11-7-2009

Global Warming, Convective Threshold and False Thermostats

Ian N. Williams
University of Chicago

Raymond T. Pierrehumbert
University of Chicago

Matthew Huber
Purdue University, mhuber@purdue.edu

Global warming, convective threshold and false thermostats

Ian N. Williams,[1] Raymond T. Pierrehumbert,[1] and Matthew Huber[2]

Received 3 August 2009; revised 5 October 2009; accepted 13 October 2009; published 7 November 2009.

[1] We demonstrate a theoretically expected behavior of the tropical sea surface temperature probability density function (PDF) in future and past (Eocene) greenhouse climate simulations. To first order this consists of a shift to warmer temperatures as climate warms, without change of shape of the PDF. The behavior is tied to a shift of the temperature for deep convection onset. Consequently, the threshold for appearance of high clouds and associated radiative forcing shifts along with temperature. An excess entropy coordinate provides a reference to which the onset of deep convection is invariant, and gives a compact description of SST changes and cloud feedbacks suitable for diagnostics and as a basis for simplified climate models. The results underscore that the typically skewed appearance of tropical SST histograms, with a sharp drop-off above some threshold value, should not be taken as evidence for tropical thermostats. **Citation:** Williams, I. N., R. T. Pierrehumbert, and M. Huber (2009), Global warming, convective threshold and false thermostats, *Geophys. Res. Lett.*, *36*, L21805, doi:10.1029/2009GL039849.

1. Introduction

[2] The feedbacks affecting tropical climate are of evident interest, both for paleoclimate and anthropogenic climate change problems. From time to time the question has arisen whether some of these feedbacks could impose a rigid upper limit on tropical temperature. Understanding of this and related issues can be greatly aided by an idealized picture of the basic workings of the tropical climate system. The basic concept, as articulated by *Pierrehumbert* [1995], rests on the fact that the state of the entire tropical free troposphere can, to a good approximation, be characterized by a single number, which is the saturated moist entropy. The one-parameter characterization holds in the vertical because moist convection keeps the free troposphere near the moist adiabat, and in the horizontal because efficient heat transports due to large-scale gravity waves and the Hadley circulation enforce horizontal temperature homogeneity. SST is determined by radiant and turbulent exchanges with the free troposphere, which determine SST relative to the free troposphere. Annual mean free-troposphere temperature is determined primarily by the top-of-atmosphere energy balance averaged over the tropics, taking into account horizontal heat exports mediated by atmosphere and ocean. The free-troposphere temperature can increase in response to increasing CO_2 or solar forcing. Since convection sets in when boundary layer air is buoyant with respect to the free troposphere, the SST threshold for deep convection will shift to warmer values as the free-troposphere temperature increases. This simple picture, when elaborated to include realistic radiation, has considerable explanatory power with regard to both tropical precipitation and temperature [*Peters and Bretherton*, 2005]. Following *Sobel and Bretherton* [2000], we'll call this the WTG ("Weak Temperature Gradient") theory.

[3] The aspects of tropical dynamics to be discussed here bear on the class of thermostats proposed by *Newell* [1979] and *Ramanathan and Collins* [1991], in which tropical SST is limited by either evaporation or cloud feedbacks which set in at a trigger temperature erroneously assumed to be a universal constant independent of atmospheric CO_2 concentration. Thermostats inevitably figure into discussions of paleoclimate and paleoclimate proxies [*Crowley*, 2000; *Kilbourne et al.*, 2004] and of cloud feedbacks relevant to future climate projections [*Stephens*, 2005]. In such reviews, it is not always made clear that the trigger class of thermostats can be decisively invalidated based on fundamental physical reasoning. A related point of confusion is the skewness of the tropical SST histogram, with sharp cutoff at high temperature, is often taken as prima facia evidence for a tropical thermostat [e.g., *Kleypas et al.*, 2008]. There is a long history of refuting the trigger class of thermostats [*Wallace*, 1992; *Hartmann and Michelsen*, 1993; *Fu et al.*, 1992; *Pierrehumbert*, 1995; *Sud et al.*, 2008], but the false notion that data and simple physical reasoning supports a tropical thermostat persists, often based on the observation of an apparent cut-off. For example, *Veron* [2008] states "... the surface temperature of the largest oceans would have been limited by the Thermal Cap of ~31C, widely believed to be the highest temperature large oceans can reach." Here, we will examine a suite of general circulation model simulations that clearly support previous work showing that this notion is false.

