albedo, and cloud feedbacks. The LW sum includes the longwave fixed-SH Planck, lapse rate, and water vapor plus cloud feedbacks. All feedback is the sum of the SW and LW feedbacks and determines climate sensitivity.

later, these differences have meridional structures and do not cancel on a zonal-average basis. An understanding of these common behavioral differences between the two model types will aid in our understanding of the overall climate system.

To further explore these consistent water vapor and lapse rate feedback behaviors, we calculate the

fixed-relative-humidity reference state feedbacks (Figs. 1d–f). For all three feedbacks, the fixed-RH versions have a smaller intermodel spread than the fixed-SH Planck feedbacks (Figs. 1a–c; Table 1), demonstrating the utility of the fixed-RH feedbacks. Additionally, the differences between AOGCM and SOM-AGCM feedbacks are smaller. Of the three fixed-RH feedbacks, the lapse rate

TABLE 2. Ensemble mean differences in global- and annual-average feedbacks ($\text{W m}^{-2} \text{K}^{-1}$) between AOGCMs and their corresponding SOM-AGCMs for the SRESa1b (A1B) and 1%to2x (1%) experiments. Shown are the Planck, lapse rate, and specific humidity feedbacks; the fixed-RH Planck and lapse rate feedbacks; the relative humidity feedback; the albedo feedback; the shortwave and longwave cloud feedbacks; the sum of all the feedbacks except cloud (Non-C = $\lambda_T + \lambda_L + \lambda_Q + \alpha$); and the sum of all the feedbacks (All). The multimodel means and standard deviations are determined using the ensemble means for each model.

Model	Scenario	λ_T	λ_L	λ_Q	$\tilde{\lambda}_T$	$\tilde{\lambda}_L$	$\tilde{\lambda}_H$	λ_α	λ_{CSW}	λ_{CLW}	Non-C	All
CCCma CGCM3.1	A1B	-0.09	-0.42	0.35	0.00	-0.17	0.00	0.00	0.12	0.09	-0.17	0.04
CCCma CGCM3.1	1%	-0.08	-0.38	0.39	0.01	-0.15	0.07	-0.05	0.18	0.17	-0.12	0.23
CCCma CGCM3.1 (T63)	A1B	-0.02	-0.07	0.09	0.00	-0.02	0.02	0.11	0.02	0.03	0.12	0.16
CSIRO Mk3.0	A1B	-0.04	-0.28	0.18	0.01	-0.11	-0.05	-0.02	-0.05	-0.10	-0.16	-0.31
CSIRO Mk3.0	1%	-0.06	-0.38	0.28	0.01	-0.15	-0.03	-0.07	0.01	0.08	-0.24	-0.14
GFDL CM2.0	A1B	-0.04	-0.20	0.20	0.00	-0.08	0.03	0.02	-0.34	-0.07	-0.02	-0.43
GFDL CM2.0	1%	-0.07	-0.29	0.28	0.00	-0.11	0.03	-0.08	-0.01	-0.18	-0.16	-0.36
GISS-ER	A1B	-0.07	-0.29	0.15	-0.00	-0.10	-0.09	-0.04	-0.19	0.26	-0.25	-0.17
INM-CM3.0	A1B	0.01	0.09	0.01	0.00	0.04	0.07	0.08	-0.03	0.00	0.19	0.16
INM-CM3.0	1%	-0.03	-0.03	0.19	0.00	-0.01	0.13	0.05	-0.10	0.25	0.17	0.32
MIROC 3.2(hires)	A1B	-0.04	-0.23	0.16	0.01	-0.10	-0.01	-0.00	0.02	0.03	-0.10	-0.05
MIROC 3.2(hires)	1%	-0.03	-0.13	0.11	0.01	-0.06	0.00	0.02	0.20	0.01	-0.03	0.18
MIROC 3.2(medres)	A1B	-0.07	-0.28	0.22	0.00	-0.12	-0.02	-0.08	0.06	0.01	-0.21	-0.14
MIROC 3.2(medres)	1%	-0.11	-0.30	0.32	0.01	-0.14	0.04	-0.11	0.21	0.24	-0.21	0.24
MPI ECHAM5	A1B	-0.02	-0.13	0.13	0.01	-0.03	0.01	0.05	0.10	0.00	0.03	0.12
MPI ECHAM5	1%	-0.03	-0.15	0.16	0.01	-0.04	0.01	0.03	0.07	0.09	0.01	0.17
MRI CGCM2.3.2a	A1B	-0.03	-0.16	0.09	-0.01	-0.07	-0.03	-0.08	-0.16	0.07	-0.19	-0.27
MRI CGCM2.3.2a	1%	-0.03	-0.19	0.10	-0.00	-0.08	-0.04	-0.09	0.06		-0.21	
NCAR CCSM3.0	A1B	-0.03	-0.09	0.08	-0.01	-0.01	-0.01	-0.02	-0.33	-0.10	-0.05	-0.48
NCAR CCSM3.0	1%	-0.02	-0.13	0.03	-0.01	-0.03	-0.08	-0.02	-0.21	0.03	-0.13	-0.31
UKMO HadGEM1	A1B	-0.02	-0.26	0.18	0.01	-0.08	-0.02	-0.08	0.07	-0.03	-0.18	-0.14
Multimodel mean	A1B	-0.04	-0.19	0.15	0.00	-0.07	-0.01	-0.01	-0.06	0.01	-0.08	-0.13
Std dev	A1B	0.03	0.13	0.08	0.01	0.06	0.04	0.06	0.16	0.10	0.14	0.22
Multimodel mean	1%	-0.05	-0.22	0.21	0.00	-0.08	0.01	-0.04	0.05	0.08	-0.10	0.04
Std dev	1%	0.03	0.12	0.12	0.01	0.05	0.06	0.06	0.14	0.14	0.13	0.27

