Large discrepancy between observed and simulated tropical precipitation trends

Richard P. Allan

Environmental Systems Science Centre, University of Reading, Reading, Berkshire, UK

Brian J. Soden, Viju O. John?

Rosenstiel School of Marine and Atmospheric Sciences, University of Miami, Miami, FL, USA

Observed and model simulated changes in precipitation are examined using reanalysis data to subsample ascending and descending branches of the tropical circulation. An emerging precipitation signal of rising trends in the ascending regions and decreasing trends in the descending regimes, detected in a variety of observational datasets, are substantially larger in magnitude than present-day model simulations and projections into the 21st century. Changes in the observing system can explain only some of this discrepancy which has important implications for future predictions of regional climate change, the reliability of the observing system and the monitoring of the global water cycle.

1. Introduction

Past changes in precipitation are thought to have profoundly affected past human societies [e.g., *Yancheva et al.*, 2007] and projected increases in the total area affected by drought and the flood risk associated with increased frequency of heavy precipitation events are expected to exert an adverse effect on agriculture, water resources, human health and infrastructure [*IPCC*, 2007a, b]. Changes in atmospheric circulation patterns and in thermodynamic properties of the circulation regimes will dictate future regional precipitation changes [e.g., *Emori and Brown*, 2005]; in planning for and adapting to such changes, it is important to be able to predict this response accurately.

There is a robust physical argument for changes in the character of precipitation in a warming world [e.g., *Trenberth et al.*, 2003; *Allen and Ingram*, 2002]. Convection typically draws moisture in from around 3-5 times the radius of the precipitating region [e.g., *Trenberth et al.*, 2003] and observational, modeling and theoretical studies suggest that atmospheric moisture will increase with warming at the rate of approximately 7 $\%K^{-1}$, primarily due to the Clausius Clapeyron relationship between saturated vapor pressure and temperature [*Wentz and Schabel*, 2000; *Soden et al.*, 2005; *Held and Soden*, 2006]. This suggests that precipitation from convective systems will increase at a similar rate [e.g., *Trenberth et al.*, 2003; *Allen and Ingram*, 2002].

Global mean precipitation is however constrained by the energy balance of the atmosphere. Models and observations suggest that atmospheric net radiative cooling (Q) will enhance with planetary warming, primarily due to increased thermal emission of a warmer atmosphere, at a rate of ~ 3 $Wm^{-2}K^{-1}$ [e.g., *Allen and Ingram*, 2002; *Allan*, 2006]. Assuming a negligible change in sensible heat transfer between the surface and the atmosphere, this suggests that precipitation (P) will vary with surface temperature, T_s, according to:

$$\frac{dP}{dTs} \sim \frac{1}{\rho_w L}\frac{dQ}{dTs}, \tag{1}$$

which is approximately 0.1 $mm\ day^{-1}K^{-1}$ or $\sim$4 % K^{-1} (ρ_w is water density and $L = 2.5 \times 10^6\ Jkg^{-1}$). Since this response is smaller than the expected convective region response, this implies that non-convective regions will experience reduced precipitation leading to greater extremes (more intense rainfall and longer dry spells) [e.g., *Emori and Brown*, 2005] and enhanced seasonality [e.g., *Chou et al.*, 2007]. Modeling studies seem to conform to this argument, with global precipitation increasing at just 1-3 $\%K^{-1}$ with evidence of drying in regions of net moisture divergence, in particular for regions at the periphery of convection [e.g., *Neelin et al.*, 2006; *Seager et al.*, 2007; *Meehl et al.*, 2005; *Wang and Lau*, 2006]. However, a recent observational study by *Wentz et al.* [2007] suggests that tropical precipitation increases at a greater rate and in line with the Clausius Clapeyron relationship.

In the present analysis we ask, can an observed precipitation response be detected and how do climate model simulations and predictions correspond with these findings? We monitor observed changes in precipitation within ascending and descending branches of the tropical circulation and compare with current model simulations of the present day and projections of future changes.

2. Data and Method

Monthly mean precipitation from the Global Precipitation Climatology Project [GPCP; *Adler et al.*, 2003] and from the Climate Prediction Center Merged Analysis of Precipitation [CMAP; *Xie and Arkin*, 1998] enhanced product (V703) were employed for the period 1979-2006. Version 6 monthly mean precipitation over the ocean was utilized from the Special Sensor Microwave Imager (SSM/I) for the period 1987-2006 [*Wentz et al.*, 2007]. Based on analysis of the separate satellite records, we use the following satellite combination: F8 (1987-1991), F11 (1992-1999) and F13 (2000-2006). The remaining satellite time-series (F10, F12, F14, F15) produce poorer agreement for overlap periods.

