Dependency of global mean precipitation on surface temperature

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Received 30 May 2008; revised 16 July 2008; accepted 18 July 2008; published 28 August 2008.

[1] We investigate the causes of temperature dependent changes in global precipitation in contemporary General Circulation Models (GCMs) subjected to a doubling of atmospheric CO_2 concentration. By analyzing the energy budget of the troposphere, we find that changes are dominated by processes robustly simulated by GCMs. Importantly, shortwave cloud feedbacks, whose uncertainty is largely responsible for the wide range of GCM temperature climate sensitivities, are shown to have little effect. This is because these mainly arise from the scattering of shortwave radiation that has little impact on the tropospheric heating that controls precipitation. Hence, we expect that the range of simulated precipitation sensitivities to temperature will not change greatly in future GCMs, despite the recent suggestion that satellite observations indicate that GCM precipitation changes are significantly in error. **Citation:** Lambert, F. H., and M. J. Webb (2008), Dependency of global mean precipitation on surface temperature, *Geophys. Res. Lett.*, *35*, L16706, doi:10.1029/2008GL034838.

1. Introduction

[2] Recently, data obtained from the Special Sensor Microwave/Imager (SSM/I) suggested that real world global precipitation responses to global warming could be significantly larger than predicted by General Circulation Models (GCMs) [*Wentz et al.*, 2007]. Annual mean SSM/I precipitation increases of $6.7 \pm 3.5\%\mathrm{K}^{-1}$ for the 20-year period 1987–2006 [*Lambert et al.*, 2008] seem large compared to 20th and 21st century GCM values of $1-3\%\mathrm{K}^{-1}$ [*Held and Soden*, 2006]. However, *Previdi and Liepert* [2008] demonstrated that GCMs can produce precipitation increases consistent with Wentz et al.'s observations over 20-year periods, because of naturally occurring climate variability and because the global mean surface temperature independent effects of external climate forcings on precipitation vary throughout the 20th century. The remaining question is whether the global mean surface temperature dependent effect of forcing on precipitation, which dominates changes on longer timescales, should be larger in GCMs.

[3] Global precipitation change can be described in terms of the tropospheric energy budget [*Mitchell et al.*, 1987; *Allen and Ingram*, 2002]. Since the heat capacity of the troposphere is negligible on annual and longer timescales, any increase in tropospheric latent heating associated with precipitation must be balanced by increases in the net radiative and sensible heat fluxes that cool the troposphere. Increases in global mean surface temperature, ΔT, are accompanied by increases in tropospheric temperature that approximately maintain a moist adiabatic lapse rate. The net effect is a tropospheric cooling that allows an increase in global precipitation. Hence precipitation increases with global warming ($k_T\Delta T$ below). However, because climate forcings can also affect the tropospheric energy budget independent of their effects through ΔT, there are direct effects individual to each forcing (ΔR_A below). The direct effect of increasing CO_2 concentration reduces the cooling of the troposphere through longwave radiation, reducing GCM precipitation increases by about 25% [*Allen and Ingram*, 2002; *Yang et al.*, 2003]. The direct effect of reductions in absorbing aerosol concentration is not routinely included in contemporary GCM simulations of the 20th century and may be partially responsible for observed precipitation increases during the past 20 years [*Previdi and Liepert*, 2008; *Lambert et al.*, 2008].

[4] Conserving energy, we write

$$L\Delta P - k_T\Delta T - \Delta R_A \simeq 0, \qquad (1)$$

where $L\Delta P$ represents changes in tropospheric latent heating due to changes in global mean precipitation, ΔP, $k_T\Delta T$ represents changes in tropospheric radiative and sensible cooling linked to changes in ΔT, and ΔR_A represents changes in cooling due to forcing and independent of ΔT. In the GCM experiments we will consider, the effects of $k_T\Delta T$ and ΔR_A are easily separable, because the former occur with ΔT change over a period of years, but the latter appear as rapid adjustments to the step change in forcing that we apply.

[5] Assuming that 21st century climate change is dominated by Greenhouse Gases (GHGs), we can expect the $k_T\Delta T$ effect of GHGs to be key to future ΔP change. It may even be dominant, as it is in GCM simulations [*Allen and Ingram*, 2002; *Held and Soden*, 2006]. Hence, here we focus on the ΔT dependent effect on precipitation and hence the primary reasons why GCM precipitation changes are found to be between $1-3\%\mathrm{K}^{-1}$.