feedback ($\tilde{\lambda}_L$) shows the largest differences, with the AOGCM feedbacks larger in magnitude (more negative) than the SOM-AGCM feedbacks (Table 2), except for one model [Institute of Numerical Mathematics Coupled Model, version 3.0 (INM-CM3.0)]. However, the differences ($-0.07 \pm 0.06 \text{ W m}^{-2} \text{K}^{-1}$ for SRESa1b and $-0.08 \pm 0.05 \text{ W m}^{-2} \text{K}^{-1}$ for 1%to2X) are barely larger than one standard deviation, whereas the fixed-SH lapse rate feedback differences are larger and more significant (Table 2). The relative humidity feedback ($\tilde{\lambda}_H$) shows no consistent dependence on model type. For some models, the relative humidity feedback is more positive in the AOGCM experiments, while others have more positive SOM-AGCM feedbacks. However, the feedback differences are all less than $0.1 \text{ W m}^{-2} \text{K}^{-1}$, and the average effect is essentially zero. Thus, in terms of global effect, the relative humidity response is similar for both AOGCMs and SOM-AGCMs. The fixed-RH Planck feedback ($\tilde{\lambda}_T$) differs by no more than $0.01 \text{ W m}^{-2} \text{K}^{-1}$ between the two model types. Since the intermodel standard deviation is small (Fig. 1d; Table 1), the fixed-RH Planck feedback is basically the same regardless of the model or scenario used.

Comparison of Figs. 1c and 1f indicates that the larger specific humidity feedbacks in AOGCMs are caused mainly by water vapor increases at fixed relative humidity (since $\tilde{\lambda}_H$ differences are smaller), implying that they correspond to larger increases in temperature. The radiative kernel feedback calculation normalizes by the global average surface air temperature, so the larger temperature increases must be localized vertically (i.e., related to lapse rate changes) and/or horizontally. In the next section, we explore the meridional aspect. Figure 1b shows that, overall, the lapse rate decreases by a larger amount in the coupled experiments. The simultaneous larger water vapor increases necessary to maintain a constant relative humidity partially compensate for the TOA energy budget effects of these lapse rate changes, but, as described above, the fixed-RH lapse rate feedback ($\tilde{\lambda}_L$; Fig. 1e) is somewhat larger in AOGCMs compared with SOM-AGCMs. (Remember that the fixed-RH lapse rate feedback parameter is the sum of λ_L and the portion of λ_Q that corresponds to a constant relative humidity for the lapse rate change.) This tendency for a more negative $\tilde{\lambda}_L$ is reflected in the total temperature plus water vapor feedback parameter ($\lambda_T + \lambda_Q + \lambda_L = \tilde{\lambda}_T + \tilde{\lambda}_L + \tilde{\lambda}_H$; Fig. 1h). With the

exception of INM-CM3.0, this sum is more negative for the AOGCMs, which, all other feedbacks being equal, would result in a smaller climate sensitivity (since climate sensitivity is proportional to $-1/\lambda$).