Atmosphere-only model simulations forced with observed sea surface temperature (AMIP3) cover the period 1979-2001; an ensemble of models (CNRM_CM3, GISS_E_R, IAP_FGOALS, INMCM3, IPSL_CM4, MIROC_HIRES, MIROC_MEDRES, MRI_CGCM2, NCAR_CCSM3, NCAR_PCM1, HadGEM1) were extracted from the World Climate Research Programme (WCRP) model archive at the Program for Climate Model Diagnosis and Intercomparison (PCMDI) archive (www-pcmdi.llnl.gov). The AMIP3 model

0094-8276/07/$5.00

fields and the observed values were bi-linearly interpolated to a common 2.5×2.5° latitude-longitude grid.

Coupled model simulations from the Climate of the 20th Century runs (1950-1999) and from the SRESA1B (stabilization at 720ppm CO_2 concentration) scenario (2000-2100) were analyzed using a model ensemble: CCCMA, CNRM_CM3, GFDL_CM2_1, GFDL_CM2_0, GISS_AOM, GISS_E_H, GISS_E_R, IAP_FGOALS, INMCM3, IPSL_CM4, MIROC_MEDRES. These models were setup from control simulations and prescribed natural and anthropogenic forcings [for a description of the model data, see www-pcmdi.llnl.gov; *Held and Soden*, 2006; *Emori and Brown*, 2005; *Meehl et al.*, 2005; *Wang and Lau*, 2006].

Area-weighted averages of the observed and simulated monthly mean fields were calculated for the tropics (30°S-30°N) and for the land-only and ocean-only regions. Additionally, vertical motion at 500 hPa (ω_{500}) was used to subsample regions of mean monthly ascending or descending motion. For the observations, ω_{500} was diagnosed from the National Center for Environmental Prediction/National Center for Atmospheric Research reanalysis 1 [NCEP; *Kalnay et al.*, 1996] for 1979-2006, or from the European Centre for Medium Range Weather Forecasts 40-year reanalysis [ERA40; *Uppala et al.*, 2005] for 1979-2001.

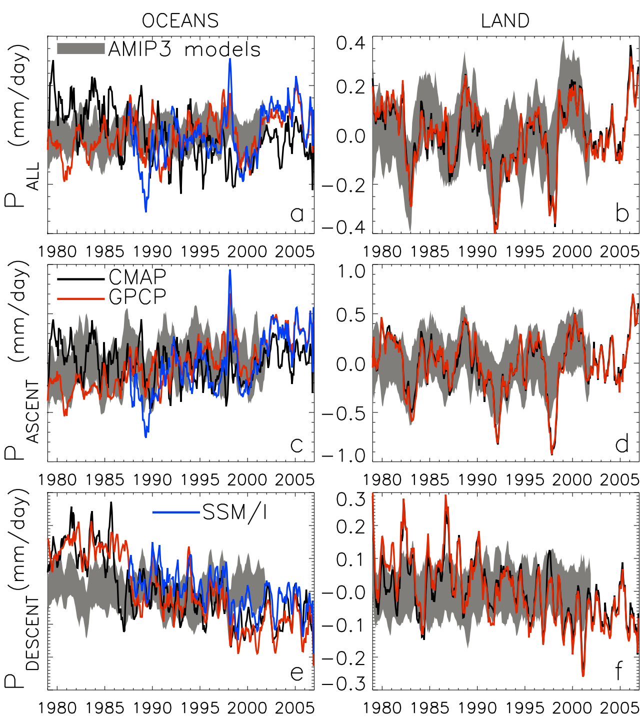

Figure 1. Deseasonalised changes in precipitation ($mm\ day^{-1}$) for observations and AMIP3 models for (a-b) the tropics (30°S-30°N) and regions of mean (c-d) ascent and (e-f) descent over land and ocean. A 5-month moving box-average was applied. Gray shading denotes the model ensemble mean ±1 standard deviation.

