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Control of land-ocean temperature contrast by ocean heat uptake

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[1] We show that the ratio of observed annual-mean land temperature change to ocean surface temperature change, ϕ, has remained almost constant during 1955–2003. This is the case, despite most of the heat capacity of the climate system lying in the oceans, and rapid variations in climate forcing. Examining seven General Circulation Models (GCMs), we find land and ocean temperature behavior comparable to observations in six cases. For three models that reproduce observed ϕ and for which we have data, we find no significant changes in future ϕ under the SRES A1B scenario. We suggest that variations in land-ocean heat flux primarily balanced by ocean heat uptake are sufficient to maintain constant ϕ. This flux is present in the six GCMs that resemble observations, suggesting that observed ϕ may remain constant into the future, even if radiative forcing is markedly different than in the past. **Citation:** Lambert, F. H., and J. C. H. Chiang (2007), Control of land-ocean temperature contrast by ocean heat uptake, *Geophys. Res. Lett.*, *34*, L13704, doi:10.1029/2007GL029755.

1. Introduction

[2] One of the most robust features of observed and modeled climate is the constant ratio of mean land to mean ocean temperature anomalies, ϕ [*Huntingford and Cox*, 2000; *Sutton et al.*, 2007]. Unlike climate sensitivity and ocean heat uptake, ϕ is well constrained by observations, because its estimation depends only on values of temperature and not on climate response time. As such, it is a powerful test of General Circulation Models (GCMs) that are used to simulate climate change.

[3] *Sutton et al.* [2007] showed that GCM ϕ is reasonably independent of global mean temperature change on 20 year timescales, and similar to observed ϕ calculated from average temperature anomalies for 1980–2004. Given transient forcings and the relatively long response time of the ocean relative to the land, there is no a priori reason to suspect that this ratio should remain constant on shorter timescales. However, we show that observed annual mean values deviate little from the mean. Modeling ϕ with a land and ocean energy balance model in which land-ocean heat transport is proportional to land-ocean temperature difference, we find that the observed constancy cannot be reproduced, implying that additional physical processes are necessary. *Shin et al.* [2006] found that GCM land-ocean heat flux variations are significant and principally tied to ocean heat uptake. Here, we show that it may be this flux that maintains constant ϕ. An implication is that future values of ϕ may be similar to those observed, even if the time dependence of future radiative forcing of climate change is different.

2. Observed and GCM Data

[4] We take observed surface temperature data from the land-based CRUTEM3 and ocean-based HadSST2 data sets prepared by the UK Met Office and the University of East Anglia for the period 1955–2003 [*Brohan et al.*, 2006]. For each gridbox, we consider only years in which there are no missing months, although this criterion is unimportant.

[5] We take 20th century temperature, and surface and Top Of Atmosphere (TOA) energy flux GCM data from the Intergovernmental Panel on Climate Change (IPCC) portal for the NCAR CCSM3 (5), GFDL CM2 0 (3), GFDL CM2 1 (3), GISS E-R (9), MIROC (3), MRI (5) and NCAR PCM (4) models (https://esg.llnl.gov:8443/index.jsp). The number of available ensemble members is given in brackets. The GCMs are driven with historical estimates of anthropogenic greenhouse gas and sulphate aerosol concentrations, volcanic aerosol concentrations and changes in solar irradiance. For full details of these and additional forcings applied to some models, see *Stone et al.* [2007]. For the GFDL CM2 0 (hereinafter GFDL0), GFDL CM2 1 (hereinafter GFDL1), MIROC and MRI models, we also have data for one run from 2000–2300 based on the SRES A1B scenario of *Nakicenovic et al.* [2000] and one unforced pre-industrial control run. A1B is a strongly greenhouse gas forced future that levels out at a CO_2 concentration of 750 ppm in 2200 (about double present-day values). These runs continue for another 100 years at constant forcing.

3. Regression Analysis of Land and Ocean Temperatures

3.1. Land-Ocean Temperature Ratio

[6] We begin by comparing annual mean land temperature anomalies, ΔT_L, to annual mean ocean temperature anomalies, ΔT_O, for 1955–2003. Anomalies are calculated with respect to 1961–90 gridbox means in each case. Using ordinary least squares regression, we find that observed $\phi = \frac{\Delta T_L}{\Delta T_O} = 1.55 \pm 0.23$ where ΔT_O is the independent variable and the uncertainties are a 5–95% range. Because both ΔT_O and ΔT_L contain noise from internal climate variability and measurement error, we re-calculate using ΔT_L as the independent variable, and find $\phi' = 1.76 \pm 0.26$. Residual autocorrelation is small in both cases. Hence, our result is broadly consistent with the *Sutton et al.* [2007] estimate of $\phi \sim 2$ for 1980–2004 mean ΔT_L and ΔT_O.

