BNL-113652-2017-JA

On the Controls of Daytime Precipitation in the Amazonian Dry Season

Virendra P. Ghatе and Pavlos Kollias

Accepted for publication in Journal of Hydrometeorology

December 2016

Environmental & Climate Science Dept.

Brookhaven National Laboratory

**U.S. Department of Energy
USDOE Office of Science (SC),
Biological and Environmental Research (BER) (SC-23)**

Notice: This manuscript has been authored by employees of Brookhaven Science Associates, LLC under Contract No. DE-SC0012704 with the U.S. Department of Energy. The publisher by accepting the manuscript for publication acknowledges that the United States Government retains a non-exclusive, paid-up, irrevocable, world-wide license to publish or reproduce the published form of this manuscript, or allow others to do so, for United States Government purposes.

DISCLAIMER

This report was prepared as an account of work sponsored by an agency of the United States Government. Neither the United States Government nor any agency thereof, nor any of their employees, nor any of their contractors, subcontractors, or their employees, makes any warranty, express or implied, or assumes any legal liability or responsibility for the accuracy, completeness, or any third party's use or the results of such use of any information, apparatus, product, or process disclosed, or represents that its use would not infringe privately owned rights. Reference herein to any specific commercial product, process, or service by trade name, trademark, manufacturer, or otherwise, does not necessarily constitute or imply its endorsement, recommendation, or favoring by the United States Government or any agency thereof or its contractors or subcontractors. The views and opinions of authors expressed herein do not necessarily state or reflect those of the United States Government or any agency thereof.

1 **On the Controls of Daytime Precipitation in the Amazonian Dry Season**

2
3
4 Virendra P. Ghatel¹ and Pavlos Kollias^{2,3}

5 1. Argonne National Laboratory; 2. Brookhaven National Laboratory,

6 3. State University of New York, Stony Brook

7
8
9
10
11
12 **Corresponding Author**

13 Virendra P. Ghatel

14 9700 South Cass avenue,

15 Lemont IL 60439

16 Email: vghate@anl.gov

17 Tel: 630-252-1609

ABSTRACT

The Amazon plays an important role in the global energy and hydrological budgets. The precipitation during the dry season (June to September) plays a critical role in maintaining the extent of the rainforest. The deployment of the Atmospheric Radiation Measurements (ARM)'s first Mobile Facility (AMF-1) in the context of the GO Amazon field campaign at Manacapuru, Brazil provided comprehensive measurements of surface, cloud, precipitation, radiation and thermodynamic properties for two complete dry seasons (2014 and 2015). The precipitation events occurring during the nighttime were associated with propagating storm systems (non-local effects), while the daytime precipitation events were primarily a result of local land-atmosphere interactions. During the two dry seasons, precipitation was recorded at the surface on 106 (43%) days from 158 rain events with 82 daytime precipitation events occurring on 64 days (60.37%). Detailed comparisons between the diurnal cycles of surface and profile properties between days with and without daytime precipitation suggested the increased moisture at low- and mid- levels to be responsible for lowering the lifting condensation level, reducing convective inhibition and entrainment, and thus, triggering the transition from shallow to deep convection. Although the monthly accumulated rainfall decreased during progression of dry season, the contribution of daytime precipitation to it increased, suggesting the decrease to be mainly due to reduction in propagating squall lines. The control of daytime precipitation during the dry season on large-scale moisture advection above the boundary layer and the total rainfall on propagating squall lines suggests that coarse resolution models should be able to accurately simulate the dry season precipitation over the Amazon basin.