3. Interannual Variability

Variability of monthly mean precipitation is displayed in Figure 1. Observed variability is larger than the model envelope with substantial differences between CMAP and GPCP over the ocean (Fig. 1a). Variability over land is consistent (Fig. 1b) and related to the large-scale circulation response to El Niño Southern Oscillation. The ocean-only SSM/I retrieval generally agrees with the GPCP record but there is a negative trend in the CMAP data, reported previously to relate to spurious use of atoll data and changes in the observing system [*Yin et al.*, 2004]. A significant positive trend in the GPCP tropical mean data (0.03 $mm\ day^{-1}dec^{-1}$) is more than double the model ensemble mean trend (Table 1).

Table 1. Linear trends and standard error for deseasonalised monthly mean precipitation 1979-2006.[a] ∗ denotes significant correlation at the 99% level, allowing for autocorrelation.

Dataset	land+ocean	ocean	land
		$dP/dt\ mm\ day^{-1}\ dec^{-1}$	
		Tropics	
GPCP	0.032±0.006*	0.044±0.009*	-0.003±0.013
CMAP	-0.052±0.009*	-0.073±0.012*	0.009±0.013
SSM/I		0.096±0.018*	
models	0.014±0.003*	0.011±0.004	0.024±0.010
		Tropical Ascent	
GPCP	0.184±0.014*	0.248±0.019*	0.022±0.026
CMAP	-0.023±0.020	-0.027±0.025	0.017±0.026
SSM/I		0.353±0.044*	
models	0.059±0.011*	0.0110±0.004	0.0237±0.010
		Tropical Descent	
GPCP	-0.102±0.005*	-0.111±0.006*	-0.080±0.008*
CMAP	-0.080±0.006*	-0.092±0.008*	-0.053±0.008*
SSM/I		-0.062±0.011*	
models	-0.006±0.003	-0.008±0.004	-0.001±0.003

[a] SSM/I, 1987-2006 period; models, 1979-2001 period

Decomposing the tropical variability into ascending and descending regions produces more coherent trends in the data. For the ascending region of the tropical oceans (Fig. 1c), an upward trend in precipitation is evident in both GPCP and CMAP data from 1990-2006 with a larger positive trend for the SSM/I data. For the period 1979-1987, a large discrepancy between CMAP and GPCP data remains for the ocean region which affects the overall trends calculated in Table 1. An upward precipitation trend of 0.18 $mm\ day^{-1}dec^{-1}$ in the GPCP data is substantially larger than the trend calculated for the entire tropics and a factor of 3 larger than the model ensemble mean trend.

For the descending portions of the tropical circulation (Fig. 1e-f) a coherent negative trend in observed precipitation is evident from all datasets over land and ocean (ranging from -0.05 to -0.11 $mm\ day^{-1}dec^{-1}$), but not detectable in the model simulations (Table 1).

4. Sensitivity to Observing System

It is clear that the observed trends in precipitation are larger than the model simulations, in particular for the descending regions. It is important to assess whether this discrepancy may be explained by spurious changes in the reanalysis ω_{500} fields or in the satellite observing systems. Figure 2a-b shows that for the descending regime, a negative precipitation trend is present for GPCP and CMAP when using either NCEP or ERA40 ω_{500}, although the trend is stronger when using ERA40 fields (blue lines). Figure 2a-b also shows the changes in precipitation calculated when applying the changing NCEP ω_{500} to a long-term monthly climatology of precipitation. Any detectable trend would relates to changes in the reanalyses fields rather than the precipitation estimates. A negative trend of around 0.025 $mm\ day^{-1}decade^{-1}$ is calculated, less than 30% of the observational trends calculated in Table 1, although this is not significant at the 95% confidence level for the GPCP climatology. This suggests that the observed changes in P are sensitive to the reanalysis fields but that this cannot explain most of the precipitation responses found in the descending region.

Fig. 2c displays increasing descent-region ω_{500} (stronger descent) although with greater variability in ERA40 than NCEP. It is possible that these changes are artifacts of the observing system [*Held and Soden*, 2006] and these may contribute to a portion of the observed trends. Changes in the areal extent of the descending regime (Fig. 2d) shows coherent variability in NCEP and ERA40 but no visible trend.

