GEOPHYSICAL RESEARCH LETTERS, VOL. 33, L16805, doi:10.1029/2006GL026547, 2006

# Observations of tropical precipitating clouds ranging from shallow to deep convective systems

Hirohiko Masunaga[1] and Christian D. Kummerow[1]

Received 10 April 2006; revised 12 June 2006; accepted 13 July 2006; published 22 August 2006.

[1] Regional and temporal variability in a broad spectrum of tropical precipitation systems is investigated in conjunction with the large-scale environment. The analysis utilizes four storm categories (Shallow, Cumulus Congestus, Deep Stratiform, and Deep Convective) determined from Tropical Rainfall Measuring Mission (TRMM) measurements. Deep Stratiform and Deep Convective systems are found to be clearly correlated with large-scale circulation deduced from a reanalysis data set, and are modulated by a distinct seasonal cycle over land. The Shallow category is practically the only component of tropical oceanic rainfall for cold sea surfaces, while it gives way to deeper systems as SST exceeds 28–29°C. The cloud horizontal scale of organized rainfall systems tends to be increasingly extensive relative to the raining portion as the system becomes larger. The present results are discussed in light of existing relevant studies. **Citation:** Masunaga, H., and C. D. Kummerow (2006), Observations of tropical precipitating clouds ranging from shallow to deep convective systems, *Geophys. Res. Lett.*, *33*, L16805, doi:10.1029/2006GL026547.

## 1. Introduction

[2] Precipitating cloud systems are a critical geophysical element involved in tropical large-scale circulation. A popular observational tool to investigate tropical convection on a planetary scale is satellite-measured outgoing longwave radiation (OLR). Satellite OLR serves as a useful proxy of deep convection, while shallow cumulus clouds are not detectable by OLR. Less attention has been paid to shallow rainfall than to deep convection because precipitation and diabatic heating ascribed to shallow cumulus are so limited that shallow clouds per se have a smaller hydrological and dynamical impact to the climate system than deep convection. Shallow cumulus, however, is a dominant member of tropical convection in number [*Johnson et al.*, 1999; *Short and Nakamura*, 2000; *Lau and Wu*, 2003] and provides a significant fraction of rainfall under the trade inversion [*Schumacher*, 2006].

[3] *Masunaga et al.* [2005, hereinafter referred to as MLK] analyzed Tropical Rainfall Measuring Mission (TRMM) data to investigate a wide spectrum of tropical precipitation systems ranging from shallow cumulus to deep convection. MLK used TRMM Precipitation Radar (PR) echo-top height and Visible/Infrared Scanner 10.8-μm brightness temperature (VIRS IR $T_b$) to classify detected rainfall events into the four storm categories of Shallow, Cumulus Congestus, Deep Stratiform, and Deep Convective. The analysis of MLK focused on testing the proposed methodology and was thus limited to two months (Februaries of 1998 and 2000) and to three Pacific domains as well as two tropical continents. In the present paper, the MLK's approach is extended to a longer time period with additional oceanic regions in order to explore more robust climatological signals recorded in the data. The data sets and analysis methods are described in section 2. The results are presented in section 3 and summarized in section 4, where implications of the present findings are discussed as well.

## 2. Data and Methodology

[4] The PR 2A23 and VIRS 1B01 products are employed for PR echo-top height (or "storm height" in the official terminology) and VIRS IR $T_b$, respectively, from the TRMM operational data archives. The TRMM satellite was boosted from the original altitude of 350 km up to 402.5 km in August 2001. The present analysis is applied to the pre-boost period so as to avoid the bias that might arise from a change in the PR sensitivity due to the orbital boost. Consequently, the time period to be investigated consists of 43 months from January 1998 to July 2001. The same period from the monthly mean data set of the National Centers for Environmental Prediction (NCEP)/National Center for Atmospheric Research (NCAR) Reanalysis data archives [*Kalnay et al.*, 1996] is used for pressure vertical velocity or $\omega$ at 500 hPa as an indicator of large-scale ascent and subsidence. The daily data set of sea surface temperature (SST) derived from TRMM Microwave Imager (TMI) measurements is provided by the Remote Sensing Systems (RSS). SSTs at four adjacent quarter-degree grid points from both the ascending and descending orbits each day are combined to mitigate the deficiency of SST data in raining scenes.