[4] The object of this Letter is to (a) demonstrate the explanatory power of WTG theory in the Intergovernmental Panel on Climate Change Fourth Assessment Report (AR4) future climate simulations and in paleoclimate (Eocene) simulations of warm climates; (b) show that use of an entropy excess coordinate in place of SST yields a compact description of cloud feedbacks suitable for use in idealized climate models; and (c) show that this description provides a convenient diagnostic of cloud forcing in simulations, and a logical target for use in monitoring tropical climate change.

2. Behavior of the Convective Threshold

[5] We examined simulations from 15 coupled ocean-atmosphere models from the AR4 archive (see auxiliary material), in which equivalent CO_2 concentration starts at

[1]Department of Geophysical Sciences, University of Chicago, Chicago, Illinois, USA.

[2]Department of Earth and Atmospheric Sciences, Purdue University, West Lafayette, Indiana, USA.

0094-8276/09/2009GL039849

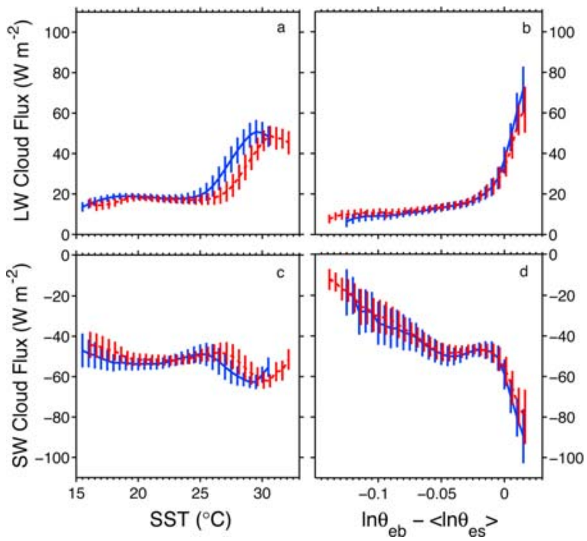

Figure 1. TOA cloud LW flux as a function of (a) SST and (b) s_{diff}; TOA cloud SW flux as a function of (c) SST and (d) s_{diff}; Solid blue and dashed red lines correspond to the ensemble median over years 0–20 and 60–80, respectively, from 15 IPCC AR4 coupled ocean-atmosphere models for the 1% per year scenario. Vertical lines indicate the interquartile range.

pre-industrial levels and increases 1% per year until doubling.[1] Cloud longwave (CLW) and shortwave (CSW) forcing for 30S to 30N was binned against SST, and histograms of SST for this region were computed. When binning cloud radiative forcing (CRF), the values were weighted by the probability of occurrence of the corresponding SST, in such a way that the integral of CRF over SST is proportional to the net CRF. The analysis was performed separately for model years 1–20 (cooler) and 61–80 (warmer). CLW forcing (Figure 1a) clearly reveals the behavior of the convective threshold. Deep convection, leading to high clouds and a sharp rise in cloud greenhouse effect, sets in above a threshold SST, but the threshold increases as one moves from a cool to a warmer climate. CLW and CSW curves (Figures 1a and 1b) shift to the right as climate warms. SST histograms for cool and warm climates separately (Figures 2a and 2c) are skewed and give the appearance of a sharp upper bound on SST; however, the location of the "cliff" in the histogram shifts to warmer temperatures as the climate warms. A similar shift was noted in other simulations [*Dutton et al.*, 2000; *Sud et al.*, 2008].