The global-average albedo feedbacks (Fig. 2a) are similar for AOGCMs and SOM-GCMs. There are discrepancies for some models, likely due to different sea ice components in SOM-AGCMs versus AOGCMs, but the signs of the discrepancies are both positive and negative, depending on the model. These results are similar to those of Yokohata et al. (2008), who find both positive and negative differences using the APRP technique, which is better able to separate feedbacks in cases where the radiation behaves nonlinearly. Our differences range from -0.12 to $0.11 \text{ W m}^{-2} \text{ K}^{-1}$ for individual ensemble members. For some models—for example, Model for Interdisciplinary Research on Climate 3.2, medium-resolution version [MIROC3.2(medres)], and Meteorological Research Institute Coupled General Circulation Model, version 2.3.2a (MRI CGCM2.3.2a)—both AOGCM experiments have smaller albedo feedbacks than the SOM-AGCM experiment, by at least $0.05 \text{ W m}^{-2} \text{ K}^{-1}$, the standard deviation of the intra-experiment albedo feedback. On the other hand, INM-CM3.0 has larger albedo feedbacks in the AOGCM experiments. Other models show opposite signed responses for the two AOGCM scenarios.

Combining all the noncloud feedbacks (Fig. 2b; Table 2), CCCma CGCM 3.1 (T63), INM-CM3.0, and MPI ECHAM5 have a reduced (less negative) noncloud feedback, while the rest of the models show an increased (more negative) noncloud feedback. Again, however, the multimodel mean differences are smaller than the standard deviation, indicating a relatively unimportant bias on climate sensitivity, or, at least, a very model-specific effect.

The shortwave cloud feedback (Fig. 2c) exhibits the largest intermodel range for all three experiments (Table 1). Although the differences between model types are smaller in comparison to this range, especially for the 1%to2x case, the magnitudes of the differences between AOGCMs and SOM-AGCMs can be comparable to those for other feedbacks (up to 0.23 or $-0.35 \text{ W m}^{-2} \text{ K}^{-1}$), and the SW cloud feedback difference is the largest difference for about a quarter of the experiments (Table 2). Furthermore, the multimodel standard deviation is largest for SW cloud feedback differences, indicating a strong influence on the total feedback. Unfortunately, there is no consistent sign of the differences, suggesting the lack of a fundamental underlying physical principle. Furthermore, the lapse rate feedback difference is actually larger in magnitude than the SW cloud feedback difference in about two-thirds

of the experiments, although it is often canceled by an opposing water vapor feedback difference. Note that this result is not in contradiction to Yokohata et al. (2008), who find that SW cloud feedbacks are responsible for much of the difference between AOGCM and SOM-AGCM feedbacks but do not specifically separate out the water vapor and lapse rate feedbacks.

Finally, the LW cloud feedback (Fig. 2d) also shows a mix of positive and negative differences between the two model types (from -0.18 to $0.28 \text{ W m}^{-2} \text{ K}^{-1}$), and the multimodel mean differences (Table 2) are less than one standard deviation and far less than the intermodel spreads within a given experiment (Table 1). Thus, as with the SW cloud feedback, the use of SOM-AGCMs to estimate LW cloud feedbacks does not introduce an obvious bias due to model type. The spread in the net cloud feedback (Fig. 2e) is dominated by the SW cloud feedbacks, since the range of LW feedbacks is smaller.

Figure 2h shows the sum of the individual feedbacks, which is proportional to the negative inverse of the climate sensitivity [Eq. (1)]. Therefore, models with more negative “All” feedbacks are less sensitive. In agreement with Yokohata et al. (2008), some models have larger AOGCM sensitivities, while others have larger sensitivities for the SOM versions. Note, however, that Yokohata et al. (2008) could not explain all of the differences in sensitivity between model types using the albedo and cloud feedback differences. The systemic differences in the water vapor and lapse rate feedbacks identified here could help resolve these climate sensitivity differences. Since the lapse rate and water vapor feedback differences tend to cancel, there is not a consistent sign of the $\lambda_L + \lambda_Q$ difference between the two model configurations, but the sum is an important contribution to the total feedback parameter for some models.

We examine the sensitivity of these consistent global-average water vapor and lapse rate feedback differences to the forcing by comparing the SRESa1b (circles) and 1%to2x (stars) experiments (Figs. 1b,c). The SRESa1b scenario includes changes in CO_2 , other greenhouse gases, and aerosols, while the 1%to2x and SOM-AGCM scenarios have only a doubling of CO_2 . In all cases (except one) both the SRESa1b and 1%to2x water vapor and lapse rate feedbacks are stronger than the SOM-AGCM feedbacks, indicating that these stronger feedbacks are a robust result, independent of the particular forcing scenario used for the AOGCM experiment.