[5] Following MLK, storm categories are defined by PR echo-top height and VIRS IR $T_b$ as illustrated in Figure 1. The Shallow category consists of cumulus clouds trapped in the boundary layer that produce moderate, warm rainfall. Cumulus Congestus events develop deeper than Shallow events but not sufficiently to reach the tropopause. Both the Deep Stratiform and Deep Convective categories have cloud tops as high as the tropopause and are differentiated from each other by PR echo-top height. Deep Stratiform events do not have an appreciable amount of frozen hydrometeors large enough to be detectable by the PR, while large ice particles are present much higher than the freezing level for Deep Convective events. Such large frozen particles are presumably generated in deep convective updraft, and then precipitate across the surrounding stratiform sec-

---

[1] Department of Atmospheric Science, Colorado State University, Fort Collins, Colorado, USA.

0094-8276/06/2006GL026547$05.00

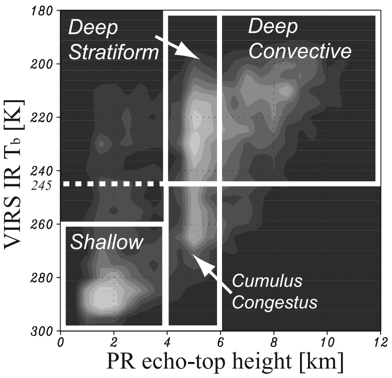

**Figure 1.** The storm categories, defined based on the joint probability density function (shaded) of PR echo-top height and VIRS IR $T_b$.

tion. The current terminology is parallel to the conventional convective/stratiform classification to the extent that a horizontal gradient in PR echo-top height, under the constraint of cold (<245 K in $T_b$) cloud tops, reflects the contrast between deep convective cores and stratiform precipitation areas.

[6] Cloud correlation scale length (CSL) and rain CSL are defined as a statistical measure of the horizontal organization of precipitating cloud systems. The cloud (rain) CSL represents how quickly cloud-top height (raining area coverage) decays with distance as deduced from a large number of observations. In contrast to the direct estimate of cloudy/raining areas by counting contiguous satellite pixels [e.g., *Nesbitt et al.*, 2000], the CSL is inapplicable to instantaneous snapshots of individual storm systems but has the advantage that the CSL is not very susceptible to the truncation by satellite swaths. The reader is referred to MLK for further details.

[7] Geographical regions investigated in this study consist of five oceanic domains and two continents: west Pacific and Australasian seas (105°E–150°E, hereafter denoted simply as the "west Pacific" for brevity), central Pacific (180°–150°W), east Pacific (120°W–90°W), Indian Ocean (60°E–90°E), Atlantic Ocean (45°W–15°W), South America, and Africa. All the regions are latitudinally bound by 15°S and 15°N. This meridional band has been expanded from that chosen by MLK (10°S–10°N) so that the intertropical convergence zone (ITCZ) remains included in the analysis during solstice seasons in which the ITCZ migrates off the equator. Landmasses and islands encompassed in the selected oceanic regions, including the Maritime Continent, are all masked out.

