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Anomaly patterns about strong convective events in the tropics and midlatitudes: Observations from radiosondes and surface weather stations

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Abstract We use 13 years (1998–2010) of rainfall estimates from the Tropical Rainfall Measuring Mission to identify high rain events located close to radiosondes. This is done in four regions: the Western Tropical Pacific, Tropical Brazil, Southeast China, and Southeast U.S. We then construct composite anomaly patterns of temperature, relative humidity, surface pressure, convective available potential energy (CAPE), geopotential height, mass divergence, relative vorticity, and potential vorticity about these high rain events. One motivation of this analysis is to identify regional differences in the interaction between strong convective events and the background atmosphere. We find, overall, that the changes in meteorological variables which occur during the evolution of strong convective events in midlatitudes are similar to the changes that occur in the tropics. In midlatitudes, however, strong convective events are associated with stronger anomalies in surface pressure and geopotential height and exhibit a warm anomaly in the lower troposphere prior to peak rainfall. In the Southeast U.S., the near-surface layer of positive CAPE that occurs prior to high rain events is thicker than in the Western Tropical Pacific. In the two midlatitude regions, the midlevel potential vorticity maximum that develops during the growth stage of high rain events acquires a downward tilt toward the surface during the decay stage, suggesting downward transport toward the surface. A conceptual model previously used to interpret the anomaly patterns of the 2 day equatorial wave is used to interpret the anomaly patterns associated with more general types of high rain events in the tropics.

1. Introduction

With some exceptions [Cotton et al., 1989; DeMott et al., 2007], most observationally based analyses of the anomaly patterns associated with strong convective events have been based on measurements over the tropical oceans. Here, we use routine radiosonde observations to construct composite anomaly patterns of a large number of meteorological variables about high rain events in four regions: the Western Tropical Pacific, Tropical Brazil, the Southeast U.S., and Southeast China. This approach enables us to identify regional similarities and differences in the interaction between deep convection and the background atmosphere. The meteorological variables we analyze include temperature, relative humidity, column water vapor, convective available potential energy (CAPE), geopotential height, mass divergence, surface pressure, relative vorticity, and potential vorticity. We first use the Tropical Rainfall Measuring Missing (TRMM) 3B42 data set to identify $2^{\circ} \times 2^{\circ}$ grid cells in which the rain rate exceeds a particular threshold. We then look for radiosonde or surface measurements that are colocated, and within 48 h, of the TRMM high rain events. To obtain better statistics, this procedure is repeated at many radiosonde stations, using a large number of rain events from 1998 to 2010. This approach has been previously followed using routine radiosonde profiles [Sherwood and Wahrlich, 1999; Mitovski et al., 2010], measurements from field campaigns [Sobel et al., 2004; Mapes et al., 2006; DeMott et al., 2007; Mapes et al., 2009], satellite data [Zelinka and Hartmann, 2009; Masunaga, 2012], and reanalysis data [Benedict and Randall, 2007; Rapp et al., 2011]. Some of the anomaly patterns shown in this paper, especially with respect to temperature and relative humidity anomaly patterns in the tropics, therefore reproduce wellknown relationships. However, radiosonde observations have not been used to directly calculate the anomaly patterns of other meteorological variables such as geopotential height and potential vorticity or the vertical profile of CAPE in the boundary layer.



Figure 1. Open squares refer to the colocated IGRA radiosonde station and the ISH weather stations used to construct the anomaly patterns. The triangles show the radiosonde arrays used to calculate the mass divergence, relative vorticity, and potential vorticity anomaly patterns. The large rectangles show the regions within which rain events were chosen to calculate the radial variation of temperature about high rain events.

2. Data Sets

2.1. Rainfall Data

The TRMM 3B42 gridded rainfall data set is constructed from several satellite-borne sensors. These include a precipitation radar, multichannel microwave radiometer, and visible and infrared sensors [*Kummerow et al.*, 1998; *Liu et al.*, 2012]. The data are available at 3 h temporal (centered at 00:00, 03:00, 06:00, 09:00, 12:00, 15:00, 18:00, and 21:00 UTC) and $0.25^{\circ} \times 0.25^{\circ}$ horizontal resolution. We used TRMM 3B42 rainfall to identify high rain events between January 1998 and December 2010.

2.2. Radiosonde Data

The Integrated Global Radiosonde Archive (IGRA) data set is stored at the National Climatic Data Center (NCDC) [*Durre et al.*, 2006]. We used data from 53 radiosonde locations from the 1998 to 2010 time period. The locations of these stations are shown in Figure 1 (open squares) and are located in four regions: 6 from the Western Tropical Pacific, 11 from Tropical Brazil, 19 from the Southeast U.S., and 17 from Southeast China. The soundings provide temperature and relative humidity profiles at variable vertical resolution. Within the Southeast U.S. region, there were on average 42 temperature and relative humidity measurements below 100 hPa. The comparable numbers for the radiosonde stations in the Western Tropical Pacific, Tropical Brazil, and Southeast China regions are 39, 26, and 14, respectively. The soundings also provide horizontal wind measurements at the surface and at the standard pressure levels of 1000, 925, 850, 700, 500, 400, 300, 250, 200, 150, and 100 hPa. Although some of the IGRA soundings are available every 6 h, the majority of the soundings are available twice per day at the 00:00 and 12:00 UTC standard synoptic times.

The divergence and vorticity anomaly calculations require simultaneous horizontal wind measurements from triangular radiosonde arrays. For these variables, we used 2 arrays from the Western Tropical Pacific, 5 arrays from Tropical Brazil, 14 arrays from the Southeast U.S., and 17 arrays from Southeast China. The average area of the Western Tropical Pacific arrays was 296,000 km² (roughly equal to the area of a 5° × 5° grid box at the

equator), 57,000 km² for the Southeast U.S. arrays (roughly equal to a $2.5^{\circ} \times 2.5^{\circ}$ grid box at 30°N), 207,000 km² for the Tropical Brazil arrays, and 31,000 km² for the Southeast China arrays.

2.3. Hourly Surface Data

Over land, convective precipitation usually peaks in the late afternoon [*Nesbitt and Zipser*, 2003]. The rain events used to construct the anomaly patterns over land also occur more frequently in the late afternoon. Within a given region, the radiosonde are usually launched at two fixed local solar times. The incomplete diurnal sampling of the radiosondes, in combination with the existence of diurnal cycles in both the rain event frequency and the meteorological variables, will introduce biases into the calculation of our anomaly patterns. It is therefore important that, where possible, the radiosonde anomaly patterns be validated against surface measurements with better temporal resolution. The Integrated Surface Hourly (ISH) data set consists of surface weather observations from about 20,000 worldwide-distributed stations [*Lott et al.*, 2001]. The data are archived at the NCDC and contain hourly and synoptic (3-hourly) surface weather observations of several variables, including temperature, relative humidity, and pressure. We used ISH data between 1998 and 2010, from surface stations, which were colocated with the 53 IGRA radiosonde stations shown in Figure 1.

2.4. Global Positioning System Data

We used Global Positioning System (GPS) radio occultation temperature measurements from the Constellation Observing System for Meteorology, Ionosphere and Climate (COSMIC) six-satellite constellation [*Anthes et al.*, 2008], downloaded from the COSMIC Data Analysis and Archive Center. There are approximately 2000 high vertical resolution (~100 m in the troposphere) COSMIC temperature profiles over the Earth per day. Over a particular region, the COSMIC profiles are distributed roughly uniformly in local solar time. They can therefore also be used to help determine whether the calculated radiosonde temperature anomaly patterns over land exhibit a diurnal bias. We used COSMIC temperature profiles within the Southeast U.S. (between 30°N and 40°N and between 85°W and 100°W) and Southeast China (between 25°N and 35°N and between 103°E and 118°E) regions shown in Figures 1c and 1d. For both regions, we restricted attention to the convective summer season (June, July, and August) from 2006 to 2010.