2. GCM Data

[6] We examine control and $2xCO_2$ conditions in a range of slab (thermodynamic mixed-layer only ocean) GCMs. We investigate both equilibria and "slab transients", in which the models make the transition to their new equilibria after the sudden doubling of atmospheric CO_2 concentration. We take data from the Quantifying Uncertainty in Model Predictions (QUMP) project [*Murphy et al.*, 2004; *Webb et al.*, 2006] for 135 GCMs based on HadSM3 [*Pope et al.*, 2000], but "physically perturbed" in their atmospheric sub-gridscale parameters in order to explore simulation uncertainty due to parameterization. Parameterized moist air plumes in 28 of these GCMs (QUMPlo-ent)

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Table 1. Number of Available Transients (n), ΔT_{eq}, k_T and Its Components, and ΔR_A[a]

			k_T						
Model	n	ΔT_{eq}	Total	CSR	CloudSW	CloudLW	SCloudLW	Sensible	ΔR_A
QUMP	70	1.4–5.2	2.1 ± 0.8	2.6 ± 0.4	0.23 ± 0.13	−0.84 ± 0.65	−0.73 ± 0.52	0.17 ± 0.44	−1.8 ± 1.4
QUMP lo-ent	15	1.8–26.4	1.5 ± 0.8	2.1 ± 0.7	0.14 ± 0.13	−0.75 ± 0.71	−0.40 ± 0.40	0.07 ± 0.43	−1.3 ± 1.7
CAM3	4	2.3	2.3 ± 0.2	2.4 ± 0.1	0.16 ± 0.04	−0.77 ± 0.11	−0.51 ± 0.14	0.40 ± 0.13	−1.1 ± 0.3
CCCma	1	3.8	2.1 ± 0.5	2.6 ± 0.3	0.08 ± 0.04	−0.72 ± 0.12	−0.43 ± 0.17	0.24 ± 0.31	−2.0 ± 1.5
ECHAM5	0	3.3	—	—	—	—	—	—	—
GFDL	1	3.4	1.7 ± 0.3	2.4 ± 0.2	0.13 ± 0.05	−0.92 ± 0.20	−0.50 ± 0.17	0.26 ± 0.18	−2.1 ± 0.8
GISS E-R	1	2.7	2.8 ± 0.4	—	—	—	—	—	−1.9 ± 0.8
HadGSM1	0	4.5	—	—	—	—	—	—	—
HadSM4	0	3.6	—	—	—	—	—	—	—
MIROC-lo	0	4.0	—	—	—	—	—	—	—
UIUC	0	2.3	—	—	—	—	—	—	—

[a]Components of k_T: Total, net clear-sky radiative cooling (CSR), net cloud SW cooling (CloudSW), net cloud LW cooling (CloudLW), net cloud surface LW cooling (SCloudLW) and net sensible cooling in the GCMs (all in Wm^{-2} K^{-1}), and ΔR_A (Wm^{-2}) from regression. Errors are 5–95% confidence intervals based on the range of best estimates for QUMP and QUMPlo-ent, and the error in the fit for the other models. QUMPlo-ent values do not contribute to the QUMP row.

show low entrainment of ambient air. As a result, these models tend to import more water vapor into the upper troposphere than other GCMs, leading to the very high climate sensitivities found in some of these runs. We retain them, nevertheless, as they will serve to demonstrate a point. The QUMP data do not explore uncertainty introduced by differences in possible GCM formulation, so to partially address this we also take data from NCAR CAM3SOM, ECHAM5, GFDL AM2.1, GISS E-R, HadGSM1, HadSM4, MIROC-lo and UIUC (see acknowledgments). Details of data availability are in Table 1.

3. Results

3.1. Climate Sensitivity

[7] Our models show a broad range of climate sensitivities, ΔT_{eq} (Table 1). As has already been shown comprehensively elsewhere, the majority of this range is due to large uncertainties in cloud radiative feedbacks [*Colman*, 2003; *Soden and Held*, 2006; *Webb et al.*, 2006; *Dufresne and Bony*, 2008]. The uncertainties cause large inter-model differences in shortwave radiative response to forcing and ΔT at the Top Of Atmosphere (TOA) (Figure 1a). Longwave feedbacks, meanwhile, are much more consistent, being dominated by changes in clear-sky fluxes. Crucially for us, these are similar functions of ΔT in each model (Figure 1b). This means that transient and equilibrium states more or less share the same distribution of longwave feedbacks, in contrast to shortwave feedbacks, which show a variety of behaviors.