[6] We wish to replace SST with a coordinate that characterizes the buoyancy controlling where deep convection occurs. Several possible coordinates yielded similar results (conditional instability, CAPE, vertical gradients of moist entropy) so we chose a variant of conditional instability that is easily applied to the model datasets. We define the entropy excess (s_{diff}) as the difference between the logarithm of sub-cloud layer pseudo equivalent potential temperature ($\ln\theta_{eb}$) averaged over the lowest two model levels (1000 and 925 hPa), and pressure-weighted mean logarithm of saturated pseudo equivalent potential temperature between 925 hPa and 500 hPa ($\langle\ln\theta_{es}\rangle$) from monthly mean temperature and relative humidity. The convective threshold is not sensitive to the exact choice of model level in the averages. The atmosphere is stable against convection when s_{diff} is negative, and is unstable when s_{diff} is significantly positive. A comparison of s_{diff} from NCEP-NCAR reanalysis [*Kalnay et al.*, 1996] to the International Satellite Cloud Climatology Project [*Rossow and Schiffer*, 1999] (see auxiliary material) shows that deep convection and associated high clouds set in when $s_{diff} > 0$, whereas low clouds begin to dominate where the boundary layer air is stable relative to the free troposphere ($s_{diff} < 0$). The onset of deep convection in the AR4 models (Figures 1b and 1d), measured by the sharp increase in the CLW and CSW forcing magnitudes, is invariant between warm and cool climates in terms of s_{diff}, which is itself nearly invariant (Figure 2d). Another way of describing the success of the new coordinate is it demonstrates that, to a good approximation, the TOA cloud radiative forcing is a function of the convective instability parameter we have defined, at least in the models. This is an improvement over using SST as a coordinate.

[7] Whether clouds have a stabilizing or destabilizing effect on surface temperature is predominantly determined by net top-of-atmosphere (TOA) cloud forcing. The primacy of the TOA radiation budget was first pointed out clearly by *Manabe and Wetherald* [1967] and emphasized in a widely-read review on radiative-convective modeling [*Ramanathan and Coakley*, 1978], but was left out of a number of subsequent studies that speculated on SST limits [e.g., *Newell*, 1979; *Ramanathan and Collins*, 1991]. Clouds that leave the TOA balance unchanged can affect SST gradients but the mean temperature is inextricably linked to the convective threshold. This result holds in idealized radiative-dynamical models [*Pierrehumbert*, 1995], the WTG approximation [*Peters and Bretherton*, 2005], and in GCMs (e.g., AR4), all of which incorporate fundamental constraints of energy conservation and convective adjustment of temperature toward a moist adiabat. If the reasoning behind "trigger-type" thermostats were valid, it would apply to the GCMs just as well as it would apply in the real world. This situation contrasts with the IRIS conjecture of *Lindzen et al.* [2001], which may be invalid for other reasons [*Hartmann and Michelsen*, 2002], but cannot be tested against an unmodified GCM.

[8] The net TOA cloud forcing as a function of s_{diff} (Figure 2b) declines sharply as s_{diff} becomes more positive, both because the high cloud greenhouse effect arising from deep convection comes to cancel the cloud albedo effect, and because very convectively unstable conditions are rare. Similarly, net CRF drops sharply in strongly stable situations, largely because such conditions are rare but there also seems to be a tendency in these models for shallow clouds to dissipate in very stable conditions. The net cloud effect is a pronounced cooling, dominated by shallow to intermediate depth clouds in mildly suppressed to mildly convective regimes. *Bony and Dufresne* [2005] came to a similar conclusion based on a dynamically defined coordinate.

[9] The s_{diff} coordinate allows us to better focus on cloud feedbacks by providing a reference to which the

[1]Auxiliary materials are available in the HTML. doi:10.1029/2009GL039849.