3. Rain Event Definition

In the construction of the radiosonde and surface station anomaly patterns, high rain events were considered to occur at locations where the mean TRMM 3B42 rainfall over a $2^{\circ} \times 2^{\circ}$ box centered at each station exceeded 1.5 mm/h over a TRMM 3 h time interval. For the calculation of the mass divergence and relative vorticity anomaly patterns, we used triangular radiosonde arrays, and rain events were defined using the average TRMM 3B42 rain rate within each array. For the Southeast U.S. and Southeast China regions, we restricted attention to June, July, and August. For the Tropical Brazil and the Western Tropical Pacific regions, we used rain events from the entire year.

Figure 2a shows the composite TRMM 3B42 $2^{\circ} \times 2^{\circ}$ mean rain event profiles from the four regions. The mean peak rain rate of close to 3 mm/h is significantly higher than the rain event threshold of 1.5 mm/h. The composite rain event profiles were constructed from events colocated with radiosonde or surface stations, rather than from rain events within the variably sized radiosonde arrays. The rain event profiles are quite similar to each other, partly reflecting the use of a common rain event threshold for all regions. High rain events develop slightly more rapidly over the three land regions than the Tropical Western Pacific. Rainfall starts to increase roughly 9 h prior to peak rainfall and drops to near background levels roughly 9 h after peak rainfall.

Within a particular region, the GPS COSMIC temperature profiles tend to be randomly distributed in space. In order to identify high rain events colocated with GPS temperature profiles, we first calculated the mean TRMM 3B42 rain rate in $2^{\circ} \times 2^{\circ}$ grid boxes, separated from each other by 1°. For example, the Southeast U.S. region shown in Figure 1c enclosed an area between $30^{\circ}N-40^{\circ}N$ and $85^{\circ}W-100^{\circ}W$. Within this region, we calculated the mean rain rate in $10 \times 15 = 150$ overlapping $2^{\circ} \times 2^{\circ}$ grid boxes and searched for GPS temperature profiles within each box. The Southeast China region shown in Figure 1d enclosed an area between $25^{\circ}N-35^{\circ}N$ and $103^{\circ}E-118^{\circ}E$ and was also partitioned into 150 overlapping boxes.



Figure 2. (a) The variation in mean rainfall during the growth and decay of the TRMM $3B42 2^{\circ} \times 2^{\circ}$ high rain events within each region. (b) The radial variation in mean rainfall about the high rain events.

Figure 2b shows the variation in average rain rate as a function of radial distance from the high rain events. The rain events of the various regions have a similar spatial scale, with the rain rate typically decaying to a near background value within about 400 km of peak rainfall.

4. Diurnal Variability

Figure 3a shows the average diurnal variation of the TRMM 3B42 rain rate within each of the four regions. As is well known [*Nesbitt and Zipser*, 2003], the rainfall rate within the three land regions peaks in the late afternoon or early evening. Within the Tropical Western Pacific region, the rainfall rate exhibits a small peak near 3 A.M. Figure 3b shows the diurnal variation in the frequency of occurrence of the rain events used to construct the radiosonde temperature anomaly patterns. The time of the diurnal peak in the rain event frequency is usually close to the time of the diurnal rainfall peak. The larger number of rain events within the Tropical Brazil region, despite the lower rain rate, is due to the use of rain events throughout the entire year, rather than June–August for the two midlatitude regions.

Figure 4 shows the number of IGRA radiosonde profiles within each region that occurred within 48 h of a rain event and could therefore be used in the construction of the temperature anomaly patterns. Within the Southeast U.S., there were approximately 21,000 profiles at 06 and 18 h local solar time available from the 1998 to 2010 time period. We used 25,000 profiles from Tropical Brazil (08 and 20 h local solar time), 55,000 profiles from Southeast China (also at 08 and 20 h local solar time), and roughly 80,000 soundings from the Tropical Western Pacific, distributed at local solar times of 00, 09, 10, 11, 12, 21, 22, and 23 h. The broader range of local solar times from the Tropical Western Pacific reflects the broader longitudinal distribution of the radiosonde stations in this region.

5. Results

5.1. Temperature Anomaly Patterns From Radiosondes

Once a rain event at a radiosonde station had been identified, we searched for radiosonde launches that had occurred within 48 h of the rain event. These soundings were then assigned to 3 h time bins, based on the difference between the radiosonde launch and rain event times. At each location, the radiosonde



Figure 3. (a) The variation in TRMM 3B42 $2^{\circ} \times 2^{\circ}$ rainfall with local solar time, averaged over the radiosonde locations of each region shown in Figure 1. (b) The diurnal variation in the number of high rain events within each 3 h time bin, calculated from all rain events used in the construction of the temperature anomaly patterns.

measurements were used to define climatological temperature profiles for each individual launch time, month, and year. These climatological profiles were then used to convert the temperature profiles near high rain events to temperature anomaly profiles. We then averaged the temperature anomaly profiles within each 3 h time bin.

The composite temperature anomaly patterns generated using the above procedure within each region are shown in Figures 5a–5c and 5e. For all regions, high rain events are associated with an upper level warming centered at 300 hPa that is coincident with peak rainfall [*Cotton et al.*, 1989; *Sherwood and Wahrlich*, 1999; *Kiladis et al.*, 2005; *Mapes et al.*, 2006; *Mitovski et al.*, 2010]. With the exception of Southeast China, the regions



Figure 4. The number of available soundings, at each local solar time, within 48 h of the rainfall events used in the construction of the temperature anomaly patterns. Different colors refer to different regions.

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Figure 5. (a, b, c, and e) Temperature anomaly patterns about the TRMM high rain events, calculated using IGRA radiosonde profiles from each region. (d and f) Temperature anomaly patterns calculated using the COSMIC GPS temperature profiles.

also show a 700 hPa lower tropospheric cooling coincident with peak rainfall. Each region shows a strong boundary layer cooling (below 900 hPa) after peak rainfall. This cooling is significantly larger and more persistent over the three land regions than over the Tropical Western Pacific. Deep convection over the three land regions is preceded by a warm anomaly in the lower troposphere. Over Tropical Brazil and Southeast China, this warming extends upward from the surface to 800 hPa but extends to 600 hPa over the Southeast U.S.

The details of the temperature anomaly patterns shown in Figures 5a–5c and 5e will be sensitive, to some degree, to the assumptions used to define the rain events. For example, the use of a larger rain event threshold would be expected to generate larger temperature anomalies. It has been shown, however, that the shapes of the temperature anomaly patterns are not sensitive to the value of the rain event threshold or the assumed event area [*Mitovski et al.*, 2010].

5.2. Temperature Anomaly Patterns From GPS

The GPS COSMIC temperature profiles were used to also construct temperature anomaly patterns of the Southeast U.S. and Southeast China regions. During the summer convective season (June, July, and August) between 2006 and 2010, there were 2111 profiles from the U.S., and 1538 profiles from China, which were



Figure 6. Within each figure, the dotted line refers to an anomaly profile calculated using 1 h ISH surface data, while the solid lines refer to anomaly profiles calculated using 12 h IGRA radiosonde data. (top row) The variation of the surface temperature anomaly during high rain events. (middle row) The variation of surface relative humidity anomaly during high rain events. (bottom row) The variation in the surface pressure anomaly during high rain events. Each column refers to a different region: (left to right) the Tropical Western Pacific, Tropical Brazil, Southeast U.S., and Southeast China regions.

colocated within 48 h of a TRMM rain event. The COSMIC temperature anomaly patterns from these two regions are shown in Figures 5d and 5f. Although the COSMIC and radiosonde temperature anomaly patterns are in good overall agreement, the COSMIC boundary layer cooling is much smaller than the boundary layer cooling obtained from the radiosonde data. There are several reasons the COSMIC temperature anomaly patterns may be less accurate near the surface. COSMIC provides vertical profiles of atmospheric refractivity, which can be used to calculate the vertical profile of atmospheric density [*Anthes et al.*, 2008]. Variations in water vapor can be expected to more strongly affect the atmospheric density at higher temperatures closer to the surface. In the absence of an independent method of constraining the water vapor profile, the relative accuracy of the COSMIC temperature retrievals can be expected to be smaller at lower altitudes. The standard deviation of the difference between COSMIC and radiosonde temperatures does increase toward the surface, though this does not appear to be associated with an increase in the mean bias [*Sun et al.*, 2010]. The refractivity gradient at the top of the boundary layer frequently exceeds the critical refraction, which can result in a systematic negative refractivity bias in the radio occultations within the boundary layer [*Xie et al.*, 2012]. Finally, the number of COSMIC temperature retrievals decreases toward the surface, especially for altitudes below 2 km [*Anthes et al.*, 2008].