3.2. Global Precipitation

[8] We calculate $k_T\Delta T$ and ΔR_A from ordinary least squares regression of $L\Delta P$ against ΔT. k_T is the gradient and ΔR_A the y-intercept. Values of k_T are in the range 1.3–3.2 Wm^{-2} K^{-1}, equivalent to 1.4–3.4%K^{-1} (excluding QUMPlo-ent) (Table 1). These are comparable to *Held and Soden's* [2006] range of 1–3%K^{-1}, even though they did not explicitly separate $k_T\Delta T$ and ΔR_A, because ΔR_A values are relatively small.

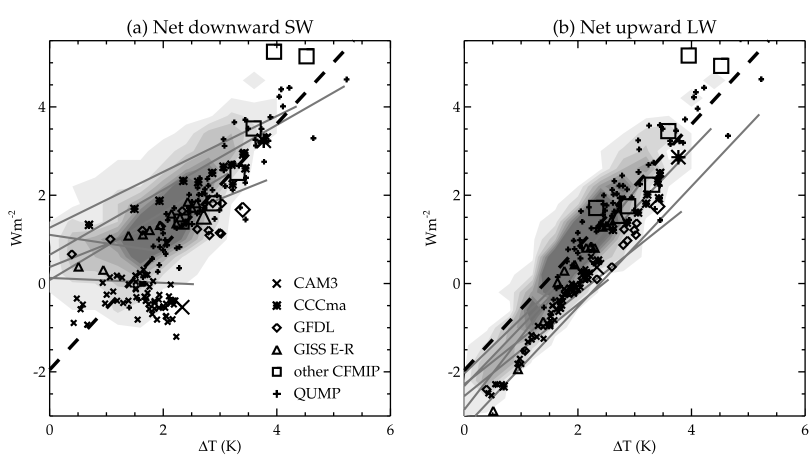

Figure 1. Net (a) incoming shortwave and (b) outgoing longwave TOA fluxes against ΔT. The shaded regions represent the density of QUMP annual-mean transient data, with darker shades meaning more data in a given region. The dashed black line is a fit to QUMP equilibrium values, and is the same in Figures 1a and 1b because the fluxes balance. (QUMPlo-ent data are not included.) The solid grey lines are least squares fits to five arbitrary ensemble members of different ΔT_{eq}. For the other models, annual mean transient data are represented by small symbols; equilibria are represented by large symbols.