1. Introduction

The Amazon forest stores substantial amount of carbon in its biomass (Saatchi et al. 2011; Gloor et al. 2012). Precipitation plays a regulating role of the Amazonian rainforest canopy at various timescales (Aragao et al. 2008; Phillips et al. 2009; Toomey et al. 2011; Saatchi et al. 2013; etc.). The region is a part of the South American Monsoon System (SAMS) that is characterized by two distinct seasons, a wet season from October to May and a dry season from June to September, with regional differences in the duration and onset of the two seasons (Liebmann and Mechoso, 2011; Raia and Cavalcanti, 2008). Rainfall occurs during both wet and dry seasons with the rainfall in the dry season being crucial for maintaining the rain forest (Toomey et al. 2011; Lewis et al. 2011). Most of the Global Climate Models (GCM) simulations part of the Intergovernmental Panel for Climate Change (IPCC) fifth assessment report (AR5) forecast the Amazonian forest to go through severe changes in meteorology and climate in the future, with reduction in its extent primarily due to a drier and longer dry season (Cook et al. 2012; Joetzjer et al. 2013). The rainfall in the dry season is affected by many factors including the Sea Surface Temperatures (SSTs) in the tropical Atlantic and Pacific oceans (Grimm, 2003, 2004; Misra, 2008; Fernandes et al. 2015; Yoon, 2016), moisture recycling due to evaporation (Juarez et al. 2007; Spracklen et al. 2012), and local effects (Li and Fu, 2004; Tanaka et al. 2014; dos Santos et al. 2014; Fitzjarrald et al. 2008).

A number of observational (Harriss et al. 1988; Kaufman et al. 1998; Silva Dias et al. 2002) and modeling (Werth and Avisaar, 2002; Anber et al. 2015; Harper et al. 2014; de Goncalves et al. 2013) studies have been conducted with a goal of improving our understanding of the controls of precipitation and the hydrological cycle in the Amazonia. Despite these efforts, considerable disagreement exists in the present climate GCM simulation of the precipitation in

the Amazonia region. This is partly due to our limited understanding of the processes responsible for causing precipitation during the dry season (Malhi et al. 2009; Mehran et al. 2014). Precipitation occurs at all hours during the dry season with the number of precipitation events decreasing during the progression of the dry season (Fig. 1 and Tanaka et al. 2014). The nighttime precipitation is due to the passage of storm systems that originate at a different region, and hence is due to nonlocal effects (Rickenbach, 2004). However, the daytime precipitation occurs due to local land-atmosphere interactions (Fitzjarrald et al. 2008). Here, observations collected during the Green Ocean Amazon (GO Amazon) field campaign are used to study the controls of daytime (local) precipitation during the Amazon dry season. In particular, the presented study aims to address the following questions: 1) Which factors control the daytime transition from shallow (non-precipitating) to deep (precipitating) convection? and 2) What causes the number of precipitation events to decrease as the dry season progresses? Hence, although the study reports the characteristics (amount, duration, etc) of precipitation events, the focus is on the factors responsible for forming precipitation producing clouds.

To answer the above questions, we have used the data collected by the Atmospheric Radiation Measurement (ARM)'s first Mobile Facility (AMF-1) during the GO Amazon field campaign (<http://www.arm.gov/sites/amf/mao/>). Data collected during the 2014 and 2015 dry season (June to September) are used in this study and are described in section 2. First, the mean diurnal cycle of surface meteorology, thermodynamics, and clouds during the dry season is described in section 3. Subsequently, the differences in the diurnal cycle of days with daytime precipitation and days with cumulus clouds (non-precipitating) are described in section 4. The progression of the daily averaged values of few key parameters during the dry season is studied

in section 5 in the context of daytime precipitation events. The article is concluded with a summary, conclusion and discussion section.

2. Data and Instrumentation

The AMF-1 was deployed at Manacapuru, Brazil (3.2°S, 60.5°W, 50 m) for the GO Amazon field campaign from January, 2014 to December, 2015. In this study, the dry season is defined from June 1 to September 30 similar to Harper et al. (2014). The AMF-1 instrumentation includes a vertically pointing Doppler cloud radar, lidars, and radiometers. Below we have only described the instruments data from whom were used in this study. A detailed description of the AMF-1 instrumentation can be found in Mather and Voyles (2013).