[7] Calculating ϕ and ϕ' for the same period and over the same areas available to observations in the GCMs, we find that most are consistent with observations, Table 1. The exception is MRI, which shows smaller values. The observed

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Table 1. Observations for 1955–2003 and Seven GCMs for 1955–2003 and Three Subsequent Periods in the A1B Scenario

	ϕ[a]				r	r_5	ϕ'	
Model	1955–2003	2004–2052	2053–2101	2102–2296	1955–2003	1955–2003	1955–2003	β[b]
Observations	1.55 ± 0.23	—	—	—	9.2×10^{-3}	2.4×10^{-3}	1.76 ± 0.26	—
CCSM3	1.52 ± 0.18	—	—	—	0.023	4.8×10^{-3}	2.12 ± 0.26	—
GFDL0	1.38 ± 0.22	1.51 ± 0.04	1.28 ± 0.12	1.40 ± 0.08	0.020	4.0×10^{-3}	1.95 ± 0.31	~0.1
GFDL1	1.49 ± 0.18	1.52 ± 0.04	1.27 ± 0.14	1.41 ± 0.09	0.021	3.3×10^{-3}	1.86 ± 0.23	~0.5
GISS E-R	1.52 ± 0.12	—	—	—	0.013	1.5×10^{-3}	2.02 ± 0.16	—
MIROC	1.49 ± 0.22	1.41 ± 0.03	1.35 ± 0.07	1.38 ± 0.04	0.016	3.3×10^{-3}	2.02 ± 0.30	~0.2
MRI	0.88 ± 0.14	1.30 ± 0.08	1.21 ± 0.30	1.28 ± 0.16	0.029	4.1×10^{-3}	1.54 ± 0.25	~1.4
PCM	1.37 ± 0.18	—	—	—	0.023	2.2×10^{-3}	1.90 ± 0.25	—

[a] ϕ calculated with ΔT_L as the independent variable.
[b] β in $\mathrm{Wm^{-2}K^{-1}}$, estimated from the A1B scenario runs.

and modeled data are shown in Figure 1. *Sutton et al.* [2007] found that ϕ calculated by their method was generally smaller in GCMs than in observations and attributed this to variability in the observations. However, they used global land and ocean values to estimate GCM ϕ, as opposed to observed area only values. Substituting global values in our analysis tends to reduce ϕ toward unity (see auxiliary material).[1]

[8] Continuing into the future, we calculate ϕ for the A1B scenario runs for 2004–2052, 2053–2101 and the stabilization period 2101–2296. Despite the much smoother radiative forcing of the A1B scenario, values of ϕ are very similar to the 20th century values, Table 1. Interestingly, A1B ϕ in MRI is now consistent with the other models.

3.2. Residual Variance

[9] The quantity $r = variance[\Delta T_L - \phi\Delta T_O]$ is a measure of the degree to which individual annual mean ΔT_L deviate from $\phi\Delta T_O$. Values of r during 1955–2003 are fairly consistent across models, although that in GISS E-R is relatively small and that in MRI is large, Table 1. Remarkably, observed r is smaller than in all of the GCMs, even though the observations contain measurement error [*Brohan et al.*, 2006]. We find similar results for 5-year mean r, r_5.

4. Observed Departure From Constant ϕ

[10] Given that the majority of the heat capacity of the climate system resides in the oceans and that 20th century forcing varied rapidly, how small a value of r should we expect in the observations? Consider the simplest two-box Energy Balance Model (EBM) of land and ocean climate. ΔT_O is given by

$$\Delta Q - \frac{\Delta T_O}{\lambda_O} + \frac{\Delta A}{1-f} = \Delta U_O, \tag{1}$$

where ΔQ is the radiative forcing due to external factors, λ_O is the ocean-only climate sensitivity parameter, $\frac{\Delta A}{1-f}$ is the anomalous atmospheric land-ocean heat transport, f is the land fraction and ΔU_O is the ocean heat uptake. ΔT_L is given by

$$\Delta Q - \frac{\Delta T_L}{\lambda_L} - \frac{\Delta A}{f} = \Delta U_L, \tag{2}$$

where λ_L is the land sensitivity parameter and ΔU_L is the land heat uptake. For simplicity, we assume with previous authors that ΔQ is globally uniform [*Huntingford and Cox*, 2000].