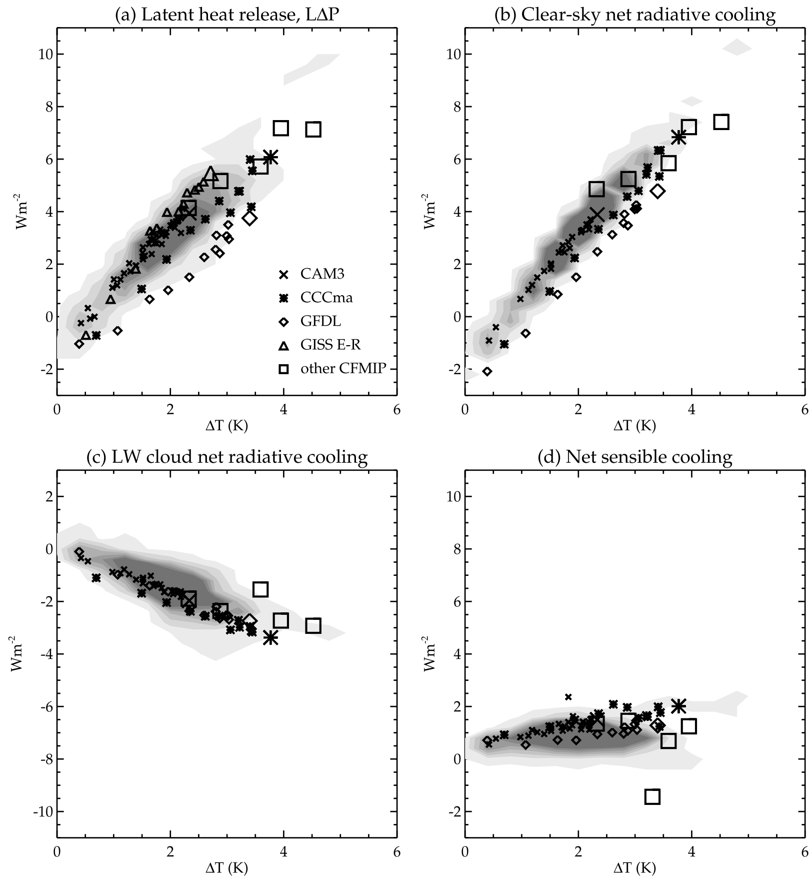

Figure 2. (a) Latent heat release, (b) net clear-sky radiative cooling, (c) LW cloud net radiative cooling (note different vertical scale) and (d) net sensible cooling. The shaded regions represent the density of QUMP annual-mean transient data, with darker shades meaning more data in a given region. QUMP equilibria are omitted for clarity, but conform to the same distribution as transients. (QUMPlo-ent data are not included.) For the other models, annual mean transient data are represented by small symbols; equilibria are represented by large symbols. Positive values lead to increases in tropospheric cooling and precipitation.

[9] We also calculate the clear-sky and cloudy longwave and shortwave components of k_T by regressing radiation components against ΔT (Table 1). We calculate cloud radiation by subtracting clear-sky radiation from all-sky radiation [*Cess et al.*, 1990].

[10] As with TOA longwave feedbacks, tropospheric latent heat release associated with precipitation is a similar function of ΔT across models (Figure 2a). Why is this so, when TOA shortwave radiative feedbacks are so inconsistent? The reason is that only feedbacks that affect tropospheric energy *absorption* affect the tropospheric energy budget [*Ramanathan et al.*, 2001; *Lambert et al.*, 2008]. These are principally changes in clear-sky longwave absorption and emission, and changes in clear-sky shortwave absorption due to changes in tropospheric water vapor concentration. The net effect of these is to increase cooling with ΔT (Figure 2b). Second most important are LW cloud effects that decrease cooling with ΔT (Figure 2c). These are dominated by changes at the surface in QUMP and CAM3, but more evenly distributed between surface and TOA in the other models (Table 1). Why longwave cloud feedbacks are similar functions of ΔT across models we do not know. However, it may be related to the method of calculation (see Discussion).

[11] Of less importance are decreases in sensible heating of the atmosphere that increase tropospheric cooling (Figure 2d), and shortwave cloud and land surface feedbacks (not shown). Shortwave cloud feedbacks are very uncertain at the TOA, but relatively unimportant to the tropospheric energy budget because these mainly affect the scattering of shortwave radiation that largely passes straight through the troposphere. Scattered radiation can only cause precipitation change by changing ΔT: it does not affect the rate of change of precipitation with temperature, k_T. These facts have been known to the aerosol community for many years [*Ramanathan et al.*, 2001; *Ramanathan and Carmichael*, 2008].

[12] The QUMPlo-ent data show lower values of k_T than QUMP mostly because the sensitivity of clear-sky radiative cooling to ΔT change is lower (Table 1). This is believed to be because parameterized convection imports much more water vapor into the upper troposphere compared with the other models.

4. Discussion

[13] The global mean surface temperature, ΔT, dependent effects of climate forcings on precipitation produce

increases of 1.4–3.4%K^{-1}, equivalent to increases in tropospheric latent heating of 1.3–3.2 $Wm^{-2} K^{-1}$ in a range of GCMs that we consider. These results and those of *Held and Soden* [2006] are apparently in disagreement with increases of 6.7 ± 3.5%K^{-1} derived from 1987–2006 observations taken from *Wentz et al.* [2007]. However, as shown by *Previdi and Liepert* [2008], GCMs can produce much larger precipitation increases during a given 20 year period, because the ΔT independent effects of climate forcings and natural variability can be significant over shorter timescales.

[14] Meanwhile, our results and those of *Held and Soden* [2006] are dominated by ΔT dependent effects on precipitation ($k_T\Delta T$). Why is the range of these quite small when the range of temperature climate sensitivity is not? The reason is that temperature dependent effects on precipitation are almost independent of shortwave cloud feedbacks [*Ramanathan et al.*, 2001; *Lambert et al.*, 2008].