We also examined the sensitivity of the observed trends to the products used. Using the GPCP guage-only product (pg1-pg2) over land and the multi-satellite product (pms) over oceans produced trends within the statistical uncertainty of the standard GPCP product. The standard CMAP product (V705) produced trends that were within the statistical uncertainty of the enhanced product. Finally, using a simple merged SSM/I product, averaging all the available satellites, reduced the positive trend over ascending regions by a half while enhanced the negative trend over the descending region by around a third

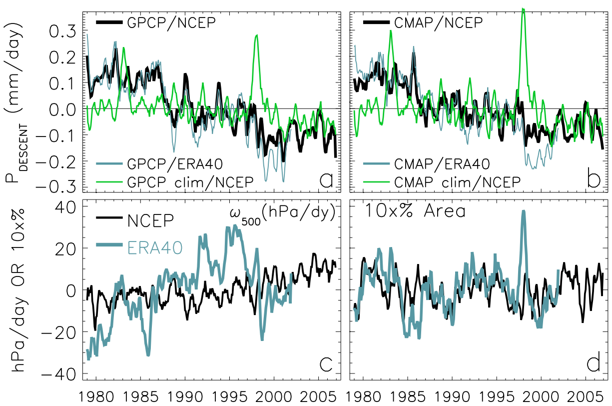

Figure 2. Deseasonalised changes in precipitation (P) over descent regions using NCEP vertical motion (ω_{500}) for (a) GPCP, (b) CMAP and (c) changes in ω_{500} and (d) areal extent of the descending region. Also shown in a-b are the changes in P when ERA40 ω_{500} is applied (light blue) and where long-term climatologies of P are combined with the changing NCEP ω_{500} (green). A 5-month moving box-average was applied.

5. Long term projections

Having established the robust nature of observed precipitation trends, we now place the variation in the context of the longer term CMIP3 climate model projections. Figure 3 shows the tropical mean percentage changes in P, relative to the reference period 1980-1999. The upward response of tropical precipitation in the models becomes increasingly clear in the 21st century; increases for the ascent region are about two thirds larger (Table 2) while for the descent region precipitation response is of variable sign but predominately reducing with time.

The disagreement between GPCP and CMAP precipitation changes for the tropics originate from the ocean ascent region before 1998 where CMAP displays a negative trend. Over the period 1979-2006 GPCP displays a positive trend of 3.7 % dec^{-1}, approximately 7 times larger than the corresponding model ensemble mean trend (note that this is larger than the difference in Table 1 since the model ensemble mean ascent region P is larger than for GPCP). Over the descending region, CMAP and GPCP estimates correspond well, showing a large negative trend of around -7 to -9 % dec^{-1}. No trend is detected in the model data for the descending regimes over this period although a negative trend of just 0.16 % dec^{-1} is calculated for the 1950-2100 period.

Table 2. Linear trends and standard error for bi-annual mean precipitation. ∗correlation above 95% significance level.

Tropics		dP/dt (% $decade^{-1}$)		
Dataset	Period	All	Ascent	Descent
GPCP	1979-2006	1.07±0.45	3.66±0.48*	-9.13±0.98*
CMAP	1979-2006	-1.55±0.48*	-0.40±0.60	-6.63±0.78*
models	1979-2006	0.30±0.08*	0.51±0.16*	0.00±0.11
models	1950-2100	0.35±0.01*	0.58±0.02*	-0.16±0.01*

6. Discussion

There is a large discrepancy between the observed and simulated precipitation changes over the tropics. A negative trend in CMAP data over the ascending-ocean region of the tropics before 1998 is thought to be spurious [e.g., *Yin et al.*, 2004]; for the remaining comparisons, robust upward trends in the ascending regime and downward trends in the descending portions of the tropical circulation are found in GPCP, CMAP and SSM/I datasets. For the tropics, the GPCP trend is about 2-3 times larger than the model ensemble mean trend, consistent with previous findings [*Wentz et al.*, 2007]. Are these changes plausible?

Assuming an approximate warming trend of 0.2 $K\ dec^{-1}$ [e.g., *Wentz et al.*, 2007], the GPCP trend for tropical ascent calculated in Table 2 approximates to about $\sim$18%K^{-1}, substantially larger than predicted by Clausius Clapeyron, and suggests increases in precipitation extremes substantially larger than predicted by models. This also implies enhanced moisture transport that is at odds with observations of a weakening tropical circulation [*Vecchi et al.*, 2006; *Wentz et al.*, 2007]. Considering the observed relationship between moisture and temperature of $\sim$4 $mm\ K^{-1}$ for ascending regions [*Allan*, 2006] and an ascending region column water vapor amount of 45 mm, the percentage increases in P are approximately double that of the water vapor changes.