5.3. Anomalies in Pressure, Temperature, and Relative Humidity at the Surface

Figure 6 shows the surface temperature, surface relative humidity, and surface pressure anomalies during high rain events within each of the four regions, calculated from both the radiosonde and surface weather station data. In general, the radiosonde anomaly profiles (Figure 6, solid lines) are in good agreement with the profiles generated by higher temporal resolution surface station measurements (Figure 6, dotted lines).

Figure 6 (top row) shows the surface temperature anomaly during high rain events. The three land regions show larger surface cooling than the Tropical West Pacific region. This is consistent with Figure 5, which also

shows that the boundary layer cooling over land exhibits greater persistence and depth. One explanation for this difference is that the relative humidity in the lower troposphere over the three land regions tends to be smaller than the relative humidity in the lower troposphere over the ocean region. Downdrafts over land would therefore be expected to exhibit stronger negative buoyancies, have a higher probability of penetrating the boundary layer, and possibly, generate more persistent cooling. It is also possible that, due to a weaker effective heat capacity over land, the sensible heat flux from the surface in response to a cold anomaly near the ground is likely to be weaker over land than the ocean.

Figure 6 (middle row) shows the anomaly in surface relative humidity during high rain events. Within all four regions, convective rainfall is associated with a sharp increase in surface relative humidity that is coincident, or slightly lags, peak rainfall. As would be expected, the size of the relative humidity increase scales with the size of the temperature decrease, with the two midlatitude land regions exhibiting the largest increases in surface relative humidity.

Figure 6 (bottom row) shows the surface pressure anomalies during high rain events. These anomalies show larger regional differences than the temperature and relative humidity anomalies. In the Western Tropical Pacific, there is a 0.5 hPa minimum in surface pressure 10 h prior to peak rainfall, followed by a small local maximum that slightly lags peak rainfall. In the Tropical Brazil region, there is a more rapid increase in surface pressure during high rain events. In the two midlatitude regions, high rain events are associated with 1–2 hPa reductions in surface pressure over a 2–3 day timescale.

The surface pressure anomalies of the radiosonde stations within the Southeast U.S., and to a lesser extent within Southeast China and Tropical Brazil, are significantly different from the surface pressure anomalies obtained from the surface stations. These deviations have a 12 h periodicity. Figure 7a (black curve) shows the average diurnal cycle of the surface pressure anomaly for the Southeast U.S. surface stations. Figure 7a (red curve) shows the diurnal cycle of the surface pressure anomaly, calculated using only those surface pressure measurements for which the rain rate of the local grid cell exceeded 1 mm/h. During the afternoon (14:00-18:00 local time), the presence of rainfall is not associated with deviations in surface pressure from the mean diurnal cycle. However, outside this afternoon and early evening time window, rainfall is associated with reduced surface pressure. Figure 7c shows that the surface pressure anomaly during high rain events depends on the time at which the rain event occurs. Each anomaly curve is constructed from high rain events having a common TRMM time, converted here to the approximate local solar time. The three curves with the largest negative anomalies in surface pressure are associated with rain events at 21:00 (Figure 7c, solid green curve), 00:00 (Figure 7c, solid blue curve), and 03:00 (Figure 7c, solid red curve). The two anomaly curves with the smallest negative anomalies in surface pressure are associated with rain events which occur at 15:00 (Figure 7c, dashed red curve) and 18:00 (Figure 7c, solid black curve). This is consistent with Figure 7a. Rain events which occur during the late evening and early morning are associated with larger decreases in surface pressure (relative to the mean surface pressure at that local time) than rain events in the late afternoon.

The existence of a diurnal cycle in the magnitude of the surface pressure anomaly associated with high rain events leads to errors when the twice daily radiosonde profiles are used to construct the surface pressure anomaly patterns. The twice daily radiosonde launches occur at local solar times of 6 and 18 h. Figure 7b shows the surface pressure anomaly constructed from the individual sondes, as well as the combined response shown previously in Figure 6. For the two individual anomalies, there is a 24 h oscillation generated by changes in the local solar time of the rain events. For example, the curve generated using the 6 local time (LT) radiosonde launches and the curve generated using the 18 LT radiosonde launches both show maxima associated with rain events occurring at 15 LT and minima associated with rain events occurring at 00 LT. The anomaly pattern is consistent with Figure 7c, which shows weaker surface pressure anomalies for rain events in the late afternoon, and stronger surface pressure anomalies for rain events at night. The combined radiosonde surface pressure anomaly profile in Figure 7b, indicated by the dashed line, is equal to the average of the two individual radiosonde profiles, weighted by the relative frequency of the rain events used to construct the two profiles. Inspection of Figure 7b shows that the 12 h oscillation in the combined radiosonde surface pressure anomaly pattern is an artifact generated from the 24 h oscillations of the two individual curves. More generally, for variables which exhibit a coupling with convective precipitation which varies over the diurnal cycle, artifacts may be introduced when using 12-hourly radiosonde profiles to construct anomaly patterns in that variable.



Figure 7. (a) The black curve shows the diurnal variation in the surface pressure anomaly, plotted against local solar time, of the surface stations within the Southeast U.S. region shown in Figure 1 (June, July, and August). The red curve was obtained from the subset of surface pressure measurements at which the local rain rate of each surface station exceeded 1.0 mm/h. (b) The solid black curve shows the surface pressure anomaly about high rain events constructed using radiosonde measurements launched at 6 LT only. The red curve shows the surface pressure anomaly about high rain constructed from radiosonde measurements at 18 LT only. The dashed black line shows the combined surface pressure anomaly of the two radiosonde launch times. (c) Each curve shows the surface pressure response, calculated using ISH surface station data within the Southeast U.S. region, to TRMM rain events occurring at a particular time. For each curve, the local time of the TRMM rain events is shown in the legend. The circles in Figure 7b indicate surface pressure anomalies calculated from radiosondes launched at 18 LT, from rain events at 15 LT. Each red circle of Figure 7b corresponds to a red circle in Figure 7c, which indicates the effect of rain-occurring events at that time on the surface pressure anomaly. In general, maxima in the curves of Figure 7b occur at times when the corresponding rain events and the surface pressure anomaly. In general, maxima in the curves of Figure 7b occur at times when the corresponding rain events are associated with a smaller surface pressure anomaly and vice versa.

5.4. Radial Distribution of the Convective Temperature Response

The temperature anomaly patterns shown in Figure 5 show the time evolution of the local temperature response to deep convection but give no indication of the spatial scale over which these temperature anomalies extend. The radiosonde measurements can be used to calculate the vertical profile of the temperature



Figure 8. The radial distribution of the temperature anomaly about high rain events in each region. The horizontal axis refers to the radial distance between a radiosonde location and a TRMM high rain event. The patterns were constructed from radiosonde launches within 3 h of a TRMM rain event.

anomaly as a function of the radial distance from the rain events. We first identified all $2^{\circ} \times 2^{\circ}$ rain events within each of the regional boxes shown in Figure 1 and then searched for radiosonde measurements within 1000 km of the rain event which had occurred within 3 h of the rain event time. The radiosonde temperature profiles were then stored in 50 km bins, based on the distance between the rain event and the radiosonde location. Each temperature profile was converted to an instantaneous anomaly profile by subtracting the climatological launch time/monthly/yearly temperature profile at that location. The temperature anomaly profiles within each 50 km bin at each pressure level were then averaged and the results shown in Figure 8.