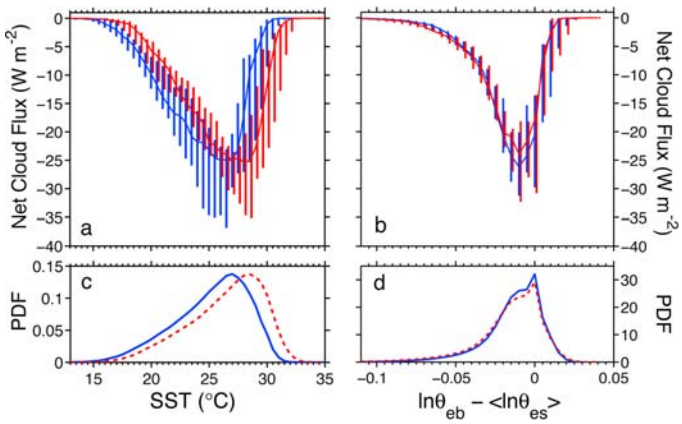

Figure 2. TOA net cloud radiative flux (CRF) as a function of (a) SST and (b) s_{diff} weighted by relative frequency of occurrence for each bin and multiplied by factors of 15.5 and 10.4 to reflect the actual value of CRF at the modes of SST and s_{diff} distributions, respectively. Solid blue and dashed red lines correspond to the ensemble median over years 0–20 and 60–80, from 15 IPCC AR4 coupled ocean-atmosphere models for the 1% per year scenario. Vertical lines indicate the interquartile range. Bottom figures are corresponding probability density functions normalized by 0.5C and 0.005 bin widths for (c) SST and (d) s_{diff}, respectively.

onset of deep convection is invariant. A comparison of the TOA forcing distribution for the warm vs. cool climates reveals that in the 15 models, the dominant effect of warming is to reduce net cloud cooling in intermediately stable conditions. This constitutes a positive cloud feedback, and we see immediately which class of convection regimes is responsible.

3. Cloud Feedbacks and Climate Sensitivity

[10] Cloud radiative feedbacks are a major source of the uncertainty in climate sensitivity estimates. In a previous study [*Bony et al.*, 2004] model cloud radiative feedbacks were composited over grid cells with subsiding vertical velocity to attribute the spread among climate sensitivity estimates to changes in shallow convection. Here, we use s_{diff}, rather than vertical velocity, to separate shallow and deep convection. We divided the ensemble mean into the 7 and 8 models exhibiting stabilizing (negative) and destabilizing (positive) tropical cloud radiative feedbacks (Table S1) and decomposed the change in spatially averaged CRF ($\bar{C}$) into three terms

$$\overline{\delta C} = \int_{-\infty}^{+\infty} C_{s_{diff}} \delta P_{s_{diff}} ds_{diff} + \int_{-\infty}^{+\infty} P_{s_{diff}} \delta C_{s_{diff}} ds_{diff} + \int_{-\infty}^{+\infty} \delta C_{s_{diff}} \delta P_{s_{diff}} ds_{diff} \quad (1)$$

where C is the CRF, P_{sdiff} is the probability of a given s_{diff}, and δ denotes a change between doubled CO_2 and preindustrial climates. In discretized form the right hand side terms are contributions from changes in the probability of s_{diff} being within a range or "bin" (first term), changes in CRF within a bin (second term), and covariation between both changes. We refer to s_{diff} bins as "convective regimes" because deep convection and associated high clouds tend to be more abundant when $S_{diff} > 0$, whereas shallow convection and low clouds dominate when $S_{diff} < 0$.

[11] Shallow convective clouds account for most of the 1.46 W m^{-2} increase in TOA radiation within convective regimes (s_{diff} bins) in destabilizing models (Table 1 and auxiliary material). The negative change (−0.21 W m^{-2}) in stabilizing models comes from within deep convective and intermediate regimes (Table 1 and auxiliary material). *Bony and Dufresne* [2005] also showed, using a vertical velocity diagnostic, that the stabilizing influence of shallow convective clouds weakens more so in models with strong positive tropical cloud feedbacks. Vertical velocity and s_{diff} diagnostics yield similar results since large-scale circulations and moist convection are coupled, however the s_{diff} diagnostic attributes a portion of the feedback to changes in the relative frequencies of convective regimes, which we denote the "convective regime" term (Table 1 and auxiliary material). The models generally decrease deep convective and intermediate regime frequencies except for GISS-EH, which increases s_{diff} (not shown). The frequency of shallow convective regimes increases in destabilizing models.