One obvious explanation is that changes in the observing system and calibration of satellite data introduce spurious variations. It is estimated that changes in the reanalysis vertical motion fields, that are potentially spurious [*Held and Soden*, 2006], may explain up to one third of the trends in the descending region. Nevertheless, agreement between multiple datasets over this regime suggest the remaining changes are from the precipitation datasets. Over ocean descent regions, the observations are dependent primarily on satellite data and it is possible that errors in satellite calibration can explain these trends. However, a downward trend is also detectable over land-descent regions where rain-guage data is utilized, suggesting that these trends are robust.

An increase in global mean precipitation with temperature of around 6%K^{-1} [*Wentz et al.*, 2007] requires an increase in the atmospheric net radiative cooling that is larger than expected [*Allen and Ingram*, 2002]. It is possible that the models do not capture decadal variability in precipitation and radiative cooling adequately, possibly relating to changes in cloud and aerosol radiative effects [e.g., *Wielicki et al.*, 2002; *Mishchenko et al.*, 2007]. Continued monitoring of tropical precipitation and further improvements in satellite calibration are required to explain the large discrepancy between observed and model predicted responses of the atmospheric hydrological cycle to warming.

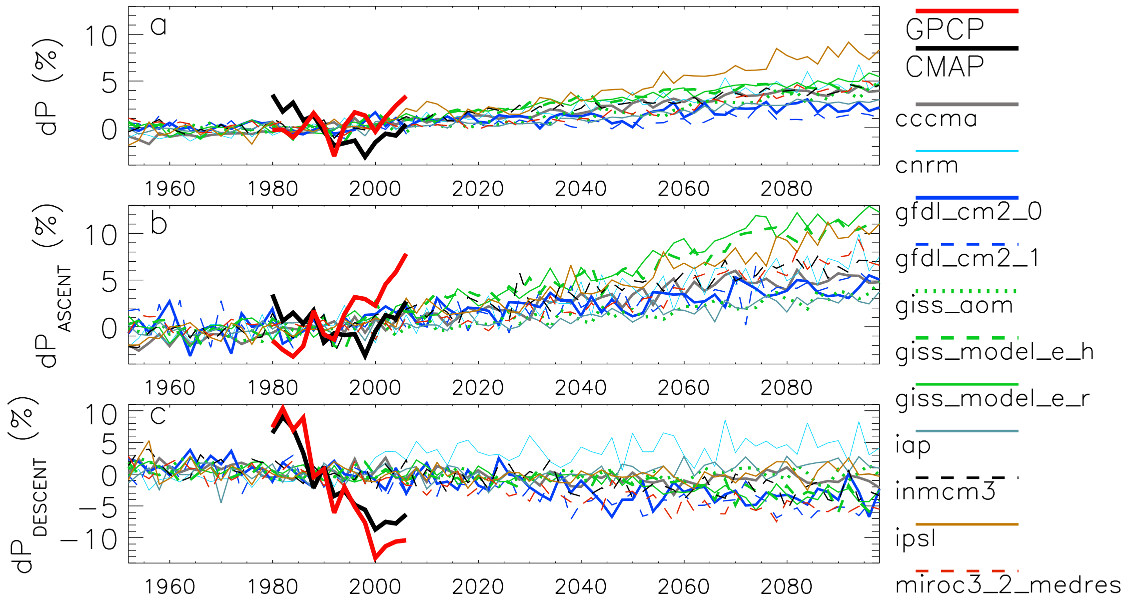

Figure 3. Percentage changes in tropical precipitation, relative to 1980-1999, simulated by CMIP3 models over the period 1950-2100 for (a) all regions, (b) ascending regions and (c) descending regions. Also shown are CMAP and GPCP observed estimates. A 2-year average is applied to the model and observational data.

Acknowledgments. Thanks to the WCRP for enabling the PCMDI model archive (www-pcmdi.llnl.gov). R.Allan was supported by NERC grant NE/C51785X/1. SSM/I data were retrieved from www.ssmi.com, GPCP data from www.ncdc.noaa.gov, and CMAP data from www.cdc.noaa.gov

References

Adler, R. F., et al. (2003), The version-2 Global Precipitation Climatology Project (GPCP) monthly precipitation analysis (1979-present), *J. Hydrometeor.*, *4*, 1147–1167.

Allan, R. P. (2006), Variability in clear-sky longwave radiative cooling of the atmosphere, *J. Geophys. Res.*, *111*(D22105), doi:10.1029/2006JD007304.