If feedbacks depend on the climate forcing, we might expect the 1%to2x feedbacks to match the SOM-AGCM feedbacks more closely than the SRESa1b feedbacks do, since the 1%to2x and SOM experiments have a

similar forcing magnitude and spatial structure. Interestingly, we find this to be the case for only some of the models. For example, for MIROC3.2(hires), the water vapor and lapse rate feedbacks in the SRESa1b experiment differ more from the SOM-AGCM feedbacks than the 1%to2x feedbacks do, while CSIRO Mark version 3.0 (CSIRO Mk3.0) and Geophysical Fluid Dynamics Laboratory Climate Model version 2.0 (GFDL CM2.0) have larger feedback differences in the 1%to2x experiments (Table 2). Furthermore, for a given model, the SRESa1b and 1%to2x points (circles and stars in Figs. 1b,c) tend to cluster together. These results suggest that the effect of the scenario on the modeled water vapor plus lapse rate feedback is not consistent across models and that the details of the forcing are of lesser importance than model type (i.e., AOGCM or SOM-AGCM).

Figures 2c and 2d suggest that the SW and LW cloud feedbacks tend to be more positive in the 1%to2x compared to the SRESa1b experiments. Seven out of nine models have more positive ensemble mean λ_{CSW} and six out of eight models have more positive λ_{CLW} in the 1%to2x experiments (Table 2). The more positive 1%to2x feedbacks can also be seen in the all LW feedbacks (Fig. 2g), and all eight 1%to2x experiments have more positive total (All) feedbacks (Fig. 2h), corresponding to a higher equilibrium climate sensitivity. (Note that $\Delta\bar{T}_{\text{as}}$ is smaller for the 1%to2x experiments than the SRESa1b experiments; see below.) However, we cannot draw conclusions here for (at least) a few reasons. First, the cloud feedback estimates, based on the adjusted ΔCRF , do not include corrections for masking due to aerosols, resulting in errors in the cloud feedback calculation for the SRESa1b case, while aerosols are unchanged in the 1%to2x and SOM-AGCM simulations. These errors can be seen in the clear-sky flux test (see Fig. S1 in the supplemental material), particularly in the SW. Although the 1%to2x cloud feedbacks (Figs. 2c,d) do not necessarily match the SOM-AGCM feedbacks any better than the SRESa1b feedbacks do, a full calculation of the cloud masking is needed, which requires both the clear-sky and all-sky forcing TOA patterns. Unfortunately, these data are not usually provided in the CMIP3 archive. Another possibility for the different cloud feedback behavior between the 1%to2x and SRESa1b cases may be due to the direct response of clouds to the atmospheric forcing (Gregory and Webb 2008), which is not treated separately by the radiative kernel technique.

Finally, these AOGCM simulations are not in equilibrium; the differences may merely reflect discrepancies in the transient feedbacks (see Fig. S1c in the supplemental material, where the “Model” net fluxes

are greater than zero for the AOGCM experiments). SRESa1b experiments have been run for a longer time (100 versus 60 yr) and have larger $\Delta\bar{T}_{\text{as}}$ (Table 1), which are closer to the SOM-AGCM values [and the circles (SRESa1b) have smaller positive net TOA fluxes than the stars (1%to2x) in Fig. S1c]. Thus, the different cloud feedback behavior between the 1%to2x and SRESa1b cases may be due to a changing cloud feedback with time. Additionally, the differences we observe between the AOGCM and SOM-AGCM feedbacks (e.g., lapse rate and water vapor) may partially reflect a variation of feedback magnitude based on the model state. Our preliminary calculations with longer 1%to2x runs suggest that the feedback parameters change as the model approaches equilibrium, but further study is beyond the scope of this paper. More work is needed to determine whether these feedback differences are due to nonlinearities in feedbacks with temperature (or other climate components) versus a dependence on the type of model (i.e., AOGCM versus SOM-AGCM). To the extent that climate responses are nonlinear (e.g., vary as a function of time/climate state), new frameworks for understanding and quantifying climate feedbacks may be required (Andrews et al. 2012).

4. Zonal-average feedback differences

To better understand the mechanisms of these feedback differences, we examine their meridional structures. Figures 3 and 4 show the zonal-average differences in feedbacks (SRESa1b feedbacks minus their corresponding SOM-AGCM feedbacks). We focus on SRESa1b feedbacks because more simulations are available, but 1%to2x feedbacks (not shown) usually exhibit similar patterns.