[15] Changes in precipitation are constrained by the tropospheric energy budget [*Mitchell et al.*, 1987; *Allen and Ingram*, 2002]. The most important components of this, clear-sky emission and absorption of longwave radiation and absorption of shortwave radiation are relatively well-understood. To first order, what is needed is a knowledge of gas absorption and emission spectra and the maintenance of an approximately moist-adiabatic lapse rate. Granted, there are uncertainties in the lapse-rate and water vapor feedbacks, but these tend to oppose each other [*Colman*, 2003; *Soden and Held*, 2006]. Although probably unphysical, the QUMPlo-ent runs serve to show that a change in clear-sky radiative feedbacks can have a large impact on precipitation change.

[16] Longwave cloud effects are also important, but these show a relatively small range across GCMs. This is puzzling because GCMs show a large spread of sub-tropical low cloud changes, which dominate the TOA net cloud feedback [*Bony and Dufresne*, 2005]. If longwave cloud tropospheric absorption is dominated by these changes, it seems unlikely that it would be robust. An alternative is that there is a "cloud masking" effect, whereby the effects of cloud on intercepting clear-sky radiation are erroneously included in cloud feedback terms calculated by the *Cess et al.* [1990] method that we use. *Soden et al.* [2004] showed that the TOA cloud masking effect is dominated by masking of the clear-sky water vapor feedback. Integrations of the SBDART radiative transfer code (see acknowledgments) suggest that the same may be true at the surface in QUMP, and that cloud masking is sufficient to explain the apparent change in total longwave cloud absorption - at least in the tropics (see auxiliary material).[1] If this is the case, then apparent changes in cloud absorption could be dominated by the climatological distribution of clouds and be robust.

[17] Shortwave cloud feedbacks are responsible for much of the large range of contemporary GCM climate sensitivities. However, these control the scattering of radiation that largely passes through the atmosphere and has little effect on the tropospheric energy budget. A change in the ratio of surface sensible heat flux to surface evaporation (the Bowen ratio) could also affect precipitation. Over much of the Earth's surface where the availability of moisture is not an issue, though, the Bowen ratio is quite well-determined by bulk aerodynamic formulae. Given the relatively small contributions of changes in shortwave cloud feedbacks and sensible heating to the perturbation tropospheric energy budget, only massive differences in future GCM fluxes would have a significant effect on precipitation. In the case of shortwave cloud feedbacks, this would almost certainly lead to large changes in climate sensitivity.

[18] Given the paucity of observations of surface longwave radiation [*Trenberth et al.*, 2007], it is possible that future observations could show that GCM downward surface longwave radiation changes - particularly cloud feedbacks - are significantly in error. This is so, even though there is broad agreement across current GCMs. For example, GCM precipitation changes in the tropics under-predict observed changes by a factor of 3 [*Allan and Soden*, 2007]. The fact that this can be seen in land-based gauge data shows that it is not merely an artifact of satellite observation. Of course, regional precipitation changes may occur without impacting on global tropospheric energy balance if there are compensating changes elsewhere. *Soden* [2000], however, found that the difference between satellite observed and GCM mean tropical precipitation is mirrored by the difference between satellite observed and GCM mean tropical downward surface longwave. There may be consistent errors in the satellite retrieval algorithms. Otherwise, it must be admitted that an error in downward longwave radiation could affect precipitation globally. We note, however, that the GCM runs considered by *Soden* [2000] did not include absorbing aerosols and could not have simulated any ΔT independent effects on precipitation.

[19] **Acknowledgments.** We thank William Ingram and Alexander Stine for excellent discussions and two reviewers for helpful reviews. We acknowledge the international modeling groups for providing data, PCMDI for collection and archiving of CMIP3 and CFMIP data, the JSC/CLIVAR WGCM and their CMIP and Climate Simulation Panel for organization, and the IPCC WG1 TSU for technical support. The IPCC Data Archive at Lawrence Livermore National Laboratory is supported by the Office of Science, U. S. Department of Energy. The CAM3 integrations were performed on the Lightning SMP cluster maintained by NCAR. The Santa Barbara DISORT Atmospheric Radiative Transfer code (SBDART) is kindly provided by Paul Ricchiazzi et al. at the University of California, Santa Barbara (http://www.crseo.ucsb.edu/esrg/sbdart/). FHL was supported by the Joint Defra and MoD Programme, GA01101, CBC/2B/0417_Annex C5.