[11] *Huntingford and Cox* [2000] showed that such a model does not maintain constant ϕ when compared to HadCM3. Here, we ask if it can replicate observations satisfactorily. We calculate EBM values of r and r_5, r_{EBM} and r_{EBM5}, when it is driven by historical ΔQ for 1955–2003 calculated from the GISS model and principally derived from *Hansen and Sato* [2001], *Lean et al.* [2002] and *Menon et al.* [2002]. (Other forcing series give similar results, but these cover the entire period of our study.) By varying model parameter values, we consider a wide range of possible climates. We calculate values of $\frac{1}{\lambda_O}$ and $\frac{1}{\lambda_L}$ consistent with global climate sensitivities of $0.9 \leq \frac{1}{\lambda} \leq 3.7\ \mathrm{Wm^{-2}K^{-1}}$, based on a study of Earth Radiation Budget Experiment data by *Forster and Gregory* [2006] and our observed estimates of ϕ (see auxiliary material). (We do *not* hold ϕ constant during our EBM runs.) We set the box-box energy transport term $\Delta A = \beta(\Delta T_L - \Delta T_O)$ in common with earlier EBM studies (e.g. *Murphy* [1995]), and estimate β from the difference between land-ocean energy transport in the final most stable 50 years of the A1B scenario runs and unforced pre-industrial control runs, Table 1. We set $\Delta U_O = c_O\frac{d\Delta T_O}{dt}$ and $\Delta U_L = c_L\frac{d\Delta T_L}{dt}$, where c_O and c_L are effective heat capacities of the ocean and land respectively. From linear trends in HadSST2 and the ocean heat content data of *Levitus et al.* [2005], we estimate c_O for 1955–2003 as $11.45\ \mathrm{Wm^{-2}K^{-1}yr^{-1}}$. This estimate is crude, and hence we allow c_O to vary between 0 and $40\ \mathrm{Wm^{-2}K^{-1}yr^{-1}}$. We allow c_L to vary between 0 and $\frac{c_O}{2}$. Adding estimates of the random errors affecting CRUTEM3 and HadSST2 (but neglecting bias, which is small during 1955–2003) [*Brohan et al.*, 2006], we produce pseudo-observations and calculate ϕ, r_{EBM} and r_{EBM5} as above.

[12] Values of r_{EBM} and r_{EBM5} are relatively large compared with observed r_{obs} and r_{obs5} for most combinations of parameters, even though the EBM does not produce internal climate variability. Figure 2 shows "confidence level" slices through the r_5 parameter space for $\frac{1}{\lambda}$ and c_O plotted against $\frac{c_L}{c_O}$, calculated by assuming that $\frac{r_{EBM5}}{r_{obs5}}$ follows an F-distribution. Above the 90% contour, where we conclude that $r_{EBM5} > r_{obs5}$, the EBM simulates unsatisfactorily large deviations from constant ϕ. Hence, we can rule out the EBM as a good model unless $\frac{1}{\lambda}$ is unrealistically large or c_O is unrealistically small for most reasonable values of $\frac{c_L}{c_O}$. (On climatological timescales, values of $\frac{c_L}{c_O}$ larger than about 0.2 seem unlikely

[1] Auxiliary materials are available at ftp://ftp.agu.org/apend/gl/2007gl029755. Other auxiliary material files are in the HTML.

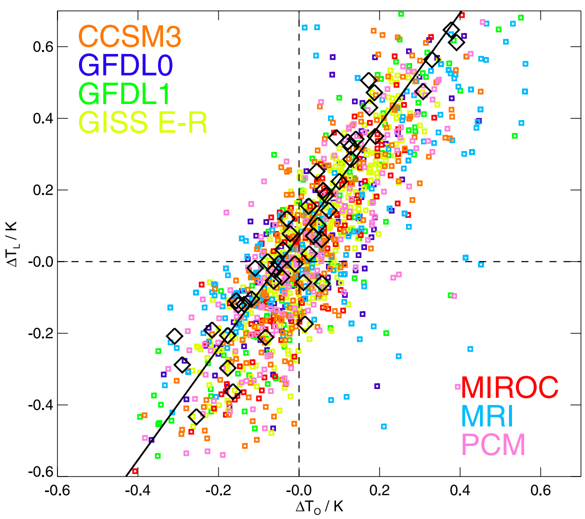

Figure 1. Annual mean ΔT_L plotted against annual mean ΔT_O for the observations (black diamonds) and seven GCMs (colored squares are individual ensemble member years). The solid line is the ordinary least squares best fit to the observations.