4. Eocene Warm Climate Simulations

[12] Proxy data from Paleocene and Eocene hothouse climates provides abundant evidence that tropical marine temperatures can rise to values far exceeding those encountered in the present climate [*Pearson et al.*, 2007; *Huber*, 2008]. It is of interest to probe the extent to which the behavior revealed in the AR4 simulations extends to sim-

Table 1. Contributions of Terms in Equation (1) to the Total Change in TOA Cloud Radiative Flux after CO_2 Doubling

Contributing Term	All Models (W/m2)	Destabilizing (W/m2)	Stabilizing (W/m2)
Convective regime	−0.51	−0.57	−0.45
Within-regime	0.68	1.46	−0.21
Covariation	0.08	0.08	0.07
Total	0.24	0.97	−0.59

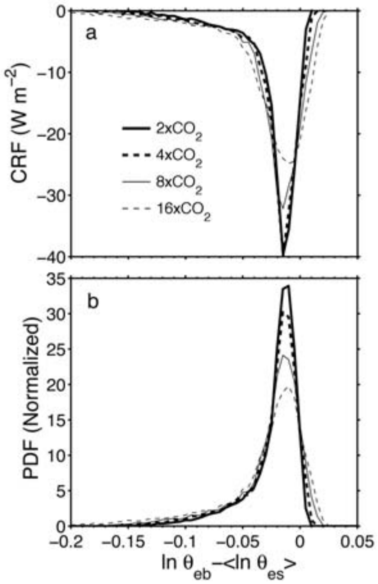

Figure 3. As in Figure 2 but for Eocene simulations, where CO_2 doubles with each simulation ($2 \times CO_2$ to $16 \times CO_2$) relative to the preindustrial level. The maximum annual average SST for each case is 34.0°, 35.7°, 37.6°, and 40.1°C, respectively. The areas under the CRF distribution curves give the net CRF, multiplied by 5.9 to reflect the actual value of CRF at the mode of the s_{diff} distribution for the EO1 simulation.

ulations of yet warmer climates, since our goal is to identify important tropical climate parameters of general applicability. We analyzed a series of Eocene simulations with CO_2 concentrations ranging from twice ($2 \times CO_2$) to sixteen times ($16 \times CO_2$) the pre-industrial level [*Liu et al.*, 2009; *Abbot et al.*, 2009]. These simulations were carried out with boundary conditions, including continental positions, ocean currents, and vegetation appropriate for ~50 Mya with the National Center for Atmospheric Research CCSM3, a heavily validated model used in the IPCC AR4.

[13] In the Eocene simulations, as for the AR4 case, the peak of the entropy-excess histogram (Figure 3) remains nearly fixed as the climate warms, indicating a continued shift in convective threshold to warmer temperatures. However we also see a broadening of the histogram, with more prevalence of relatively high instability. Evidently, the model's convective parameterization is somewhat less effective at eliminating convective available potential energy in very warm climates. Whether the real atmosphere exhibits similar behavior is an unresolved question.

[14] CRF retains a similar character for all the Eocene simulations, with net TOA cloud cooling dominated by convective systems of intermediate stability. In these simulations, the cloud feedback neither strongly amplifies nor attenuates the warming. The possibility remains that clouds in the real Eocene atmosphere may behave differently, but these simulation results suffice to show that cloud properties must be tied to the entropy difference rather than SST, and that a widespread occurrence of high clouds similar to present-tropical high clouds would not have a pronounced cooling effect on the tropics. For that, one either needs to change the character of high clouds so that their albedo effect becomes more dominant, or to increase the net cooling properties of the intermediate clouds that dominate in these simulations.