The zonal-average fixed-SH Planck feedback (Fig. 3a) demonstrates that even though global-average feedbacks may be similar between AOGCMs and SOM-AGCMs, the spatial structures can be quite different. The Planck feedback is derived from the surface air temperature change. Since feedback parameters are all relative to an increase in the global average surface air temperature of 1 K, differences in the Planck feedback must represent differences in the spatial structure of the surface air temperature increase. The fixed-SH Planck feedback is more negative (i.e., larger in magnitude) throughout the tropics for almost all models, indicating that the tropical temperature increase is larger for AOGCMs (for the same $\Delta\bar{T}_{\text{as}}$). The Planck feedback is generally stronger throughout the Northern Hemisphere (NH) subtropics and midlatitudes, and many, but not all, AOGCMs show a larger feedback in the Arctic. The

FIG. 3. As in Fig. 1, but for annual zonal-average feedbacks in SOM-AGCM experiments subtracted from feedbacks in SRESa1b AOGCM experiments.

feedback difference generally becomes more positive (weaker AOGCM feedback) moving from the tropics to the Southern Hemisphere midlatitudes, and almost all AOGCMs have weaker Planck feedbacks in the Southern Ocean compared with SOM-AGCMs. Thus, AOGCMs tend to have larger surface air temperature changes in the tropics and smaller changes in the Southern Ocean. Note that although $\Delta \bar{T}_{as}$ is the same for both model types, the global average Planck feedback difference does not have to be zero, since temperature changes at some locations are more effective at altering

the TOA energy budget than those at other locations. In general, tropical changes are more effective than high-latitude changes (Soden et al. 2008). In fact, Fig. 1a shows the AOGCMs have slightly larger (more negative) global average fixed-SH Planck feedbacks.

For the models where multiple ensemble members are available, the overall patterns of zonal feedback (for the Planck as well as other feedbacks) are similar across ensemble members, although the patterns have different magnitudes or are shifted to be more positive or negative. To the extent that the patterns are similar, the underlying

FIG. 4. As in Fig. 2, but for annual zonal-average feedbacks in SOM-AGCM experiments subtracted from feedbacks in SRESa1b AOGCM experiments.

dynamics of a given model are likely similar, though of a greater or lesser magnitude.

Figure 3b shows a broad region of more negative lapse rate feedbacks for the AOGCMs, spanning both the tropics and midlatitudes. For the same ΔT_{as} , the tropical and subtropical lapse rates decrease more in the AOGCMs. In the tropics and subtropics, the lapse rate remains close to the moist adiabatic lapse rate (which decreases with temperature). Consistent with our Planck

feedback results, this implies a larger temperature change in the tropics relative to high latitudes. Following Fig. 3b of Soden and Held (2006), we plot the ratio of the tropical-average (30°N–30°S) to global-average ΔT_{as} versus the global-average lapse rate feedback for all experiments (Fig. 5). The SOM-AGCM experiments (squares) tend to have smaller lapse-rate feedbacks and tropical-to-global-average ratios than their corresponding AOGCMs, with only a few exceptions. The water vapor

FIG. 5. Ratio of the tropical-average (30°N–30°S) to global-average surface air temperature changes as a function of the global-average lapse rate feedback, after Soden and Held (2006), for SRESa1b AOGCM (circles), 1%to2x AOGCM (stars), and doubled CO₂ SOM-AGCM (squares) experiments. Colors are as in Fig. 1.

feedback is also related to ratio (not shown); however, the correlation is less, since water vapor and lapse rate feedbacks are not perfectly correlated.

The water vapor feedback differences (Fig. 3c) are strongest in the tropics, decreasing toward the poles. With a few exceptions, the AOGCM feedbacks are larger (more positive) than the SOM-AGCM feedbacks from about 30°S to 90°N. Three models, however, have weaker equatorial water vapor AOGCM feedbacks (INM-CM3.0, MRI CGCM2.3.2a, and NCAR CCSM3.0). Note that these models have some of the closest correspondences between model types for the global-average λ_0 feedback (Fig. 1c; Table 2) due to cancellation of the generally stronger tropical and subtropical feedbacks of these AOGCMs by their weaker feedbacks at the equator. INM-CM3.0 does not have as negative of an equatorial AOGCM to SOM-AGCM difference for the 1%to2x scenario (not shown), corresponding to a more positive global average feedback difference, and indicating some scenario dependence of the zonal feedbacks. The broad increase in water vapor feedback indicates that, for the same increase in $\Delta \bar{T}_{as}$, the AOGCMs have larger increases in water vapor over a wide tropical region, compared with the corresponding SOM-AGCMs. Dessler and Wong (2009) have linked larger water vapor feedbacks to larger tropical temperature changes

relative to global changes, and we do in fact find larger increases in tropical atmospheric temperature relative to the high latitudes for the AOGCMs (Fig. S3d of the supplemental material). That is, the AOGCM simulations, for the same global average surface air temperature change, have more peaked atmospheric temperature changes, with larger temperature changes than SOM-AGCMs in the tropics and smaller temperature changes in high latitudes (again in agreement with the implied temperature changes for the Planck results).