based on the work of *Beltrami et al.* [2002] and *Huang* [2006].) r_{EBM5} is very insensitive to varying β and ϕ (not shown). Replacing $c_O \frac{d\Delta T_O}{dt}$ with observed ocean heat uptake values from *Levitus et al.* [2005], also available from 1955–2003, only increases values of r_{EBM5}. If we instead plot confidence level slices for annual mean r_{EBM}, then almost the entire parameter space lies above the 90% contour. However, the appropriate distribution of $\frac{r_{EBM}}{r_{obs}}$ is difficult to determine because of interannual autocorrelation in values of r_{EBM}.

[13] Speculating that small r_{obs5} was obtained by chance, we employ the mean value calculated from the six GCMs that reproduce observed ϕ, $r_{GCM5} = 3.2 \times 10^{-3}$. In this case the 90% confidence region is somewhat smaller, Figure 2. Even accepting r_{GCM5}, we conclude that the EBM lacks the necessary physical processes to simulate the observed constancy of 20th century land-ocean temperature contrast.

5. Land-Ocean Heat Transport

[14] What other processes can explain the small departure of observed temperatures from mean ϕ? Inspection of equations 1 and 2 reveals two options. First, the effective heat capacity of the ocean under rapid volcanic and solar forcing may be smaller than under more gradual anthropogenic greenhouse gas and aerosol forcing, allowing ΔT_O to respond more rapidly to rapid forcing changes. We imagine that this must be the case to some extent, as a short term spike in forcing is not expected to mix deeply into the ocean, even though volcanic eruptions have had a significant impact on ocean heat content [*Church et al.*, 2005]. Proving that c_O is variable in the observations or GCM data is difficult, however, as ocean heat content shows variations uncorrelated to ΔT_O. Nevertheless, we note that regression estimates of GCM c_O are smaller on below 5 year timescales than on 5–10 year timescales. 5–10 year c_O values, on the other hand, are consistent with 10–20 year values (see auxiliary material).

[15] The second possibility is that constant ϕ is maintained by a powerful land-ocean atmospheric heat flux, ΔA, that retards changes in ΔT_L. Subtracting equation 2 from equation 1 and dividing by ΔT_O gives

$$\frac{\phi}{\lambda_L} - \frac{1}{\lambda_O} = \frac{1}{\Delta T_O}\left[\Delta U_O - \Delta U_L - \frac{\Delta A}{f(1-f)}\right]. \quad (3)$$

Requiring ϕ almost constant, means setting the left-hand side almost constant: $-\alpha \simeq \frac{\phi}{\lambda_L} - \frac{1}{\lambda_O}$. Rearranging, we find

$$\Delta A \simeq f(1-f)(\Delta U_O - \Delta U_L + \alpha \Delta T_O). \quad (4)$$