A vertically pointing 94-GHz Doppler cloud radar known as the W-band ARM Cloud Radar (WACR) was part of the AMF-1. The WACR reports the full Doppler spectrum and its first three moments at 2 second temporal and 30 m range resolution. Also part of the AMF-1 was a laser ceilometer that operated at 905 nm wavelength and reported the values of first three cloud base heights at 15 second temporal and 30 m range resolution. The WACR and ceilometer data were combined to produce estimates of noise filtered radar reflectivity, cloud boundaries, and cloud fraction as described by Kollias et al. (2009) at a 5 second temporal and 30 m range resolution. A microwave radiometer (MWR) reported the sky brightness temperatures at 23.8 GHz and 31.4 GHz, from which the column integrated water vapor (IWV) and Liquid water path (LWP) were retrieved at a 20-second resolution. Also part of the AMF-1 instrumentation was a surface meteorological station that reported temperature, moisture, pressure, and winds every minute. The minute values of rain rate as reported by the Present Weather Detector (PWD) system that operates at 875 nm and has sensitivity of $0.05 \text{ mm hour}^{-1}$ were used in this study. The

height of the lifting condensation level (LCL) was calculated using the formulation provided by Bolton (1980) from the meteorological station data. Half-hourly values of surface sensible heat flux (SHF), latent heat flux (LHF), and surface friction velocity (U^*) were calculated using the data collected by the Eddy Correlation (ECOR) flux measurement system. A Surface Energy Balance System (SEBS) reported the ground heat flux and the net ecosystem exchange (SHF+LHF) on half-hourly timescales. Due to persistent failure in a component of the ECOR LHF sensor, the ECOR reported LHF measurements were intermittent. Additionally, at times due to lower wind speeds at night ($<1 \text{ m s}^{-1}$), the half hour period used to calculate the eddy covariances was not sufficient to capture the large eddies within the boundary layer. Hence, a second estimate of LHF is calculated from SEBS measurements by subtracting the ECOR reported SHF, from the SEBS reported net ecosystem exchange. Hereafter, this flux is referred as the SEBS LHF. The differences in the values of SEBS LHF and ECOR LHF are caused by energy imbalance (Wohlfahrt and Widmoser, 2013; Burba, 2013), with SEBS LHF being mostly higher than ECOR LHF. It is beyond the scope of this article to probe the causes of this difference, however we believe the SEBS LHF to be much closer to the real exchange of moisture between the surface and the atmosphere than the ECOR LHF (personal communication with David Cook).

Minute values of upwelling and downwelling radiation in the shortwave and longwave spectrum were measured by the radiation platform. Balloon borne radiosondes were launched every 6 hours (06, 12, 18, 00 UTC) that measured the tropospheric temperature, moisture, pressure, and wind fields. The local time was four hours behind the UTC. The Convective Inhibition (CIN) and Convective Available Potential Energy (CAPE) was calculated from the profiles of equivalent potential temperature (θ_e) and saturation equivalent potential temperature

(θ_{es}) using the method described by Kollias et al. (2009). The 50-m value of θ_e was used as a proxy for the surface value to avoid surface layer effects, and the level with θ_{es} value equal to that was termed as the level of free convection (LFC). The lower tropospheric stability was assessed using the mean value of the lapse rate of potential temperature $\frac{\partial \theta}{\partial z}$ and mixing ratio $\frac{\partial q}{\partial z}$ between 1 and 3 km, similar to Gentine et al. (2013). The convective triggering potential (CTP), the low-level humidity index (HI_{low}), and the inversion Bowen ratio (B_{inv}) were calculated using the definition by Findell and Eltahir (2003), with CTP and B_{inv} calculated for 1-3 km layer and HI_{low} calculated for the 500-1500 m layer.

The horizontal winds were retrieved every 15-min from the Plan Position Indicator (PPI) scans performed by the Doppler Lidar (DL) that operated at 1.5 micrometer wavelength and 30 m range resolution. The hourly reanalysis output from the European Center for Medium range Weather Forecasting (ECMWF) model operating at 137 levels and 0.56° grid resolution is used to characterize the large-scale environment. In particular, the ECMWF reported large-scale vertical velocity and the horizontal advection tendency of temperature and moisture are used in this study. The model output reported the total and physical tendency of temperature and moisture every hour, from which the horizontal advection terms of temperature and moisture were calculated. The hourly averaged values of the aforementioned parameters are used to calculate the diurnal cycle (section 3 and 4) and the daily mean values (section 5).

3. Average Diurnal Cycle

The averaged diurnal cycle was calculated using hourly values of respective parameters, with the hourly averages calculated for hours at the beginning of the hour e.g. zeroth hour

corresponds to values between midnight and 1 am, and 11th hour corresponding to values between 11 am and noon.