When we consider the fixed-RH framework, we see once again that including the water vapor changes needed to maintain constant RH reduces the differences in the Planck (Fig. 3d) and lapse rate (Fig. 3e) feedbacks. However, local changes in relative humidity are responsible for the sharp peaks and troughs in the subtropics and tropics of some models (Fig. 3f). Figure S3c of the supplemental material confirms that the broad increase in water vapor feedback from the SOM-AGCMs to the AOGCMs can be reproduced by assuming constant relative humidity, following the procedure of Colman (2004).

The differences in surface albedo feedbacks between AOGCMs and SOM-AGCMs have opposite signs in the northern versus southern high latitudes. The NH surface albedo feedback (Fig. 4a) is generally more positive (larger) in the AOGCMs, whereas the Southern Hemisphere albedo feedback is generally more positive in the SOM-AGCMs, in agreement with Boer and Yu (2003c) and Hwang and Frierson (2010). The surface albedo feedback is partially determined by initial sea ice in the Southern Ocean, which SOM-AGCMs generally overestimate, resulting in larger sea ice feedbacks (Yokohata et al. 2008). Additionally, in a warmer climate, changes in the net vertical transport of heat into the Southern Ocean result in net heat uptake, damping the surface temperature change and reducing the sea ice loss for AOGCMs compared with SOM-AGCMs (e.g., Manabe et al. 1991; Gregory 2000). In the Arctic, increased poleward ocean heat transport in AOGCMs (Holland and Bitz 2003) results in larger temperature and sea ice changes compared with SOM-AGCMs; the initial sea ice condition also influences the ice–albedo feedback. In agreement with Fig. 3 of Yokohata et al. (2008), we find most AOGCMs have a larger or near-neutral surface albedo feedback in the NH compared with SOM-AGCMs, though the Southern Hemisphere albedo feedback differences are normally of larger magnitude. However, for some models, these NH surface albedo feedback increases do cancel the Southern Hemisphere feedback decreases.

Relative to the SOM-AGCMs, the larger Arctic and smaller Antarctic temperature responses of the AOGCMs

(Fig. 3a) result in a larger water vapor increase in the Arctic and a smaller increase in the Antarctic and, hence, a slightly more positive water vapor feedback in the Arctic, with a more negative water vapor feedback in the Antarctic (Fig. 3c). Furthermore, at high latitudes, the lapse rate is controlled by not the moist adiabatic lapse rate, but rather baroclinic instability, and high-latitude lapse rate feedbacks are positive (Bony et al. 2006). Therefore, the larger NH temperature change in AOGCMs corresponds to a more positive lapse rate feedback in the Arctic and a more negative lapse rate feedback in the Antarctic compared with SOM-AGCMs (Fig. 3b).

The higher tropical-to-global- ΔT_{as} ratios for AOGCMs, as well as the interhemispheric gradient of temperature responses, suggest differences in meridional heat transport between the two model types. Held and Soden (2006) find that the ensemble average column poleward heat transport increases for both AOGCMs and SOM-AGCMs. The present analysis (Fig. 6) indicates that, while the ensemble average suggests an increase in poleward heat transport, individual SOM-AGCMs and AOGCMs show a wide range of meridional heat transport responses to an imposed forcing. Similar to Hwang and Frierson (2010), models tend to have increased poleward column heat transport in a warmer climate, but the magnitude varies quite a bit from model to model. Some models even show equatorward heat transport (e.g., CCSM3 in the Southern Hemisphere). When comparing the responses of the two model types (Figs. 6b,d), AOGCMs tend to have a smaller increases in NH poleward heat transport than SOM-AGCMs, in agreement with Held and Soden (2006), and a mix of larger and smaller increases in the Southern Hemisphere (depending on both the model and the scenario). Note that the feedbacks themselves alter the TOA radiation, in turn affecting the poleward energy flux (Zelinka and Hartmann 2012).