The dominant feature of the radial temperature anomaly pattern is an upper level (300 hPa) warming that extends roughly 1000 km outward from the rain event. All four regions also exhibit a strong near-surface (below 900 hPa) cooling that extends roughly 300 km outward from the rain event. The Western Tropical Pacific Ocean and Southeast U.S. regions also exhibit a strong lower tropospheric cooling between 800 hPa and 500 hPa. This lower tropospheric cooling is less pronounced over Tropical Brazil and absent over Southeast China. The regional differences in the local convective temperature response shown in Figure 5 are therefore not restricted to the immediate environment of deep convective events but extend to larger spatial scales.

5.5. Relative Humidity

The IGRA radiosonde data set supplies vertical profiles of water vapor pressure, which can be converted to relative humidity. At temperatures below 0°C, we normalized the vapor pressure by the saturation vapor pressure over ice [*Emanuel*, 1994]. In general, radiosonde measurements of vapor pressure are less accurate at colder temperatures [*Kley et al.*, 2000]. The relative humidity was therefore not calculated for pressure levels above 200 hPa. Figure 9 shows the relative humidity response to high rain events, calculated using a procedure similar to that used for the temperature anomaly patterns. Within each region, peak rainfall is associated with values of relative humidity that exceed 80% at most altitudes. The gradual moistening of the



Figure 9. The variation in relative humidity during the evolution of high rain events in each region. The horizontal axis refers to time since peak rainfall.

lower troposphere prior to high rain events has been attributed to the influence of congestus clouds [*Takayabu et al.*, 1996; *Mapes et al.*, 2006], but this interpretation has been recently challenged [*Hohenegger and Stevens*, 2013; *Masunaga*, 2013]. Rain events within the Southeast U.S. occur within a background atmosphere that is significantly drier than the other three regions, particularly in the upper troposphere.

5.6. Total Column Water Vapor Anomaly

The relative humidity and temperature measurements of the radiosonde profiles can be used to calculate the total column water vapor, defined here as the mass of water vapor per unit area between the surface and 200 hPa (also known as the precipitable water). Figure 10 shows the column water vapor anomaly during the



Figure 10. The variation of the total column water vapor anomaly about high rain events in each region. The total column water vapor anomaly is a mass-weighted vertical integral of the specific humidity anomaly between 1000 and 200 hPa, as measured by a radiosonde profile.



Figure 11. (a and b) The variation in boundary layer CAPE during high rain events in the Western Tropical Pacific and Southeast U.S. regions. The horizontal axis refers to time since peak rainfall. (c and d) Similar to Figures 11a and 11b but show the CAPE anomaly.

growth and decay of high rain events within each region. Rain events in the Southeast U.S. occur at much lower values of column water than the other three regions (not shown). However, the change in column water during high rain events in the Southeast U.S. is very similar to the change in Southeast China and the Western Tropical Pacific (7–10 kg/m²). Rain events in Tropical Brazil are associated with smaller increases in column water (less than 4 kg/m²).

5.7. Convective Available Potential Energy

The existence of air parcels near the surface with positive convective available potential energy (CAPE) is a precondition for moist convection [*Sherwood and Wahrlich*, 1999]. We calculate CAPE by starting with air parcels near the surface whose pressure, temperature, and relative humidity are equal to the values given by a radiosonde profile at a particular pressure level. We lift the air parcel subject to the assumptions of constant moist static energy and total water (i.e., no condensate removal), no mixing, and no ice formation. Unfortunately, only the radiosonde profiles of the Western Tropical Pacific and Southeast U.S. regions had sufficient vertical resolution to resolve the vertical structure of boundary layer CAPE.

The evolution of CAPE and the CAPE anomaly during the growth and decay of high rain events within these two regions is shown in Figure 11. In both regions, the decay stages of high rain events are associated with strongly negative CAPE anomalies which maximize near the surface several hours after peak rainfall. This CAPE reduction is presumably associated with some combination of preferential entrainment of higher moist static energy air parcels from the boundary layer by convective updrafts and by the injection of lower moist static energy air into the boundary layer from the midtroposphere [*Zipser*, 1977; *Sherwood and Wahrlich*, 1999].

Figure 11 also shows that the near-surface layer of positive CAPE is considerably thicker in the Southeast U.S. region than in the Western Tropical Pacific region. Model simulations suggest that convective storms which grow from a deeper near-surface moist layer, especially in the range of 1.5–2 km, can experience significantly less updraft dilution and develop stronger updraft speeds than those which grow from thinner moist layers [*McCaul and Cohen*, 2002]. In this case, Figure 11 may partially explain the well-known fact that convective updrafts which occur over the ocean exhibit smaller updraft speeds than those which occur over land [*Lucas et al.*, 1994], though other factors are probably also involved [*Robinson et al.*, 2008].

Due to the diurnal cycle in CAPE over land [*Dai et al.*, 1999] and the higher frequency of rain events during the late afternoon, some of the CAPE enhancement within the Southeast U.S. shown in Figure 11b prior to peak rainfall can be attributed to the diurnal cycle. The CAPE anomaly, in which the diurnal effect has been reduced by subtracting from each CAPE measurement the mean value of CAPE for each particular radiosonde launch time, station, and month, is shown in Figures 11c and 11 d. Within the Western Tropical Pacific, there is very little change in CAPE prior to peak rainfall. Within the Southeast U.S., the largest positive CAPE anomaly (~ 600 J/kg) occurs for parcels originating from 950 hPa roughly 6 h prior to peak rainfall.

Figure 11 also shows that there is a 12 h oscillation in the Southeast U.S. CAPE and CAPE anomaly patterns that mainly occurs prior to peak rainfall. This oscillation is an artifact generated by the same mechanism which produces the surface pressure oscillation shown in Figure 7. Although Figure 6 (top row) shows that the surface temperature errors in the Southeast U.S. introduced by this sampling bias are small, these errors have a larger relative impact on CAPE.

5.8. Geopotential Height

Geopotential height profiles were calculated from the surface pressure and temperature and relative humidity profiles of individual radiosondes. The geopotential height at the surface of each location was set equal to the local surface station altitude. We used stations whose altitude was less than 200 m. We then iteratively calculated the geopotential height upward from the surface using the hypsometric equation. This equation defines the geopotential height (GPHT) at a pressure level n (p_n) in terms of the geopotential height at a lower pressure level n-1 (p_{n-1}), the virtual temperature at levels n-1 ($T_{v,n-1}$) and n ($T_{v,n}$) and the pressure thickness dp.

$$\mathsf{GPHT}_n = \mathsf{GPHT}_{n-1} + \mathsf{R}_\mathsf{d}\left(\frac{T_{\mathsf{v},n-1} + T_{\mathsf{v},n}}{2\mathsf{g}}\right) d\,\mathsf{In}p. \tag{1}$$

In this equation, p refers to pressure in Pascals, R_d to the specific gas constant of dry air (287.04 J kg⁻¹ K⁻¹), T_v to the virtual temperature of an air parcel, and g to standard gravity (9.806 m s⁻²), and d ln p is defined as

$$d\ln p = \frac{p_{n-1} - p_n}{0.5(p_{n-1} + p_n)}.$$
(2)

To obtain the geopotential height anomaly, we subtracted from each geopotential height the mean geopotential height for that station, month, and radiosonde launch time. We then grouped the resulting geopotential height anomaly profiles of each region in 3 h time bins about their respective rain events. Figure 12 shows the geopotential height anomaly patterns associated with the high rain events of each region. The anomaly pattern of the Southeast U.S. region exhibits a 12 h oscillation. This oscillation is stronger in the lower troposphere and can be attributed to the 12 h oscillation in lower tropospheric surface pressure discussed earlier.

The geopotential height anomaly patterns of the four regions show some common features. Each region shows a strong positive geopotential height anomaly in the upper troposphere centered at 200 hPa. This anomaly is generated mainly by the upper tropospheric 300 hPa warm anomaly shown in Figure 5. Figure 12 also shows that the positive upper tropospheric geopotential height anomalies within each region rapidly decrease above 200 hPa. This decrease with height can be attributed to the intense narrow cold anomalies centered near 150 hPa shown in Figures 5 and 8, which prevent the positive upper tropospheric geopotential height anomalies associated with strong convective events from penetrating into the lower stratosphere [Holloway and Neelin, 2007].