The average diurnal cycle of the surface meteorological parameters during the dry season is shown in Figure 2. The average surface air temperature was minimum at 6 am followed by a rapid increase until about 11 am and then remained constant until 4 pm. The range of the temperature as denoted by the standard deviation bars suggests greater fluctuations in daytime temperatures than in the nighttime temperatures. The averaged water vapor mixing ratio at the surface was also minimum at 6 am, but was highest at 8am followed by another local minimum at 3 pm and maximum at 7 pm. It is difficult to explain the cause of the two peaks (and minimum) in the diurnal cycle of the mixing ratio from these observations. The latter might be due to the loss of mixed layer water vapor to the free troposphere and deepening of the mixed layer thereby redistributing the vapor over a deeper layer, with the total loss not being compensated by the LHF. The winds at the surface were southerly and weak ($<1 \text{ m s}^{-1}$) during nighttime, and westerly during the daytime with peak speed of 3 m s^{-1} at noon. Similar to Oliveira and Fitzjarrald (1993) and Oliveira and Fitzjarrald (1994), the deviations of hourly averaged surface winds from the diurnal mean were termed as the thermal circulations. The polar plot of the averaged thermal circulation (not shown) exhibited an elliptical trajectory that rotated clockwise with time consistent with previous observations. The rain rate did not exhibit a distinct diurnal cycle as other parameters with the standard deviation values much greater than the means, suggesting a large variability in the amount and timing of rain events. The surface pressure at the AMF-1 also exhibited a distinct diurnal cycle with minimums at 3 am and 4 pm, and maximums at 9 am and 11 pm. Broadly synthesizing the mean diurnal cycle of meteorological parameters at Manacapuru during the dry season follows the classic summertime

diurnal cycle of continental boundary layer shown by Stull (1988) and the diurnal cycles reported in that region (Tanaka et al. 2014 etc.). The diurnal changes in the winds and pressure are consistent with the previous findings of a river breeze resulting due to temperature and pressure contrast between the land surface and the river (Oliveira and Fitzjarrald, 1993).

The mean diurnal cycle of SHF, LHF and U^* during the dry season is shown in Figure 3. The SHF and LHF were near zero at night and exhibited maximum values around noon (ECOR SHF $\sim 100 \text{ Wm}^{-2}$, ECOR LHF $\sim 250 \text{ Wm}^{-2}$, and SEBS LHF $\sim 400 \text{ Wm}^{-2}$). The higher value of LHF than SHF are consistent with water saturated land conditions despite the dry season. The standard error of the mean of ECOR SHF is higher than both the ECOR LHF and SEBS LHF implying relative changes in SHF to be greater than those in the LHF. The surface friction velocity (U^*) is very low ($< 0.1 \text{ m s}^{-1}$) at nighttime with a peak of $\sim 0.25 \text{ m s}^{-1}$ at noon. The surface convective velocity scale (w^*) exhibited a diurnal cycle similar to the SHF and LHF with a peak of $\sim 1.4 \text{ m s}^{-1}$ at noon (not shown). The significantly lower value of U^* compared to w^* suggests the role of shear in maintaining turbulence within the boundary layer to be smaller than that of the surface turbulent fluxes.

The mean diurnal cycle of downwelling shortwave radiation (SWD), upwelling shortwave radiation (SWU), downwelling longwave radiation (LWD) and upwelling longwave radiation (LWU) at the surface is shown in Figure 4. The downwelling and upwelling shortwave radiation follows the solar cycle with a Gaussian shape and peak around noon. The average shortwave albedo of the land is about 0.20 with standard deviation of 0.02. Broken cloud conditions are responsible for the observed SWD standard deviation, however, no significant departure from the Gaussian shape is observed. The diurnal cycle of LWD loosely follows that of surface air temperature with an almost constant value at night $\sim 400 \text{ Wm}^{-2}$ with a rapid increase