## 3. Results

[8] The time series of each storm category is shown in Figure 2 in terms of the monthly frequency of occurrence. The Deep Stratiform and Deep Convective categories dominate the Shallow category in the west Pacific except for the first 5 months in 1998, during which the 1997–98 El Niño episode was still active. The overall trend is inverted in the central Pacific, where the Shallow category is dominant over others once the El Niño ceases. The Cumulus Congestus category closely resembles the Shallow category in the west Pacific, but stays constantly less frequent than the Shallow category in the central Pacific. The east Pacific is similar to the central Pacific in terms of the average population of each storm category, whereas a noticeable annual cycle, which is absent in the west and central Pacific, manifests itself in the east Pacific: deep systems overwhelm Shallow events from April through October. This seasonal cycle might be related to the enhanced activity of the intraseasonal oscillation in the east Pacific during boreal summer [*Knutson and Weickmann*, 1987; *Maloney and Hartmann*, 2000]. Deep Stratiform and Deep Convective systems are more frequent in the Indian Ocean than in the central and east Pacific, although not as much as in the west Pacific. None of the members of the storm categories are active in the Atlantic Ocean, where only Shallow events exhibit a modest seasonal cycle that peaks in July. Atlantic Shallow events are found suppressed near the African coast in their least frequent phase of the annual cycle (not shown).

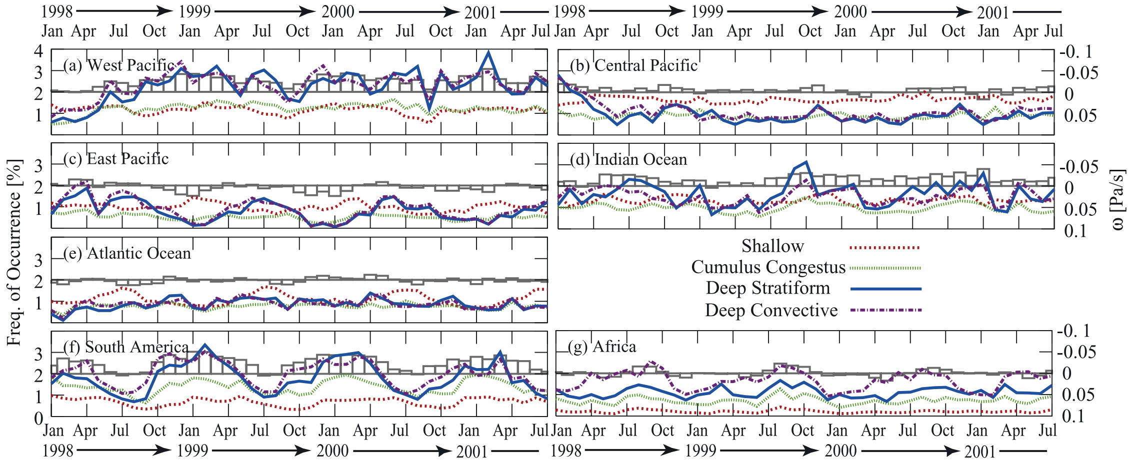

**Figure 2.** The frequency of occurrence of Shallow (red), Cumulus Congestus (green), Deep Stratiform (blue), and Deep Convective (purple) events as well as ω at 500 hPa (histogram) in [Pa/s] labeled on the right-hand axis. Negative ω, indicating large-scale ascent, is defined upward. Each variable is a regionally averaged, monthly mean value shown in a time series from January 1998 to July 2001. Different panels are for different regions as indicated.

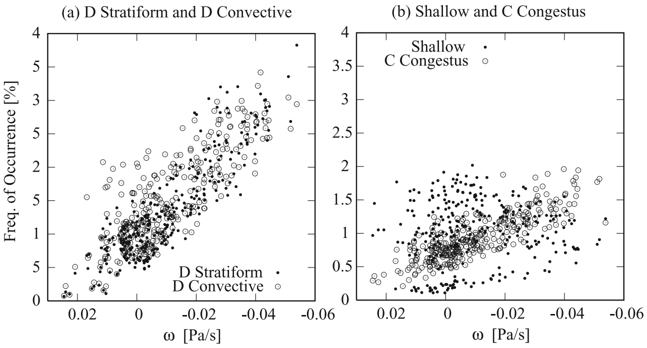

**Figure 3.** (a) Deep Stratiform and Deep Convective and (b) Shallow and Cumulus Congestus plotted in the frequency of occurrence versus ω plane. A single plot represents a regionally averaged, monthly mean value. Negative ω, indicating large-scale ascent, is defined rightward.