In the two tropical regions, the negative midlevel geopotential anomaly is largest at 500 hPa and roughly symmetric about peak rainfall. The onset of negative geopotential height anomalies in the lower tropospheric begins prior to the lower tropospheric cooling shown in Figure 5a and can be mainly attributed to local reductions in surface pressure. For example, Figure 6 (bottom row, first column) shows that the surface pressure of the Western Tropical Pacific region decreases by roughly 0.6 hPa prior to peak rainfall. This would generate a geopotential height decrease of roughly 5.2 m, comparable with the lower tropospheric geopotential height anomalies prior to peak rainfall shown in Figure 12a.

In the two midlatitude regions, negative surface pressure anomalies several days prior to peak rainfall contribute to negative geopotential height anomalies in the boundary layer. As peak rain rates develop, the cold anomalies in the boundary layer and midlevels contribute to the intensification and upward extension of



Figure 12. The variation in geopotential height (GPHT) during high rain events in each region. The horizontal axis refers to time since peak rainfall.

these negative geopotential height anomalies to 400 hPa. These negative midlevel geopotential height anomalies are much stronger, deeper, and more persistent than those observed in the tropics.

5.9. Mass Divergence

At a given pressure level, the net horizontal mass inflow or outflow from a convecting region can be estimated using horizontal wind measurements from radiosonde arrays. We calculated the mass divergence of the triangular arrays shown in Figure 1 using the following expression.

$$\nabla \cdot \vec{V} = \sum \frac{\vec{V} \cdot \vec{n} dI}{A}$$
(3)

 $V \cdot \vec{n}$ refers to the wind component normal with respect to a line that connects two points of the array, *dl* to the distance between the two points, and *A* to the surface area of the array.

The number of divergence profiles available to construct the anomaly patterns is usually much less than the number of available temperature or relative humidity profiles, partly because a divergence calculation requires a complete ring of simultaneous horizontal wind measurements. The native resolution TRMM 3B42 data set was used to calculate the average rainfall rate within the arrays shown in Figure 1. The rain event frequency is in general smaller for larger arrays. On average, there were 164 divergence profiles available per 3 h time bin from the 2 arrays in the Western Tropical Pacific, 205 profiles from the 14 arrays in the Southeast U.S., 414 profiles from the 17 arrays in the Southeast China, and 102 profiles from the 5 arrays in the Equatorial Brazil.

Figure 13 shows the divergence anomaly patterns of each region. The four patterns are quite similar. The dominant feature of each pattern is a strong divergence maximum in the upper troposphere (200 hPa). This divergence maximum occurs near peak rainfall but in three of the regions shows a tendency for stronger persistence after peak rainfall. Each region also shows a strong convergence feature in the boundary layer prior to peak rainfall, extending from the surface to 800 hPa. In the Western Tropical Pacific, this feature occurs 6 h prior to peak rainfall. For rain events within the three land regions, the time difference between the boundary layer divergence maximum and peak rainfall is roughly equal to 3 h.



Figure 13. The variation in the mass divergence anomaly during high rain events in each region. The horizontal axis refers to time since peak rainfall.

The divergence anomaly pattern of the Western Tropical Pacific region shown in Figure 13 exhibits an antisymmetric midlevel divergence dipole about peak rainfall [*Thompson et al.*, 1979; *Mapes et al.*, 2006, 2009; *Mitovski et al.*, 2010]. Within the three land regions, there is also a tendency for midlevel divergence to occur prior to peak rainfall and midlevel convergence to occur after peak rainfall. However, the timing of these divergence features is closer to peak rainfall than in the Western Tropical Pacific.

5.10. Relative Vorticity

Relative vorticity was calculated from the horizontal wind measurements of a radiosonde array using the following expression:

ζ

$$\Sigma = \Sigma \frac{V_p dI}{A}$$
(4)

 V_p refers to the wind component parallel to a line connecting two points in the array, dl to the distance between the two points, and A to the surface area of the array.

Unfortunately, the relative vorticity anomaly patterns of the two tropical regions were irregular and did not show strong features. This can probably be attributed to the weakness of the Coriolis parameter at the latitudes characteristic of the radiosonde arrays of these two regions, which would weaken the relative impact of the divergence term on the local vorticity budget. Figures 14a and 14b show the relative vorticity anomaly patterns of the two midlatitude regions. Within the Southeast China and Southeast U.S. regions, positive relative vorticity anomalies develop in the boundary layer 1 day prior to peak rainfall. These features are presumably associated with the coincident development of boundary layer convergence. At peak rainfall, the positive vorticity anomalies intensify and grow vertically to 400 hPa. The timing in the growth and vertical extent of these vorticity anomalies is consistent with the midlevel convergence features shown in Figure 13.



Figure 14. (a and b) The variation in the relative vorticity anomaly about high rain events, calculated from radiosonde arrays in the Southeast U.S. and China. (c and d) The variation in potential vorticity (PV) about high rain events. (e and f) The variation in the potential vorticity anomaly about high rain events. The horizontal axis refers to time since peak rainfall. The potential vorticity is expressed in potential vorticity units (PVU), where 1 PVU = $10^{-6} \text{ m}^2 \text{ s}^{-1}$ K kg⁻¹.

5.11. Potential Vorticity

The production of potential vorticity (PV) within convective systems is believed to have a role in the development of the midlevel jet [*Franklin et al.*, 2006]. PV was calculated from the temperature and horizontal wind profiles of the radiosonde arrays using the following expression:

$$\mathsf{PV} = (\zeta + f) \left(-g \frac{\partial \theta}{\partial p} \right) \tag{5}$$

In this equation, ζ refers to relative vorticity, *f* to the Coriolis parameter, *g* to standard gravity (9.806 m s⁻²), θ to potential temperature, and *p* to pressure.

The PV and PV anomaly patterns about high rain events within the Southeast U.S. and Southeast China regions are shown in Figures 14c–14f, respectively. The tropopause is indicated by the strong vertical gradient in PV near 200 hPa. In both regions, high rain events are associated with strong negative PV anomalies near the tropopause. The development of this anomaly is consistent with a reduction in convective heating in going from the troposphere to the stratosphere.

Figures 14e and 14f also show that high rain events within the two regions are associated with positive PV anomalies that extend from the surface to 400 hPa. The stratiform type temperature response shown in Figures 5 and 8 would be expected to be associated with positive stability and PV anomalies in the



Figure 15. (a) The relative humidity tendency pattern (dRH/dt) calculated from the Western Tropical Pacific relative humidity pattern shown in Figure 9a. (b) The observed temperature tendency pattern (dT/dt), calculated from the Western Tropical Pacific temperature anomaly patterns shown in Figure 5a. (c) The vertical pressure velocity calculated from the Western Pacific divergence anomaly pattern shown in Figure 13a. (d) The temperature tendency pattern due to vertical advection, calculated from the vertical pressure velocity and stability anomaly patterns.

midtroposphere [*Hertenstein and Schubert*, 1991; *May et al.*, 1994; *Franklin et al.*, 2006]. It has been shown that convective rainfall variability within the western half of the Southeast U.S. region shown in Figure 1c is modulated by PV anomalies generated along Rocky Mountains [*Li and Smith*, 2010]. This source of convective variability is likely to have some influence on the overall PV anomaly in this region. Figure 14e also shows that high rain events in the Southeast U.S. are also associated with strong positive PV anomalies near the surface. This increase is consistent with the increase in boundary layer stability that occurs during high rain events.

5.12. Tendencies

Figure 15a shows the relative humidity tendency about high rain events in the Western Tropical Pacific, calculated from the relative humidity anomaly pattern shown in Figure 9a. The tendency pattern is dominated by moistening during the growth stage of high rain events at all altitudes but especially in the upper troposphere. During the decay stage, drying is strongest in the lower troposphere. Figure 15b shows the temperature tendency about high rain events in the Western Tropical Pacific, calculated from the temperature anomaly patterns shown in Figure 5a. The pattern is dominated by the development of a stratiform temperature response prior to peak rainfall and the erosion of this response after peak rainfall.