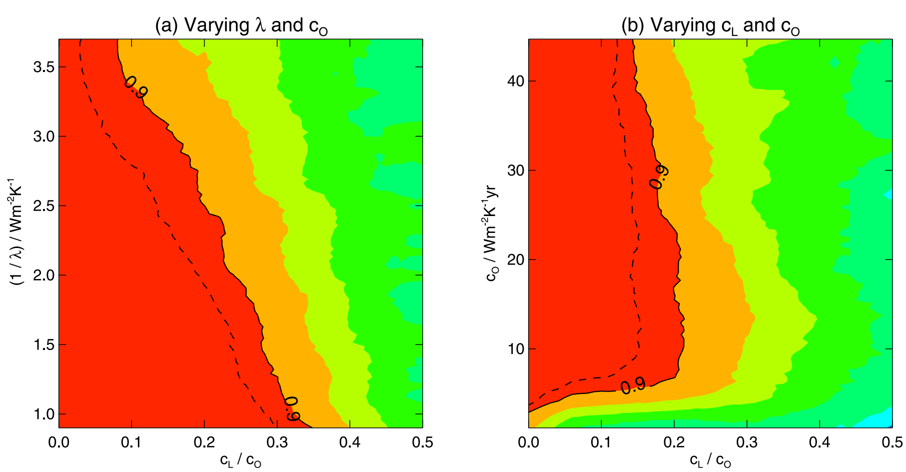

Figure 2. The confidence level at which $r_{EBM5} > r_{obs5}$, assuming that $\frac{r_{EBM5}}{r_{obs5}}$ is distributed *F*. The labeled solid contour and red region to the left of each panel indicate 90% confidence. From left to right, confidence decreases by 10% at each color boundary. The dashed line is the 90% contour where we use GCM mean r_5. Both panels are smoothed to reduce the effect of variations in EBM pseudo measurement error, allowing us to plot clean contours. (a) Varying $\frac{1}{\lambda}$ and $\frac{c_L}{c_O}$ with c_O = 11.45 Wm^{-2}K^{-1}yr^{-1}, ϕ = 1.55 and β = 0.19. (b) Varying c_O and $\frac{c_L}{c_O}$ with ϕ = 1.55, β = 0.19 and $\frac{1}{\lambda}$ = 2.3 Wm^{-2}K^{-1}.

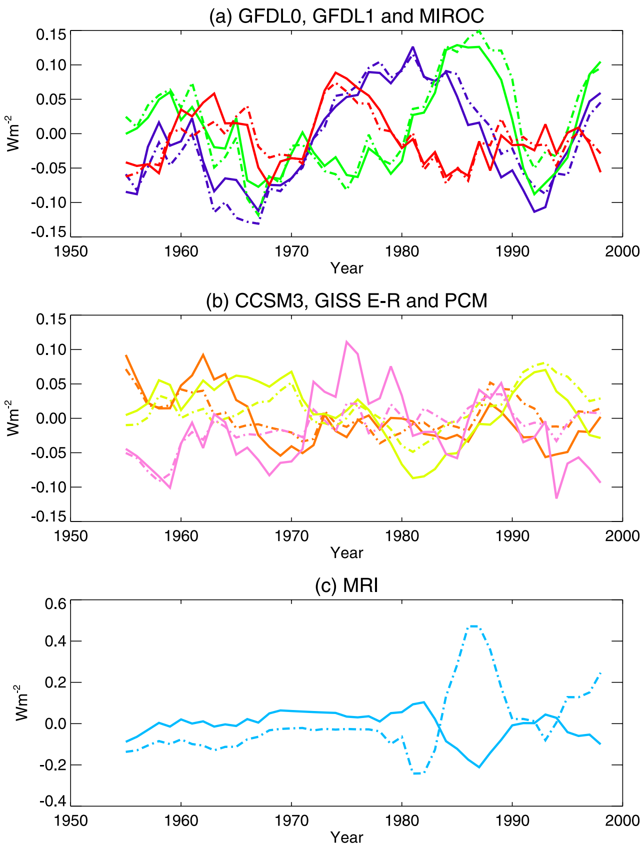

Figure 3. Values of ΔA (solid lines) and $f(1-f)(\Delta U_O - \Delta U_L + \alpha\Delta T_O)$ (dot-dashed lines) in seven GCMs. (a) GFDL0 (dark blue), GFDL1 (green) and MIROC (red), for which α is available and r is small. (b) CCSM3 (orange), GISS E-R (yellow) and PCM (pink), for which α is unavailable. (c) MRI (light blue), for which α is available and r is large. Note the different y-axis scale on this panel.

Because ΔT_L is now a function of ΔT_O, the last term, $\alpha\Delta T_O$, is equivalent to our previous estimate of $\Delta A = \beta(\Delta T_L - \Delta T_O)$, with $\alpha = \frac{\beta(\phi-1)}{f(1-f)}$. In Section 4 we saw that this term alone is unable to maintain constant ϕ. Land heat uptake is also quite small [*Beltrami et al.*, 2002] (although it is large in some GCMs). Hence, if changes in effective ocean heat capacity are unimportant, variations in land-ocean heat transport must be primarily balanced by ocean heat uptake.

[16] Because the heat capacity of the atmosphere is small on climatological timescales, summing TOA and surface radiative, sensible and latent heat fluxes over land or ocean yields ΔA. (Here we use global rather than observed area only quantities.) Land and ocean estimates of ΔA are in good agreement in all seven cases once we remove a global-mean correction due to non-conservation of energy in some GCM atmospheres. We then compare this to $f(1-f)(\Delta U_O - \Delta U_L + \alpha\Delta T_O)$, omitting $\alpha\Delta T_O$ for CCSM3, GISS E-R and PCM, for which we have no value of α, Figure 3.