*Matsui et al.* [2004] showed the observational evidence that low clouds are typically not raining in this region, where aerosol concentration and lower-tropospheric stability are relatively high. The annual cycle is prominent over the South American and African continents especially for deep systems, as expected from the monsoonal changes of climate regime in these regions. The Deep Convective category prevails over others in Africa.

[9] The temporal variation of the Deep Stratiform and Deep Convective categories is generally well synchronized with the 500-hPa ω velocity (histogram in Figure 2), suggesting a close connection between the development of deep precipitation systems and large-scale circulation in all study areas. Figure 3a confirms that a higher (lower) population of Deep Stratiform and Deep Convective systems is well correlated with the large-scale ascent (subsidence). A similar result was obtained by *Del Genio and Kovari* [2002]. This correlation is still clear for Cumulus Congestus events although the sensitivity to ω is weaker than deep systems. No appreciable relationship with ω is found for the Shallow category (Figure 3b).

[10] The frequency of rainfall occurrence is further broken down by SST for the oceanic domains. Figures 4a–4d shows the percent occurrence computed individually for different SSTs. When SST is lower than 27°C, Shallow events are practically the only component of tropical precipitation. Deep Convective systems, however, sharply increase with SST for warmer sea surfaces, leading the Deep Convective category to overwhelm other storm categories once SST exceeds 28–29°C. The Deep Stratiform category also increases with SST although not as markedly as the Deep Convective. The diverse oceanic regions delineate a tightly bundled group of curves, topped by the Atlantic Ocean and east Pacific and bottomed by the west Pacific and Indian Ocean for Deep Stratiform and Deep Convective systems. Interestingly, this regional gradient is seemingly contrary to what was shown in Figure 2. A clue to reconcile the apparent discrepancy lies in the SST histogram (Figure 4e). SSTs higher than 29°C occur much more often in the Indian Ocean and west Pacific than in the Atlantic Ocean and east Pacific. The high population of deep systems in the west Pacific and Indian Ocean is therefore largely owed to the frequent occurrence of warm sea surface, and the opposite argument applies to the Atlantic Ocean and east Pacific.

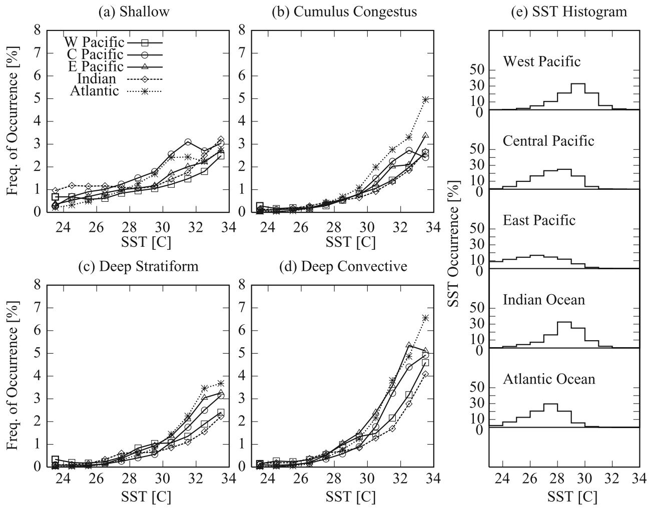

**Figure 4.** The frequency of occurrence partitioned by SST for (a) Shallow, (b) Cumulus Congestus, (c) Deep Stratiform, and (d) Deep Convective, separated by different oceanic regions as indicated Figure 4a. (e) The SST histogram.