to a peak value of $\sim 425 \text{ Wm}^{-2}$ near noon, followed by a gradual decrease. This is expected as the cloud coverage is low ($<25\%$) and for the same amount of water vapor, an 8 K increase in the temperature will roughly cause an increase in the longwave flux by $\sim 44 \text{ Wm}^{-2}$ (11%). The diurnal cycle of LWU follows the solar cycle with values of $\sim 440 \text{ Wm}^{-2}$ at night, a peak of $\sim 525 \text{ Wm}^{-2}$ around noon followed by a gradual decrease. The net longwave flux at the surface is always negative denoting transfer of heat from the surface to the atmosphere. This is consistent with the low negative values of SHF and LHF during nighttime. The LWP and IWV as reported by the MWR did not exhibit a distinct average diurnal cycle (Figure 4). Although the averaged LWP exhibited a local peaked near noon, the standard error of the mean was much higher during that time. The IWV was $\sim 4.6 \text{ cm}$ during the entire day with standard deviation of $\sim 0.6 \text{ cm}$.

Radiosondes were launched four times a day at 2 am, 8 am, 2 pm, and 8 pm local time (Fig 5). Noticeably, apart the near surface values, the averaged profiles did not exhibit significant changes with height. On average, the potential temperature increased with height and the mixing ratio decreased with height in all soundings suggesting the atmosphere to be statically stable, except for a weak mixed layer in the 1 pm averaged sounding. The averaged wind speed increased almost linearly with height from the surface to 2 km in all of the soundings, and then remained almost constant. The wind direction was southwesterly in the lower 2 km in the 2 am and 8 am sounding, and were westerly in the 2 pm and 8 pm sounding. Above 2 km the winds were westerly at all times. Although similar to the J1 type profile shown by Oliveira and Fitzjarrald (1993) a vertical directional wind shear was observed below 1 km, unlike their observations the maximum of the wind speed was at a higher altitude ($\sim 2 \text{ km}$) and did not change with time. The profiles of wind speed and wind direction as reported by the Doppler Lidar also did not show any presence of nocturnal low level jet. The mean profiles of equivalent

potential temperature (θ_e) and saturation equivalent potential temperature (θ_{es}) suggested CAPE and CIN to be present during all sounding launch times, with lowest CIN values at 8 am and highest CAPE values at 2 pm local time. The average CAPE and CIN at 8 am were 758.31 J kg⁻¹ and -88.66 J kg⁻¹, while the same at 2 pm were 681.40 J kg⁻¹ and -136.91 J kg⁻¹ respectively. The average diurnal cycle of wind speed and wind direction as reported the Doppler Lidar (not shown) was consistent with that from the surface meteorological station and the soundings.

The average diurnal cycle of hydrometeor fraction, LCL and first cloud base height during the dry season is shown in Figure 6. The LCL was near surface during the nighttime and peaked to a value of 1065 m at 3 pm. The higher values of hydrometeor fraction (>20%) are seen for the low level clouds during the daytime and high level clouds during the later afternoon and at night. As the radar pulse of the WACR is susceptible to attenuation due to water vapor and liquid water, the deduced coverage of high-level clouds might be underestimated. Cirrus clouds are regularly observed over the Amazon with coverage of ~55% during the dry season (Gouveia et al. 2016). As data from Micro-Pulse Lidar (MPL) was not used in this study, and the presence of high-level clouds was only detected from the WACR, the reported coverage of high-level clouds represent a lower bound. However, the maximum hourly averaged rain rate observed at the surface during the study period was 4.94 mm hour⁻¹, with only 0.5% of the minute values higher than 5 mm hour⁻¹, suggesting the attenuation due to rainwater to be minimal. The close matching of LCL and first cloud base height to the higher coverage of low-level clouds between 9 am and 5 pm confirm the clouds to result from surface based convection rather than advection. The differences in the LCL and the cloud base height during the nighttime together with near-zero SHF and LHF suggest the boundary layer to be decoupled from the surface during these times, consistent with previous studies (Martin et al. 1988; Fitzjarrald and Moore, 1990). The

coverage of low level clouds is ~8% from 3-8 am and is discussed later in section 4. The high level clouds possibly are a result of deep convection around the ARM site consistent with a peak in late afternoon and ~10% coverage around midnight. As the tropical cirrus clouds typically last for 19-30 hours advecting over 600-1000 km (Luo and Rossow, 2004), it is plausible for the origin of these clouds to be at a location with conditions far different than those at the AMF-1.