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

Chou, C., J. Tu, and P. Tan (2007), Asymmetry of tropical precipitation change under global warming, *Geophys. Res. Lett.*, *submitted*.

Emori, S., and S. J. Brown (2005), Dynamic and thermodynamic changes in mean and extreme precipitation under changed climate, *Geophys. Res. Lett.*, *32*(L17706), doi:10.1029/2005GL023,272.

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

IPCC (2007a), *Climate Change 2007: The Physical Science Basis. Contribution of Working Group I to the Fourth Assessment Report of the Intergovernmental Panel on Climate Change*, 996 pp., Cambridge University Press, Cambridge, United Kingdom and New York, NY, USA.

IPCC (2007b), *Climate Change 2007: Impacts, Adaptation and Vulnerability. Working Group II Contribution to the Intergovernmental Panel on Climate Change Fourth Assessment Report, Summary for Policy Makers*, 23 pp.

Kalnay, E., et al. (1996), The NCEP/NCAR 40-year reanalysis project, *Bull. Amer. Met. Soc.*, *77*, 437–471.

Meehl, G. A., J. M. Arblaster, and C. Tebaldi (2005), Understanding future patterns of increased precipitation intensity in climate model simulations, *Geophys. Res. Lett.*, *32*(L18719), doi:10.1029/2005GL023,680.

Mishchenko, M. I., I. V. Geogdzhayev, W. B. Rossow, B. Cairns, B. E. Carlson, A. A. Lacis, L. Liu, and L. D. Travis (2007), Long-term satellite record reveals likely recent aerosol trend, *Science*, *315*(5818), 1543.

Neelin, J. D., M. Munnich, H. Su, J. E. Meyerson, and C. E. Holloway (2006), Tropical drying trends in global warming models and observations, *PNAS*, *103*(16), 6110–6115.

Seager, R., et al. (2007), Model Projections of an Imminent Transition to a More Arid Climate in Southwestern North America, *Science*, *316*(5828), 1181–1184.

Soden, B. J., D. L. Jackson, V. Ramaswamy, M. D. Schwarzkopf, and X. Huang (2005), The radiative signature of upper tropospheric moistening, *Science*, *310*, 841–844.

Trenberth, K. E., A. Dai, R. M. Rasmussen, and D. B. Parsons (2003), The changing character of precipitation, *Bull. Amer. Met. Soc.*, *84*, 1205–1217.

Uppala, S. M., et al. (2005), The ERA-40 re-analysis, *Quart. J. Roy. Meteorol. Soc.*, *131*, 2961–3012.

Vecchi, G. A., B. J. Soden, A. T. Wittenberg, I. M. Held, A. Leetmaa, and M. J. Harrison (2006), Weakening of tropical pacific atmospheric circulation due to anthropogenic forcing, *Nature*, *441*, 73–76.

Wang, H., and K. M. Lau (2006), Atmopsheric hydrological cycle in the tropics in twentieth century coupled climate simulations, *Int. J. Climatol.*, *26*, 655–678.

Wentz, F. J., and M. Schabel (2000), Precise climate monitoring using complementary satellite data sets, *Nature*, *403*, 414–416.

Wentz, F. J., L. Ricciardulli, K. Hilburn, and C. Mears (2007), How much more rain will global warming bring?, *Science*, p. doi:10.1126/science.1140746.

Wielicki, B. A., et al. (2002), Evidence for large decadal variability in the tropical mean radiative energy budget, *Science*, *295*, 841–844.

Xie, P., and P. A. Arkin (1998), Global monthly precipitation estimates from satellite-observed outgoing longwave radiation, *J. Climate*, *11*, 137–164.

Yancheva, G., et al. (2007), Influence of the intertropical convergence zone on the east asian monsoon, *Nature*, *445*, 74–77.

Yin, X., A. Gruber, and P. Arkin (2004), Comparison of the GPCP and CMAP merged guage-satellite monthly precipitation products for the period 1979-2001, *J. Hydromet.*, *5*, 1207–1222.

R. P. Allan, Environmental Systems Science Centre, University of Reading, Harry Pitt Building, Whiteknights, Reading, Berkshire, RG6 6AL, UK. (rpa@mail.nerc-essc.ac.uk)

B. J. Soden, Rosenstiel School of Marine and Atmospheric Sciences, University of Miami, Rikenbacker Causeway, Miami, FL, USA. (bsoden@rsmas.miami.edu)