[11] The CSL estimated for Deep Stratiform and Deep Convective events is investigated next to examine the horizontal organization of precipitating clouds. The Shallow and Cumulus Congestus categories are not considered here because they are usually not accompanied with substantially organized systems [see MLK]. A slight increase in the cloud CSL is barely recognizable as the frequency of occurrence increases (Figure 5a), where the correlation coefficient is 0.236 for the Deep Stratiform category and 0.416 for the Deep Convective. Although these correlations are statistically significant compared to the correlation coefficient of ±0.148 at the 99% significance level, the tendency is so subtle that it is not necessarily concluded that a large-scale environment favorable to the development of deep systems promotes their horizontal organization, and vice versa. On the other hand, the cloud CSL and rain CSL are closely linked with each other (Figure 5b). The cloud CSL is about

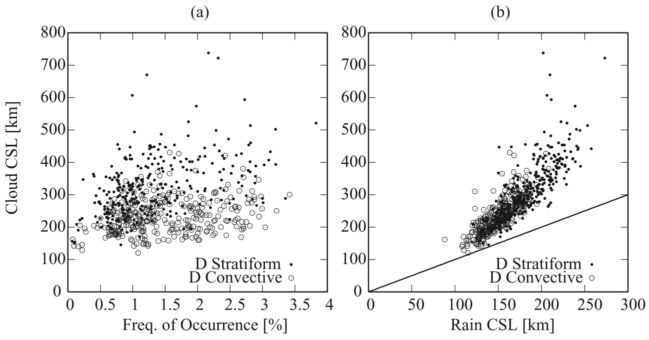

**Figure 5.** Scatter plots for (a) the cloud CSL versus the frequency of occurrence and (b) the cloud CSL versus the rain CSL. A single plot represents a regionally averaged, monthly mean value. The cloud CSL exceeds the rain CSL above the solid line given in Figure 5b.

the same in magnitude as the rain CSL at their bottom end of ∼100 km, while the cloud CSL tends to be 1.5 times or nearly twice as large where the rain CSL exceeds 200 km. This result suggests that cirrus anvil clouds detrained from deep convection could gain increasingly large spatial extent compared to the size of the raining portion itself as the system is organized on a greater horizontal scale.

## 4. Discussion and Conclusions

[12] The present work seeks to understand the variability in a broad spectrum of tropical precipitation systems as a function of the large-scale environment. The analysis method follows that proposed by MLK, introducing four storm categories (Shallow, Cumulus Congestus, Deep Stratiform, and Deep Convective) determined by the combination of TRMM PR echo-top height and VIRS IR $T_b$. Deep systems are clearly correlated with large-scale circulation, and are modulated by a distinct seasonal cycle over land. Shallow cumulus is virtually disconnected from large-scale dynamics. Shallow systems are therefore a main provider of tropical oceanic rainfall in the presence of large-scale subsidence, which suppresses deep convection, outside the deep Tropics as observed by *Schumacher and Houze* [2003].

[13] The dependence of raining cloud populations on SST are qualitatively consistent across diverse oceanic regions in the Tropics. Shallow events are ubiquitous over the entire SST range of tropical oceans, growing slightly in population as SST increases. Deeper systems do not develop at all for cold SSTs (<∼27°C) but increase rapidly over warmer sea surfaces. As a result, the dominant mode of tropical rainfall switches from shallow cumulus to deep convective systems around an SST threshold of 28–29°C, as also implied by preceding observational work [*Waliser and Graham*, 1993; *Del Genio and Kovari*, 2002; MLK]. The high sensitivity to SST for deep systems over warm ocean could be attributed to local thermodynamic conditions nonlinearly changing with underlying SST [*Sobel and Bretherton*, 2000].