Zonal SW cloud feedbacks (Fig. 4c) show no consistent signs at any latitude, in agreement with Fig. 7 of Yokohata et al. (2008). They do not find one specific process responsible for the differences in cloud feedbacks between the two model types but rather find different dominant processes for different models. Similarly, the LW cloud feedbacks (Fig. 4d) show no obvious model-independent behavior.

5. Conclusions

Feedbacks in AOGCMs differ from those in SOM-AGCMs. Most previous work examining these differences has focused either on cloud versus noncloud feedbacks or on details of the SW feedbacks. By using

the radiative kernel technique, we are able to isolate differences in LW feedbacks (lapse rate and water vapor) as well as analyze them further with the fixed-RH framework. Using 6 times more simulations, we confirm the earlier results of Yokohata et al. (2008), especially concerning the varied model behavior of the cloud and surface albedo feedbacks; however, the difference between AOGCM and SOM-AGCM feedbacks is more complicated than a simple relationship between SW cloud feedback and total feedback. The standard deviation of the SW cloud feedback difference is larger than any of the individual noncloud feedbacks, but the standard deviation of the sum of the noncloud feedback differences is only 10% smaller, indicating a significant role of the noncloud feedbacks in determining the difference between AOGCM and SOM-AGCM climate sensitivities. Additionally, the SW cloud feedback difference varies greatly among models and does not even have a consistent sign difference. Thus, without knowing something about a model, we cannot estimate this difference ahead of time. There is no common correction that can be applied to all SOM-AGCMs, for example. This work focuses on those aspects of the differences in feedbacks between model types that are consistent across models. These similarities provide information about the underlying dynamics of the model, tying feedbacks to the latitudinal and vertical structure of the temperature changes.

Both global-average (positive) water vapor and (negative) lapse rate feedbacks are consistently larger (more positive and more negative, respectively) in AOGCMs. Cancellation between global-average lapse rate and water vapor feedback differences partially “hides” them when compared, for example, to SW cloud feedback differences, but the consistencies of these differences across models point to common fundamental processes. Additionally, in terms of model development and validation, it is important that models do not get the “right” answer for the “wrong” reasons.

Meridional patterns of the feedbacks are different, indicating altered interactions with dynamics, with implications for other feedbacks. Zonal-average water vapor feedback differences between the two model types (AOGCM minus SOM-AGCM) peak in the tropics, although a few models have compensating differences in the deep tropics. Zonal-average lapse rate feedbacks are more negative in AOGCMs than SOM-AGCMs at most latitudes except in the Arctic. Global-average surface albedo feedbacks show no consistent differences between model types due to large variability of and near-cancellation by Arctic and Antarctic responses. On the other hand, cloud feedbacks, which are more dependent on the specific details of the

FIG. 6. Change in northward heat transport between the control and experiment scenarios for (a) SRESa1b, (c) 1%to2x, and (e) SOM-AGCMs experiments, and differences between the (b) SRESa1b or (d) 1%to2x heat transport changes and the corresponding SOM-AGCM change. Colors are as in Fig. 1.

modeled processes, show no consistent differences, globally or zonally, even though the differences are often large in magnitude.

Although most of these TOA flux differences become small when averaged globally or combined with other covarying feedbacks, the large local differences affect regional energy budgets. Boer and Yu (2003b) propose that the spatial pattern of feedbacks determine, to first order, the spatial pattern of the temperature response. Thus, improving the meridional structure of feedbacks in models could improve projected regional responses to anthropogenic forcing. Also, as emphasized by Zelinka and Hartmann (2012), small global-average feedbacks can still have a large effect on anomalous energy transport due to their high variation with latitude. Near-zero global feedback differences can indirectly affect climate sensitivity through this anomalous energy transport.

The consistent zonal difference patterns (for non-cloud feedbacks) suggest that similar large-scale processes are involved for almost all models. Tropical water vapor feedbacks are larger in AOGCMs due to larger increases in tropical atmospheric temperature for the same global average surface air temperature change, and the increased lapse rate feedback is related to the ratio of tropical to global ΔT_{as} as well. Feedbacks calculated using the fixed-RH framework indicate that most of the broad pattern of water vapor feedback differences results from changes in specific humidity needed to maintain a constant relative humidity with the larger tropical atmospheric temperature response. Large interhemispheric contrasts in the sign of high-latitude surface albedo, lapse rate, and water vapor feedback differences decrease the global averages and implicate differences in oceanic heat transport, which is essentially held fixed in SOM-AGCMs experiments.