5. Conclusions

[15] The entropy excess of the sub-cloud layer relative to the free troposphere defines a threshold for convection that is invariant as mean tropospheric temperature increases in future and past (Eocene) greenhouse climate simulations. Shallow and deep convective clouds correspond to negative and positive entropy excess, respectively, making entropy excess a useful coordinate on which to evaluate cloud radiative feedbacks in response to CO_2 doubling. This diagnostic is similar in spirit to that of *Bony et al.* [2004] but does not require evaluation of vertical velocity. Our results reveal small changes in partitioning between deep and shallow convective regimes that could mask the separate contribution of cloud microphysical parameterizations to feedbacks if analyzed using a purely dynamical diagnostic. An important target for long-term satellite cloud monitoring would be to see if the observed changes in tropical cloud radiative forcing are tracking the changes expected from models, as summarized in Figure 2.

[16] Analysis of AR4 and Eocene simulations clearly reveals the fallacy of "trigger-type" tropical thermostats. As climate warms in these simulations, the onset of deep convection and strong cloud feedbacks shifts to warmer SST values, but is approximately invariant in the entropy excess coordinate. Possible inadequacies of the GCMs in representing the real world do not compromise our argument, since the convective threshold is primarily determined by the same fundamental physics that GCMs incorporate as constraints. Hypotheses for the sources of uncertainty in GCM cloud feedbacks are needed, but should not rely on a fixed convective threshold SST independent of changes in the top-of-atmosphere radiation budget.

[17] Deep convection coincides with the peak frequency in the negatively skewed SST distribution because deep convection promotes skewness in the distribution and not because of any intrinsic upper bound on the mean temperature.

[18] **Acknowledgment.** This work was supported by National Science Foundation grants ATM0123999, ATM0933936, and P2C2 program grant ATM0902780.

References

Abbot, D. S., M. Huber, G. Bousquet, and C. C. Walker (2009), High-CO_2 cloud radiative forcing feedback over both land and ocean in a global climate model, *Geophys. Res. Lett.*, *36*, L05702, doi:10.1029/2008GL036703.

Bony, S., and J. L. Dufresne (2005), Marine boundary layer clouds at the heart of tropical cloud feedback uncertainties in climate models, *Geophys. Res. Lett.*, *32*, L20806, doi:10.1029/2005GL023851.

Bony, S., et al. (2004), On dynamic and thermodynamic components of cloud changes, *Clim. Dyn.*, *22*(2–3), 71–86, doi:10.1007/s00382-003-0369-6.

Crowley, T. J. (2000), CLIMAP SSTs re-revisited, *Clim. Dyn.*, *16*(4), 241–255, doi:10.1007/s003820050325.

Dutton, J. F., et al. (2000), The effect of global climate change on the regions of tropical convection in CSM1, *Geophys. Res. Lett.*, *27*(19), 3049–3052, doi:10.1029/2000GL011542.

Fu, R., et al. (1992), Cirrus-cloud thermostat for tropical sea-surface temperatures tested using satellite data, *Nature*, *358*(6385), 394–397, doi:10.1038/358394a0.

Hartmann, D. L., and M. L. Michelsen (1993), Large-scale effects on the regulation of tropical sea-surface temperature, *J. Clim.*, *6*(11), 2049–2062, doi:10.1175/1520-0442(1993)006<2049:LSEOTR>2.0.CO;2.

Hartmann, D. L., and M. L. Michelsen (2002), No evidence for iris, *Bull. Am. Meteorol. Soc.*, *83*(2), 249–254, doi:10.1175/1520-0477(2002)083<0249:NEFI>2.3.CO;2.

Huber, M. (2008), A hotter greenhouse?, *Science*, *321*(5887), 353–354, doi:10.1126/science.1161170.

Kalnay, E., et al. (1996), The NCEP/NCAR 40-year reanalysis project, *Bull Am Meteorol. Soc.*, *77*(3), 437–471, doi:10.1175/1520-0477(1996)077<0437:TNYRP>2.0.CO;2.

Kilbourne, K. H., et al. (2004), A fossil coral perspective on western tropical Pacific climate similar to 350 ka, *Paleoceanography*, *19*, PA1019, doi:10.1029/2003PA000944.