[14] Shallow systems are more likely dependent on local thermodynamic conditions rather than large-scale dynamics. The thermodynamic role of shallow precipitation in tropical atmospheres has been discussed in the literature. Increasing SST would gradually expedite the precipitation efficiency of shallow cumulus clouds [*Lau and Wu*, 2003; *Rapp et al.*, 2005], resulting in free-tropospheric moistening that preconditions the development of deep convection [*Kemball-Cook and Weare*, 2001]. The eventual deep convection is followed by stratiform rain that peaks at a later time as observed at various spatial and temporal scales such as mesoscale squall lines [*Houze*, 1977], cloud clusters [*Leary and Houze*, 1979], the quasi-2-day oscillation [*Takayabu et al.*, 1996], and the Madden-Julian Oscillation [*Lin et al.*, 2004; *Kiladis et al.*, 2005]. The Deep Stratiform category appears weaker in its sensitivity to SST than the Deep Convective (Figure 4). This result could arise in part from a temporal lag of stratiform rain occurrence behind deep convection which may have already begun to modify the underlying SST. One might argue that it could be also related to the SST-dependent precipitation efficiency of deep convective clouds as discussed by *Lindzen et al.* [2001] in their adaptive iris hypothesis. An extensive examination of the iris hypothesis based on TRMM measurements is given by *Rapp et al.* [2005].

[15] Variation in the population of Deep Stratiform and Deep Convective systems shows only a subtle systematic tendency with the horizontal organization of the storms as measured by the CSL. The organization of precipitation systems inferably involve dynamic processes that are not describable by large-scale environment alone, such as mesoscale momentum transport [*Houze et al.*, 2000]. The cloud and rain CSLs are highly correlated with each other, with the cloud CSL growing more rapidly than the rain CSL above a common lower end of ∼100 km. This finding has two intriguing implications. First, so-called super cloud clusters identified from OLR measurements are indeed a practical measure of organized precipitation systems, except that the horizontal scale of high clouds is increasingly "excessive" relative to the underlying raining area as the system becomes larger. The second implication arises in light of the Earth's energy budget. Tropical convective systems are not only the source of latent heating but also contribute to the radiative energy budget through the associated high clouds [*L'Ecuyer et al.*, 2006]. The relative importance of the cloud radiative heating/cooling is then expected to be enhanced as the rainfall system develops on a greater horizontal scale.

[16] **Acknowledgments.** Stimulating discussions with Tristan L'Ecuyer and Toshi Matsui and the helpful comments of the reviewers are greatly appreciated. The TRMM data sets (1B01 and 2A23) were provided by the National Aeronautics and Space Administration (NASA) Goddard Space Flight Center (GSFC). NCEP/NCAR Reanalysis data were provided by the NOAA Cooperative Institute for Research in Environmental Sciences (CIRES) Climate Diagnostics Center (CDC), Boulder, Colorado, USA, from their Web site at www.cdc.noaa.gov. TMI SST data, available at www.remss.com, are produced by the RSS and sponsored by the NASA Earth Science REASoN DISCOVER Project. This research is supported by NOAA's office of Global Program Grant NA17RJ1228#15 and NASA's Precipitation Program Grant NAG5-13694.

## References

Del Genio, A. D., and W. Kovari (2002), Climatic properties of tropical precipitating convection under varying environmental conditions, *J. Clim.*, *15*, 2597–2615.

Houze, R. A., Jr. (1977), Structure and dynamics of a tropical squall-line system, *Mon. Weather Rev.*, *105*, 1540–1567.

Houze, R. A., Jr., S. S. Chen, D. E. Kingsmill, Y. Serra, and S. E. Yuter (2000), Convection over the Pacific warm pool in relation to the atmospheric Kelvin-Rossby wave, *J. Atmos. Sci.*, *57*, 3058–3089.

Johnson, R. H., T. M. Rickenbach, S. A. Rutledge, P. E. Ciesielski, and W. H. Schubert (1999), Trimodal characteristics of tropical convection, *J. Clim.*, *12*, 2397–2418.

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

Kemball-Cook, S. R., and B. C. Weare (2001), The onset of convection in the Madden-Julian oscillation, *J. Clim.*, *14*, 780–793.