AOGCMs have larger Arctic ocean heat transport increases in response to imposed forcings than SOM-AGCMs, resulting in larger surface and atmospheric temperature changes, leading to larger (more positive) surface albedo and water vapor feedbacks and smaller (less negative) lapse rate feedbacks. However, when combined with the larger radiative damping (i.e., Planck feedback), the noncloud feedback parameter for NH middle and high latitudes (Fig. 4b) can be either larger or smaller for AOGCMs when compared to SOM-AGCMs. In the tropics and subtropics, AOGCMs have larger temperature and water vapor changes, with a net result of a more negative feedback parameter than SOM-AGCMs. In the high southern latitudes, AOGCMs have smaller temperature changes than SOM-AGCMs, generally leading to weaker surface albedo and water vapor and stronger lapse rate feedbacks. Again, the net sign of the noncloud feedbacks can be either larger or

smaller. For some models [e.g., Goddard Institute for Space Studies Model E-R (GISS-ER)] the noncloud feedbacks result in more cooling (versus the SOM-AGCM) in low latitudes and less cooling at high latitudes. This would require less poleward heat transport, whereas other models have a tendency for heat to move anomalously northward or southward. The influence of the cloud feedbacks, of course, complicates this heat transport effect, and a complete analysis of the meridional heat budget is beyond the scope of this paper.

These differences in feedbacks between the AOGCMs and the SOM-AGCMs indicate that $\Delta X/\Delta \bar{T}_{\text{as}}$ is different. One possibility is a dependence of the climate response on the initial condition. The dependence on climate state has been identified as a factor in other ensembles, such as perturbed-physics experiments with SOM-AGCMs (Yoshimori et al. 2011). SOM-AGCMs generally do not replicate the sea surface temperature and sea ice patterns of their corresponding AOGCMs (Yokohata et al. 2008; Bitz et al. 2012). Development of a SOM that does produce the same ocean surface and sea ice state as its corresponding AOGCM requires care. Bitz et al. (2012) obtain a higher sensitivity when they update the SOM for CCSM to better match the AOGCM climatology and include a dynamical sea ice model. A dependence on initial condition (climate state) would suggest that the effective sensitivity changes with time, since time dependence is a result of a varying climate state. Alternately (or additionally), $\Delta X/\Delta \bar{T}_{\text{as}}$ may vary because differences in ocean dynamics and heat transport alter the climate response. These heat transport differences may decrease as the models approach new steady states, so that equilibrium coupled model feedback magnitudes may be the same as the SOM-AGCM feedbacks. On the other hand, persistent differences in the meridional structure of atmospheric and ocean heat transport may result in equilibrium feedback differences, even after the net ocean heat uptake goes to zero. We are currently studying longer AOGCM experiments to quantify how equilibrium feedbacks depend on scenarios and model.

Acknowledgments. This work was supported by National Science Foundation Grant ATM-0904092. We acknowledge the modeling groups, the Program for Climate Model Diagnosis and Intercomparison (PCMDI), and the WCRP's Working Group on Coupled Modelling (WGCM) for their roles in making available the WCRP CMIP3 multimodel dataset. Support of this dataset is provided by the Office of Science, U.S. Department of Energy. Discussions with Cecilia Bitz, Isaac Held, Mark Zelinka, and Alex Jonko and comments from anonymous reviewers contributed to this work.

APPENDIX

Expansions of Model Names

Expansions of model names are given in Table A1.

TABLE A1. Expansions of model names.

CCCma CGCM3.1	Centre for Climate Modeling and Analysis Coupled GCM, version 3.1
CSIRO Mk3.0	Commonwealth Scientific and Industrial Research Organisation Mark version 3.0
GFDL CM2.0	Geophysical Fluid Dynamics Laboratory Climate Model version 2.0
GISS-ER	Goddard Institute for Space Studies Model E-R
HadCM2	second climate configuration of the Met Office Unified Model
HadGEM1	Hadley Centre Global Environmental Model version 1
INM-CM3.0	Institute of Numerical Mathematics Coupled Model, version 3.0
MIROC 3.2(hires)	Model for Interdisciplinary Research on Climate 3.2, high-resolution version
MIROC 3.2(medres)	Model for Interdisciplinary Research on Climate 3.2, medium-resolution version
MPI ECHAM5	Max Planck Institute ECHAM5
MRI CGCM2.3.2a	Meteorological Research Institute Coupled General Circulation Model, version 2.3.2a
NCAR CCSM3.0	National Center for Atmospheric Research Community Climate System Model, version 3

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