Kleypas, J. A., et al. (2008), Potential role of the ocean thermostat in determining regional differences in coral reef bleaching events, *Geophys. Res. Lett.*, *35*, L03613, doi:10.1029/2007GL032257.

Lindzen, R. S., M. D. Chou, and A. Y. Hou (2001), Does the earth have an adaptive infrared iris?, *Geophys. Res. Lett.*, *82*(3), 417–432.

Liu, Z. H., M. Pagani, D. Zinniker, R. DeConto, M. Huber, H. Brinkhuis, S. R. Shah, R. M. Leckie, and A. Pearson (2009), Global cooling during the Eocene-Oligocene climate transition, *Science*, *323*(5918), 1187–1190, doi:10.1126/science.1166368.

Manabe, S., and R. T. Wetherald (1967), Thermal equilibrium of atmosphere with a given distribution of relative humidity, *J. Atmos. Sci.*, *24*(3), 241–259, doi:10.1175/1520-0469(1967)024<0241:TEOTAW>2.0.CO;2.

Newell, R. E. (1979), Climate and the ocean, *Am. Sci.*, *67*(4), 405–416.

Pearson, P. N., B. E. van Dongen, C. J. Nicholas, R. D. Pancost, S. Schouten, J. M. Singano, and B. S. Wade (2007), Stable warm tropical climate through the Eocene epoch, *Geology*, *35*(3), 211–214, doi:10.1130/G23175A.1.

Peters, M. E., and C. S. Bretherton (2005), A simplified model of the Walker circulation with an interactive ocean mixed layer and cloud-radiative feedbacks, *J. Clim.*, *18*(20), 4216–4234, doi:10.1175/JCLI3534.1.

Pierrehumbert, R. T. (1995), Thermostats, radiator fins, and the local runaway greenhouse, *J. Atmos. Sci.*, *52*(10), 1784–1806, doi:10.1175/1520-0469(1995)052<1784:TRFATL>2.0.CO;2.

Ramanathan, V., and J. A. Coakley (1978), Climate modeling through radiative-convective models, *Rev. Geophys.*, *16*(4), 465–489, doi:10.1029/RG016i004p00465.

Ramanathan, V., and W. Collins (1991), Thermodynamic regulation of ocean warming by cirrus clouds deduced from observations of the 1987 El-Nino, *Nature*, *351*(6321), 27–32, doi:10.1038/351027a0.

Rossow, W. B., and R. A. Schiffer (1999), Advances in understanding clouds from ISCCP, *Geophys. Res. Lett.*, *80*(11), 2261–2287.

Sobel, A. H., and C. S. Bretherton (2000), Modeling tropical precipitation in a single column, *J. Clim.*, *13*(24), 4378–4392, doi:10.1175/1520-0442(2000)013<4378:MTPIAS>2.0.CO;2.

Stephens, G. L. (2005), Cloud feedbacks in the climate system: A critical review, *J. Clim.*, *18*(2), 237–273, doi:10.1175/JCLI-3243.1.

Sud, Y. C., G. K. Walker, Y. P. Zhou, G. A. Schmidt, K.-M. Lau, and R. F. Cahalan (2008), Effects of doubled CO_2 on tropical sea surface temperatures (SSTs) for onset of deep convection and maximum SST: Simulations based inferences, *Geophys. Res. Lett.*, *35*, L12707, doi:10.1029/2008GL033872.

Veron, V. E. N. (2008), Mass extinctions and ocean acidification: Biological constraints on geological dilemmas, *Coral Reefs*, *27*, 459–472, doi:10.1007/s00338-008-0381-8.

Wallace, J. M. (1992), Effect of deep convection on the regulation of tropical sea-surface temperature, *Nature*, *357*(6375), 230–231, doi:10.1038/357230a0.

M. Huber, Department of Earth and Atmospheric Sciences, Purdue University, 1397 Civil Engineering Bldg., West Lafayette, IN 47907, USA.

R. T. Pierrehumbert and I. N. Williams, Department of Geophysical Sciences, University of Chicago, 5734 S. Ellis Ave., Chicago, IL 60637, USA. (inw@uchicago.edu)