Diagnostic interpretations of the anomaly patterns which occur during the evolution of strong convective events usually attribute the observed patterns to a combination of convective and large-scale forcings [*Yanai et al.*, 1973]. The large-scale forcings refer, for example, to the temperature and relative humidity tendencies associated with the mean vertical motion within a radiosonde array and to horizontal advection. The convective forcings then refer to the sum of the remaining physical tendencies which cannot be attributed to the large-scale flow. These include not only the tendencies associated with moist convective circulations internal to the radiosonde array but also to the tendencies generated by other forms of turbulent mixing as well as radiative heating and cooling. For a convective system, the particular partitioning of the observed (or residual) tendency into large-scale and convective components will depend on the size of the radiosonde array used to diagnose the mean vertical velocity within the array.

Figure 15c shows the evolution of the domain averaged vertical motion (pressure velocity) within the two Western Tropical Pacific radiosonde arrays shown in Figure 1. It was calculated from the divergence anomaly pattern shown in Figure 13a, the climatological divergence profile of each array, and by imposing the boundary

condition $\omega = 0$ at the 100 hPa pressure level. During the growth stage of high rain events, there is strong upward motion throughout the troposphere, with somewhat larger ascent in the lower troposphere. After peak rainfall, the vertical motion is mainly confined to the upper troposphere [*Mapes et al.*, 2006]. Figure 15d shows the temperature tendency pattern associated with the vertical motion pattern shown in Figure 15c, calculated by multiplying the calculated pressure velocity by the appropriate static stability. The heating rate from the large-scale vertical motion appears to account for some features of the observed temperature tendencies. In particular, the lower tropospheric warming after peak rainfall appears to be partially due to the weakness of the lower tropospheric dynamical cooling during the decay stage of high rain events.

6. Discussion in Terms of Two-Mode Dynamical Response

Most of the rainfall variability in the Tropical West Pacific is associated with clusters of enhanced convective activity propagating parallel to the equator and which have horizontal flows similar to those exhibited by equatorially trapped shallow water waves [*Kiladis et al.*, 2009]. One of the more common wave types has a period of roughly 2 days, usually propagates westward with a speed of 10–30 m/s and is referred to as the 2 day wave [*Takayabu*, 1994]. The temperature, divergence, and specific humidity anomaly patterns generated by these waves [*Kiladis et al.*, 2009] are similar to the anomaly patterns obtained here by averaging over all tropical high rain events. The anomaly patterns of the 2 day waves can be reproduced by a simple model in which the dynamical response to convective heating is dominated by two vertical modes [*Haertel and Kiladis*, 2004]. The first vertical mode is associated with upward or downward vertical motion which extends throughout the troposphere, peaks at midlevels and can be mainly thought of as the dynamical response to full depth convective heating. The dynamical heating from this mode is roughly equal to the full depth convective heating but almost exactly out of phase. As illustrated in Figure 16a, the full depth heating generated by deep convection is therefore almost exactly canceled locally by the adiabatic cooling generated by the upward motion of the first mode. As a result, the induced subsidence heating and drying from deep convection is effectively exported to larger spatial scales.

The convective heating and induced vertical motion patterns associated the second vertical mode [Haertel and Kiladis, 2004] are illustrated in Figure 16c. During the growth stage of high rain events, there is convective warming in the lower troposphere (e.g., from cumulus congestus clouds) and convective cooling in the upper troposphere. The large-scale upward motion in the lower troposphere and downward motion in the upper troposphere of mode 2 give rise to the observed midlevel divergence that occurs prior to peak rainfall, shown in Figure 13a. As with the first vertical mode, the heating tendencies associated with convection and the dynamical response oppose one another. However, because the magnitude of the dynamical heating exceeds the magnitude of the convective heating, the observed temperature tendency during the growth stage of high rain events is in phase with the dynamically induced cooling of the lower troposphere and the dynamical warming of the upper troposphere. After peak rainfall, the convective heating of the second mode is associated with convective cooling in the lower troposphere and convective warming in the upper troposphere (e.g., as would arise from precipitating stratiform anvil clouds). The induced dynamical motion is downward in the lower troposphere and upward in the upper troposphere, as required to generate the observed midlevel convergence. The magnitude of the dynamical heating again exceeds the magnitude of the convective heating, so that the net heating rate from the second mode is positive in the lower troposphere and negative in the upper troposphere, giving rise to the observed residual heating pattern which damps the stratiform temperature anomaly pattern generated during the growth stage of high rain events.

The net upward lower tropospheric motion of the second mode would favor moistening of the lower troposphere during the growth stage. Conversely, the net downward motion of the second mode would favor the drying of the lower troposphere during the decay stage. These effects are consistent with the observed relative humidity tendencies shown in Figure 15a. However, for the upper troposphere, it is not possible to interpret the observed relative humidity tendency in terms of the expected residual vertical motion of the second mode. This is particularly true during the growth stage where the upper troposphere warms (suggesting net descent), but the observed moistening is quite strong. This is presumably a reflection of the relatively greater role of detrained condensate in affecting the moisture budget of the upper troposphere.

In the above conceptual model [Haertel and Kiladis, 2004], the growth and decay of the dominant stratiform temperature response is attributed to an imbalance between the convective and dynamical heating of the



Figure 16. (a) The convective heating and large-scale circulation associated with mode 1. The convective heating is indicated with the red oval and the large-scale dynamical motions by the arrows. (b) Rainfall and column water anomalies about high rain events in Western Tropical Pacific region. (c) The convective heating and large-scale circulations associated with mode 2. Dark blue and red ovals show observed cold and warm temperature anomalies. This figure has been adapted and modified from *Haertel and Kiladis* [2004, Figure 10].

second mode. It is not clear why the dynamical heating tendency of the second mode should be larger than that of the convective heating tendency. However, presumably, there is a self selection procedure whereby the size, propagation speed, and heating patterns of observed convective systems are self selected in ways that are favorable to their continued existence. This would include the rapid removal of the deep convective

heating signature via the mode 1 dynamical response, so that the rainfall system continues to have sufficient CAPE, and the midlevel divergence during the growth stage, which would increase the column moist static energy [*Haertel et al.*, 2008].

The midlevel divergence and associated upward lower tropospheric motion prior to peak rainfall would also promote the development of positive column water vapor anomalies during the growth stage of high rain events. Column water affects the growth of convective instabilities by modifying the effect of entrainment on updraft buoyancy [*Sherwood*, 1999; *Sherwood* and *Wahrlich*, 1999; *Raymond*, 2000]. For rain events in the Western Tropical Pacific, Figure 10 shows that column water starts to increase up to 48 h prior to peak rainfall. This is much earlier than the onset of increases in rainfall (about 12 h prior to peak rainfall) and suggests that the induced lower tropospheric uplift from the second mode may extend a considerable distance in front of a rain event. The radial temperature anomalies shown in Figure 8 support this viewpoint by showing that the lower tropospheric cold anomalies extend roughly 600 km outward from high rain events, again, much larger than the size of the rain event itself. A slow synoptic-scale rise in column water starting roughly 36 h prior to peak rainfall also occurring prior to rainfall increases has been previously observed in other data sets [*Holloway and Neelin*, 2010].

The development of the stratiform temperature response during the growth stage of high rain events contributes to increases in midlevel stability and PV. During the decay stage, the net downward motion in the lower troposphere from the convective and large-scale circulations (which contribute to the observed decrease in lower tropospheric relative humidity) would also be expected to transport the enhanced midlevel PV toward the surface. Although we could not calculate PV anomaly patterns of the two tropical regions, the PV patterns shown in Figure 14 for the Southeast U.S. and Southeast China regions give support for the existence of a tongue of higher PV air extending downward toward the surface from the 400 hPa level after peak rainfall. The downward advection of the midlevel PV maximum after peak rainfall may play a role in hurricane genesis. In the "top down" theory of hurricane genesis [*Gjorgjievska and Raymond*, 2013], a midlevel positive PV anomaly generated by the stratiform heating profile of a mesoscale convective system is advected toward the surface by evaporative cooling, where it then initiates the formation of a warm core anticyclone [*Bister and Emanuel*, 1997].