Kiladis, G. N., K. H. Straub, and P. T. Haertel (2005), Zonal and vertical structure of the Madden-Julian oscillation, *J. Atmos. Sci.*, *62*, 2790–2809.

Knutson, T. R., and K. M. Weickmann (1987), 30–60 day atmospheric oscillations: Composite life cycles of convection and circulation anomalies, *Mon. Weather Rev.*, *115*, 1407–1436.

Lau, K. M., and H. T. Wu (2003), Warm rain processes over tropical oceans and climate implications, *Geophys. Res. Lett.*, *30*(24), 2290, doi:10.1029/2003GL018567.

Leary, C. A., and R. A. Houze Jr. (1979), The structure and evolution of convection in a tropical cloud cluster, *J. Atmos. Sci.*, *36*, 437–457.

L'Ecuyer, T. S., H. Masunaga, and C. D. Kummerow (2006), Variability in the characteristics of precipitation systems in the tropical Pacific. Part II: Implications for atmospheric heating, *J. Clim.*, *19*, 1388–1406.

Lin, J., B. Mapes, M. Zhang, and M. Newman (2004), Stratiform precipitation, vertical heating profiles, and the Madden-Julian Oscillation, *J. Atmos. Sci.*, *61*, 296–309.

Lindzen, R. S., M.-D. Chou, and A. Y. Hou (2001), Does the Earth have an adaptive infrared iris?, *Bull. Am. Meteorol. Soc.*, *82*, 417–432.

Maloney, E. D., and D. L. Hartmann (2000), Modulation of eastern north Pacific hurricanes by the Madden-Julian Oscillation, *J. Clim.*, *13*, 1452–1460.

Masunaga, H., T. S. L'Ecuyer, and C. D. Kummerow (2005), Variability in the characteristics of precipitation systems in the tropical Pacific. Part I: Spatial structure, *J. Clim.*, *18*, 823–840.

Matsui, T., H. Masunaga, and R. A. Pielke Sr. (2004), Impact of aerosols and atmospheric thermodynamics on cloud properties within the climate system, *Geophys. Res. Lett.*, *31*, L06109, doi:10.1029/2003GL019287.

Nesbitt, S. W., E. J. Zipser, and D. J. Cecil (2000), A census of precipitation features in the tropics using TRMM: Radar, ice scattering, and lightning observations, *J. Clim.*, *13*, 4087–4106.

Rapp, A. D., C. Kummerow, W. Berg, and B. Griffith (2005), An evaluation of the proposed mechanism of the adoptive infrared iris hypothesis using TRMM VIRS and PR measurements, *J. Clim.*, *18*, 4185–4194.

Schumacher, C. (2006), Shallow tropical convection: How often does it rain?, *Bull. Am. Meteorol. Soc.*, *87*, 23–25.

Schumacher, C., and R. A. Houze Jr. (2003), The TRMM Precipitation Radar's view of shallow, isolated rain, *J. Appl. Meteorol.*, *42*, 1519–1524.

Short, D., and K. Nakamura (2000), TRMM radar observations of shallow precipitation over the tropical oceans, *J. Clim.*, *13*, 4107–4124.

Sobel, A. H., and C. S. Bretherton (2000), Modeling tropical precipitation in a single column, *J. Clim.*, *13*, 4378–4392.

Takayabu, Y.N., K.-M. Lau, and C.-H. Sui (1996), Observation of a quasi-2-day wave during TOGA COARE, *Mon. Weather Rev.*, *124*, 1892–1913.

Waliser, D. E., and N. E. Graham (1993), Convective cloud systems and warm-pool sea-surface temperatures: Coupled interactions and self-regulation, *J. Geophys. Res.*, *98*(D7), 12,881–12,893.

---

C. D. Kummerow and H. Masunaga, Department of Atmospheric Science, Colorado State University, Fort Collins, CO 80523-1371, USA. (kummerow@atmos.colostate.edu; masunaga@atmos.colostate.edu)