References

Allan, R. P., and B. J. Soden (2007), Large discrepancy between observed and simulated precipitation trends in the ascending and descending branches of the tropical circulation, *Geophys. Res. Lett.*, *34*, L18705, doi:10.1029/2007GL031460.

Allen, M. R., and W. J. Ingram (2002), Constraints on future changes in climate and the hydrologic cycle, *Nature*, *419*, 224–232.

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.

Cess, R. D., et al. (1990), Intercomparison and interpretation of climate feedback processes in 19 atmospheric general circulation models, *J. Geophys. Res.*, *95*, 16,601–16,615.

Colman, R. (2003), A comparison of climate feedbacks in general circulation models, *Clim. Dyn.*, *20*, 865–873.

Dufresne, J.-L., and S. Bony (2008), An assessment of the primary sources of spread of global warming estimates from coupled atmosphere-ocean models, *J. Clim.*, in press.

Held, I. M., and B. J. Soden (2006), Robust responses of the hydrological cycle to global warming, *J. Clim.*, *19*, 5686–5699.

Lambert, F. H., A. R. Stine, N. Y. Krakauer, and J. C. H. Chiang (2008), How much will precipitation increase with global warming?, *Eos Trans. AGU*, *89*(21), doi:10.1029/2008EO210001.

[1]Auxiliary materials are available in the HTML. doi:10.1029/2008GL034838.

Mitchell, J. F. B., C. A. Wilson, and W. M. Cunnington (1987), On CO_2 climate sensitivity and model dependence of results, *Q. J. R. Meteorol. Soc.*, *113*, 293–322.
Murphy, J. M., D. M. H. Sexton, D. N. Barnett, G. S. Jones, M. J. Webb, M. Collins, and D. J. Stainforth (2004), Quantification of modelling uncertainties in a large ensemble of climate change simulations, *Nature*, *430*, 768–772.
Pope, V. D., M. L. Gallani, P. R. Rowntree, and R. A. Stratton (2000), The impact of new physical parametrizations in the Hadley Centre climate model: HadAM3, *Clim. Dyn.*, *16*, 123–146.
Previdi, M., and B. Liepert (2008), Interdecadal variability of rainfall on a warming planet, *Eos Trans. AGU*, *89*(21), doi:10.1029/2008EO210002.
Ramanathan, V., and G. Carmichael (2008), Global and regional climate changes due to black carbon, *Nat. Geosci.*, *1*, 221–227.
Ramanathan, V., P. J. Crutzen, J. T. Kiehl, and D. Rosenfeld (2001), Aerosols, climate and the hydrological cycle, *Science*, *294*, 2119–2124.
Soden, B. J. (2000), The sensitivity of the tropical hydrological cycle to ENSO, *J. Clim.*, *13*, 538–549.
Soden, B. J., and I. M. Held (2006), An assessment of climate feedbacks in coupled ocean-atmosphere models, *J. Clim.*, *19*, 3354–3360.
Soden, B. J., A. J. Broccoli, and R. S. Hemler (2004), On the use of cloud forcing to estimate cloud feedback, *J. Clim.*, *17*, 3661–3665.
Trenberth, K. E., et al. (2007), Observations: Surface and atmospheric climate change, in *Climate Change 2007: The Physical Science Basis. Contribution of Working Group I to the Fourth Assessment Report of the Intergovernmental Panel on Climate Change*, edited by S. Solomon et al., pp. 235–336, Cambridge Univ. Press, Cambridge, U. K.
Webb, M. J., et al. (2006), On the contribution of local feedback mechanisms to the range of climate sensitivity in two GCM ensembles, *Clim. Dyn.*, *27*, 17–38.
Wentz, F. J., L. Ricciardulli, K. Hilburn, and C. Mears (2007), How much more rain will global warming bring?, *Science*, *317*, 233–235, doi:10.1126/science.1140746.
Yang, F., A. Kumar, M. E. Schlesinger, and W. Wang (2003), Intensity of hydrological cycles in warmer climates, *J. Clim.*, *16*, 2419–2423.

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