7. Summary

We have used the TRMM 3B42 gridded rainfall data set to identify $2^{\circ} \times 2^{\circ}$ rain events in four regions: the Western Tropical Pacific, Tropical Brazil, Southeast China, and Southeast U.S. Within each region, we selected rain events in close proximity to radiosonde or surface weather stations. These measurements were then used to construct composite anomaly patterns of a large number of meteorological variables about high rain events. One motivation of this analysis was to determine the regional similarities and differences in the interaction between strong convective events and the background atmosphere. The second motivation was to help determine the pathways by which the atmosphere returns to a balanced state following moist convection and to help understand how these pathways affect the evolution of convective systems.

Our analysis shows that there are many similarities in the interaction between strong convective events and the background atmosphere between the four different regions. With the partial exception of Southeast China (which appears to lack the 700 hPa cooling maximum), deep convection occurs in association with the development of a stratiform type temperature response in the background atmosphere. In all four regions, the upper tropospheric warming extends roughly 1000 km outward from deep convective events, while the lower tropospheric cooling extends roughly 600 km. The upper tropospheric warming and lower tropospheric cooling generate increased stability at midlevels. The midlevel stability increase can be expected to enhance convective detrainment in the midtroposphere and promote the development of cumulus congestus clouds. High rain events in each of the four regions are associated with strong boundary layer cooling during the decay stage. This boundary layer cooling was stronger in the three land regions than in the Western Tropical Pacific.

In two of the four regions, we were able to calculate the time evolution of the vertical structure of CAPE in the boundary layer during high rain events. In agreement with previous studies [*Sherwood and Wahrlich*, 1999] the boundary layer cooling after peak rainfall contributes to the development of strongly negative CAPE anomalies throughout the boundary layer in both regions. Within the Southeast U.S. region, the CAPE

enhancement prior to peak rainfall was roughly 200 J/kg. The near-surface layer of positive CAPE was much deeper over land than the ocean. Within the Southeast U.S. region, the positive CAPE layer extended to roughly 800 hPa, whereas within the Western Tropical Pacific region it extended to only 900 hPa.

There were regional variations in the interaction between strong convective events and the background atmosphere. In the two midlatitude regions, the middle and lower tropospheric negative geopotential height anomalies that developed during and after high rain events were much stronger, more vertically coherent, and more persistent than in the two tropical regions. These more strongly negative geopotential height anomalies can be partly attributed to the more strongly negative surface pressure anomalies that develop during midlatitude convection. An additional difference is that each of the three land regions exhibited warming in the lower troposphere prior to peak rainfall, a feature absent from the tropical ocean region. Finally, the midlevel stratiform convergence of the two midlatitude regions occurred closer in time to peak rainfall than in the two tropical regions.

Within the Southeast U.S. region, rain events which occur during the evening are associated with larger reductions in surface pressure than those which occur during the day. It is not clear whether the larger surface pressure reductions at night were driven by a diurnal change in the nature of deep convection, or whether the larger surface pressure reduction at night is associated with more strongly forced synoptic environments for convective events. However, if convection does indeed force the larger nighttime surface pressure reductions, diurnal changes in behavior of deep convection may contribute to the observed diurnal cycle of surface pressure in this region.

Convective circulations modify the temperature, relative humidity, and surface pressure of the background atmosphere and contribute to the generation of geopotential height anomalies. These geopotential height anomalies, in concert with any preexisting dynamical forcings, contribute to the large-scale circulations which occur in association with convective events. The observed anomaly patterns therefore result from both the direct effects of convection and the large-scale flow. We have suggested that a two-mode dynamical model originally developed to explain the anomaly patterns of the 2 day wave [*Haertel and Kiladis*, 2004], and later used to interpret anomalies associated with the Madden-Julian oscillation [*Haertel et al.*, 2008], is an appropriate conceptual model for interpreting the mean anomaly patterns of high rain events over the tropical oceans. In this model, the negative midlevel geopotential height anomalies that develop during the growth stage of convective events help drive a convergent midlevel inflow toward high rain events during the decay stage that requires, by mass balance, compensatory lower tropospheric uplift in the background atmosphere. This uplift appears to be responsible for the development of positive column water vapor anomalies which develop up to 48 h prior to peak rainfall and roughly 36 h prior to the onset of increases in rain rate.

We were able to calculate the vorticity and potential vorticity anomaly patterns of high rain events in the Southeast China and the Southeast U.S. regions. In both of these regions, high rain events are associated with negative anomalies in both upper tropospheric relative and potential vorticity which extend into the lower stratosphere. Below 400 hPa, convective events are associated with positive anomalies in relative vorticity and potential vorticity. During the decay stage of high rain events, the midlevel positive PV anomaly appears to be transported toward the surface by the net downward circulation.

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References

Anthes, R. A., et al. (2008), The COSMIC/FORMOSAT-3 Mission: Early Results, *Bull. Am. Meteorol. Soc., 89*, 313–333, doi:10.1175/BAMS-89-3-313. Benedict, J., and D. A. Randall (2007), Observed characteristics of the MJO relative to maximum rainfall, *J. Atmos. Sci., 64*, 2332–2354, doi:10.1175/JAS3968.1.

Bister, M., and K. A. Emanuel (1997), The genesis of Hurricane Guillermo: TEXMEX analyses and a modeling study, *Mon. Weather Rev.*, 125, 2662–2682, doi:10.1175/1520-0493(1997)125<2662:TGOHGT>2.0.CO;2.

Cotton, W. R., M.-S. Lin, R. L. McAnelly, and C. J. Tremback (1989), A composite model of mesoscale convective complexes, *Mon. Weather Rev.*, 117, 765–783, doi:10.1175/1520-0493(1989)117<0765:ACMOMC>2.0.CO;2.

Dai, A., F. Giorgi, and K. E. Trenberth (1999), Observed and model-simulated diurnal cycles of precipitation over the contiguous United States, J. Geophys. Res., 104, 6377–6402, doi:10.1029/98JD02720.

DeMott, C., D. A. Randall, and M. Khairoutdinov (2007), Convective precipitation variability as a tool for general circulation model analysis, J. Clim., 20, 91–112, doi:10.1175/JCLI3991.1.

Durre, I., R. S. Vose, and D. B. Wuertz (2006), Overview of the integrated global radiosonde archive, J. Clim., 19, 53–68, doi:10.1175/JCLI3594.1. Emanuel, K. A. (1994), Atmospheric Convection, pp. 580, Oxford Univ. Press, New York.

Franklin, C. N., G. J. Holland, and P. T. May (2006), Mechanisms for the generation of mesoscale vorticity features in tropical cyclone rainbands, *Mon. Weather Rev.*, 134, 2649–2669, doi:10.1175/MWR3222.1.

Gjorgjievska, S., and D. J. Raymond (2013), Interaction between dynamics and thermodynamics during tropical cyclogenesis, Atmos. Chem. Phys. Discuss., 13, 18,905–18,950, doi:10.5194/acpd-13-18905-2013.

Haertel, P. T., and G. N. Kiladis (2004), Dynamics of 2-day equatorial waves, J. Atmos. Sci., 61, 2707-2721, doi:10.1175/JAS3352.1.

Haertel, P. T., G. N. Kiladis, A. Denno, and T. M. Rickenbach (2008), Vertical-mode decompositions of 2-day waves and the Madden–Julian oscillation, J. Atmos. Sci., 65, 813–833, doi:10.1175/2007JAS2314.1.

Hohenegger, C., and B. Stevens (2013), Preconditioning deep convection with cumulus congestus, J. Atmos. Sci., 70, 448–464, doi:10.1175/ JAS-D-12-089.1.