[17] In the GFDL0, GFDL1 and MIROC models, the agreement is excellent, Figure 3a. It is less good in CCSM3, GISS E-R and PCM, but still quite convincing, Figure 3b. This does not appear to be because $\alpha\Delta T_O$ is omitted from our estimate in these models, as employing a range of values of α does not help. In MRI, there is no agreement, and TOA flux differences are the largest component of ΔA, Figure 3c. Hence, $f(1-f)(\Delta U_O - \Delta U_L + \alpha\Delta T_O)$ approximates ΔA in three of four GCMs for which we have all necessary data, and is an important part of ΔA in the GCMs for which we do not have estimates of $\alpha\Delta T_O$. The exception is MRI, which produces different values of ϕ under the A1B scenario and large r relative to the other models. While the variations of ΔA in Figure 3 are fairly small, dividing by f (equivalent to multiplying by about 3), reveals heat flux variations over land of the order of 0.5–1 $\mathrm{Wm^{-2}}$. These are comparable to changes in radiative forcing over the same period. Looking across the six GCMs apart from GISS E-R over land, we find that Root Mean Square (RMS) variations in ΔA are comparable to variations in ΔU_L and TOA fluxes. Over the ocean, however, RMS variations in ΔA are only about 30% the size of ΔU_O and TOA variations, because of the difference in area of land and ocean. Hence, land-ocean heat flux is more important to ΔT_L than ΔT_O. (ΔU_L and ΔU_O RMS variations are $\sim$10 times larger in GISS E-R than the other models.)

6. Discussion

[18] The observed annual-mean ratio of land to ocean temperature change, ϕ, has remained almost constant during the past 50 years. This is the case, despite rapid variations in radiative forcing that affect climate due to large volcanic eruptions. In six of seven GCMs, we find values of ϕ consistent with the observations for 1955–2003. Extending our analysis into the 21st and 22nd centuries under the A1B scenario in three of the six GCMs that reproduce observations, we find no significant changes in ϕ, suggesting that land-ocean temperature contrast is reasonably independent of radiative forcing timescale and global mean temperature change (see also *Sutton et al.* [2007]).

[19] *Sutton et al.* [2007] argued that 20 year mean ϕ is a function of the relative importance of latent and sensible heating changes that accompany warming over land and ocean. In our language, they stated that ϕ depends on the ratio of climate sensitivity parameters, $\frac{\lambda_L}{\lambda_O}$. We see that at equilibrium, or on longer timescales where we may neglect land and ocean heat uptake, subtracting equation 2 from 1 yields

$$\phi \simeq \frac{\frac{1}{\lambda_O} - \alpha}{\frac{1}{\lambda_L}}. \quad (5)$$

Hence, where we can neglect α, our result is the same as theirs.

[20] The departure of observed land and ocean temperatures from constant ratio is small enough to suggest that ϕ is actively maintained by the climate system. This could be because effective ocean heat capacity is smaller on shorter timescales, allowing the ocean to respond more rapidly to sharp changes in forcing than would otherwise be expected. Although we have not fully explored this possibility, we note that GCM effective ocean heat capacities are consistent on 5–10 year and 10–20 year timescales. An alternative possibility is that a powerful land-ocean heat flux, ΔA, primarily balanced by ocean heat uptake, acts to retard land temperature change under rapid forcing. Such a land-ocean

heat flux is common to most GCMs prepared for the IPCC Fourth Assessment Report [*Shin et al.*, 2006]. While it is not surprising that the atmosphere adjusts to changes in ΔU_O, the prospect that it controls ϕ is interesting. It seems plausible that this is the case in the six GCMs that reproduce observed ϕ, while it is not in the MRI model, which does not reproduce observed ϕ. If real ΔA is similar, then the contribution of land-ocean heat flow changes to the land energy budget is comparable to that of radiative forcing. ΔA may be even larger than predicted by GCMs if observed ocean heat uptake variability is as estimated by *Levitus et al.* [2005], although these results are disputed [*Gregory et al.*, 2004].

[21] The ratio of observed annual-mean land to ocean temperature change has remained almost constant since the 1950s. We have shown that it is likely that constant ϕ is actively maintained by the climate system, and that a land-ocean heat flux primarily balanced by ocean heat uptake is a realistic solution.

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