Holloway, C. E., and J. D. Neelin (2007), The convective cold top and quasi equilibrium*, J. Atmos. Sci., 64, 1467–1487, doi:10.1175/JAS3907.1.
 Holloway, C. E., and J. D. Neelin (2010), Temporal relations of column water vapor and tropical precipitation, J. Atmos. Sci., 67, 1091–1105, doi:10.1175/2009JAS3284.1

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, doi:10.1175/JAS3520.1.

Kiladis, G. N., M. C. Wheeler, P. T. Haertel, K. H. Straub, and P. E. Roundy (2009), Convectively coupled equatorial waves. *Rev. Geophys.*, 47, RG2003, doi:10.1029/2008RG000266.

Kley, D., J. M. Russell III, and C. Phillips (Eds.) (2000), SPARC Assessment of Upper Tropospheric and Stratospheric Water Vapor, WMO/TD-143, World Clim. Res. Programme, Geneva.

Kummerow, C., W. Barnes, T. Kozu, J. Shiue, and J. Simpson (1998), The Tropical Rainfall Measuring Mission (TRMM) sensor package, J. Atmos. Oceanic Technol., 15, 809–817, doi:10.1175/1520-0426(1998)015<0809:TTRMMT>2.0.CO;2.

Li, Y., and R. B. Smith (2010), The detection and significance of diurnal pressure and potential vorticity anomalies east of the Rockies, J. Atmos. Sci., 67, 2734–2751, doi:10.1175/2010JAS3423.1.

Liu, Z., D. Ostrenga, W. Teng, and S. Kempler (2012), Tropical Rainfall Measuring Mission (TRMM) precipitation data and services for research and applications, *Bull. Am. Meteorol. Soc.*, 93, 1317–1325, doi:10.1175/BAMS-D-11-00152.1.

Lott, N., R. Baldwin, and P. Jones (2001), The FCC integrated surface hourly database: A new resource of global climate date. NOAA, Nat. Clim. Data Center Tech. Rep., 2001–01, 42.

Lucas, C., E. J. Zipser, and M. A. Lemone (1994), Vertical velocity in oceanic convection off tropical Australia, J. Atmos. Sci., 51, 3183–3193, doi:10.1175/1520-0469(1994)051<3183:VVIOCO>2.0.CO;2.

Mapes, B. E., S. Tulich, J. Lin, and P. Zuidema (2006), The mesoscale convection life cycle: Building block or prototype for large-scale tropical waves?, *Dyn. Atmos. Oceans*, 42, 3–29, doi:10.1016/j.dynatmoce.2006.03.003.

Mapes, B. E., J. Bacmeister, M. Khairoutdinov, C. Hannay, and M. Zhao (2009), Virtual field campaigns on deep tropical convection in climate models, J. Clim., 22, 244–257, doi:10.1175/2008JCLI2203.1.

Masunaga, H. (2012), A satellite study of the atmospheric forcing and response to moist convection over tropical and subtropical oceans, J. Atmos. Sci., 69, 150–167, doi:10.1175/JAS-D-11-016.1.

Masunaga, H. (2013), A satellite study of tropical moist convection and environmental variability: A moisture and thermal budget analysis, J. Atmos. Sci., 70, 2443–2466, doi:10.1175/JAS-D-12-0273.1.

May, P. T., G. J. Holland, and W. L. Ecklund (1994), Wind profiler observations of tropical storm Flo at Saipan, *Weather Forecast.*, 9, 410–426, doi:10.1175/1520-0434(1994)009<0410:WPOOTS>2.0.CO;2.

McCaul, E. W., and C. Cohen (2002), The impact on simulated storm structure and intensity of variations in the mixed layer and moist layer depths, *Mon. Weather Rev.*, *130*, 1722–1748, doi:10.1175/1520-0493(2002)130<1722:TIOSSS>2.0.CO;2.

Mitovski, T., I. Folkins, K. von Salzen, and M. Sigmond (2010), Temperature, relative humidity, and divergence response to high rainfall events in the tropics: Observations and models, J. Clim., 23, 3613–3625, doi:10.1175/2010JCLI3436.1.

Nesbitt, S. W., and E. J. Zipser (2003), The diurnal cycle of rainfall and convective intensity according to three years of TRMM measurements, J. Clim., 16, 1456–1475, doi:10.1175/1520-0442-16.10.1456.

Rapp, A. D., C. Kummerow, and L. D. Fowler (2011), Interactions between warm rain clouds and atmospheric preconditioning for deep convection in the tropics. J. Geophys. Res., 16, D23210 doi:10.1029/2011JD016143.

Raymond, D. (2000), Thermodynamic control of tropical rainfall, Q. J. R. Meteorol. Soc., 126, 889-898, doi:10.1002/qj.49712656406.

Robinson, F. J., S. C. Sherwood, and Y. Li (2008), Resonant response of deep convection to surface hot spots, J. Atmos. Sci., 65, 276–286, doi:10.1175/2007JAS2398.1.

Sherwood, S. C. (1999), Convective precursors and predictability in the Tropical Western Pacific, *Mon. Weather Rev.*, 127, 2977–2991, doi:10.1175/1520-0493(1999)127<2977:CPAPIT>2.0.CO;2.

Sherwood, S. C., and R. Wahrlich (1999), Observed evolution of tropical deep convective events and their environment, *Mon. Weather Rev.*, 127, 1777–1795, doi:10.1175/1520-0493(1999)127<1777:OEOTDC>2.0.CO;2.

Sobel, A. H., S. E. Yuter, C. S. Bretherton, and G. N. Kiladis (2004), Large-scale meteorology and deep convection during TRMM KWAJEX, Mon. Weather Rev., 132, 422–444, doi:10.1175/1520-0493(2004)132<0422:LMADCD>2.0.CO;2.

Sun, B., A. Reale, D. J. Seidel, and D. C. Hunt (2010), Comparing radiosonde and COSMIC atmospheric profile data to quantify differences among radiosonde types and the effects of imperfect collocation on comparison statistics, J. Geophys. Res., 115, D23104, doi:10.1029/2010JD014457.

Takayabu, Y. N. (1994), Large-scale cloud disturbances associated with equatorial waves Part II: Westward-propagating inertio-gravity waves, J. Meteorol. Soc. Jpn, 72, 451–465.

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, doi:10.1175/1520-0493(1996)124<1892:OOAQDW>2.0.CO;2.

Thompson, R. M., Jr., S. W. Payne, E. E. Recker, and R. J. Reed (1979), Structure and properties of synoptic-scale wave disturbances in the Intertropical Convergence Zone of the eastern Atlantic, J. Atmos. Sci., 36, 53–72, doi:10.1175/1520-0469(1979)036<0053:SAPOSS>2.0.CO;2.

Xie, F., D. L. Wu, C. O. Ao, A. J. Mannucci, and E. R. Kursinski (2012), Advances and limitations of atmospheric boundary layer observations with GPS occultation over southeast Pacific Ocean, *Atmos. Chem. Phys.*, *12*, 903–918, doi:10.5194/acp-12-903-2012.

Yanai, M., S. Esbensen, and J.–. H. Chu (1973), Determination of bulk properties of tropical cloud clusters from large-scale heat and moisture budgets, J. Atmos. Sci., 30, 611–627, doi:10.1175/1520-0469(1973)030<0611:DOBPOT>2.0.CO;2.

Zelinka, M. D., and D. L. Hartmann (2009), Response of humidity and clouds to tropical deep convection, J. Clim., 22, 2389–2404, doi:10.1175/2008JCLI2452.1.

Zipser, E. J. (1977), Mesoscale and convective-scale downdrafts as distinct components of a squall-line structure, *Mon. Weather Rev.*, 105, 1568–1589, doi:10.1175/1520-0493(1977)105<1568:MACDAD>2.0.CO;2.

Hertenstein, R. F. A., and W. H. Schubert (1991), Potential vorticity anomalies associated with squall lines, Mon. Weather Rev., 119, 1663–1672, doi:10.1175/1520-0493(1991)119<1663:PVAAWS>2.0.CO:2.