The average diurnal cycle of large-scale vertical velocity and of the large-scale horizontal advective tendencies of temperature and moisture are shown in Figure 7. On average, nighttime horizontal advection is weak. During daytime (8 am to 4 pm) a shallow cold air advection confined below 900 hPa is observed. The depth of the cold air advection during the daytime is on average consistent with the depth of the westerly wind layer shown in the soundings (Figure 5) and the changes in the surface pressure caused by the thermal gradient between land and the river. The depth of this layer is consistent with that reported by the previous studies (Oliveira and Fitzjarrald, 1993). Above this shallow layer, the advection is warm throughout the day. On average around local noon, a moist advection is present from 950 hPa to 750 hPa. Consistent with the tropical location, the large-scale vertical velocity follows the solar cycle and is weak during the nighttime with high negative values (upward motion) from 8 am to 4 pm and a peak at around noon at 950 hPa. It is difficult to explain the positive moisture advection around noon that coincides with the largescale upward motion. The low-level cloudiness simulated by the ECMWF model (not shown) peaks between 8 am and noon, and hence is inconsistent with the peak in the noontime peak in the largescale vertical velocity.

4. Diurnal Cycle during Cumulus and Precipitation Days

Cloudy days are classified with respect to their potential to produce measurable precipitation at the ground in two categories: “cumulus” if no precipitation is measured and “precipitating” if precipitation is measured with rain rates higher than $0.05 \text{ mm hour}^{-1}$. The classification follows that proposed by Zhang and Klein (2010 and 2013). A total of 244 days were analyzed (122 over two years). Low level clouds were omnipresent throughout the two dry seasons and only 4 days (2%) are classified as “clear sky” with maximum ceilometer cloud fraction lower than 5%. Although the ceilometer recorded cloud fraction was less than 5% on these days, the Total Sky Imager (TSI) reported presence of shallow clouds at the ARM site. Precipitation was recorded at the surface on 106 days (43.44%). The precipitation days were further classified as daytime [9am – 9pm], nighttime [10 pm - 9 am] or mixed (daytime and nighttime) based on the time of occurrence of the precipitation events as captured by the surface rain gage. Based on this classification, 64 days were classified as daytime precipitation, 16 days were classified as nighttime precipitation, and 26 days were classified as mixed. As reported by previous studies (e.g. Tanaka et al. 2014; Pereira Filho et al. 2002 etc.), the nighttime precipitation occurring over that region is mostly due to propagating storm system and not produced by local effects. Additionally, the wind direction was northerly during the nighttime events (not shown), while it was similar to the average (Fig 2) during daytime events. Here, we have focused on the causes of daytime precipitation that might have local controlling factors. Therefore, the term daytime is omitted and we focus only the mean diurnal cycle of cumulus only days (139 days) and precipitation days (64 days). To gain insights on the continuity of convection, we also contrasted the diurnal cycles of cumulus and precipitation days that were preceded by cumulus days. Cumulus days that were preceded by cumulus days were termed as CuToCu and the daytime precipitation days that were preceded by cumulus days were termed as

CuToPrecip. Total of 89 CuToCu days and 30 CuToPrecip days were identified. The conclusions from this comparison between the averaged diurnal cycle of CuToCu and CuToPrecip days were similar to those discussed below, and the comparison plots and a table are provided as supplemental material. As the start times of most of the daytime rain events was around noon, we have only discussed variables that exhibited differences before noon for the classification as it is not possible to distinguish if the afternoon observations are consequences of the rain or are characteristics of the environment responsible for producing the rain.

The averaged diurnal cycle of hydrometeor fraction for cumulus days and precipitation days along with that of LCL and first cloud base height is shown in Figure 8. The LCL was lower during precipitation days than that during cumulus days. Most of the precipitation occurred between 11 am and 5 pm and convective anvil outflow is observed over the site for the rest of the day. Most of the rainfall occurred between noon-4 pm during the daytime precipitation days. High-level cirrus clouds were also observed in later afternoon and the nighttime during cumulus days as well but with lower coverage and thicknesses. We suspect the cirrus clouds to be present from 6am-noon during precipitation days, as reliable lidar data is not available during high sun-angle conditions, and the sensitivity of the radar decreases at higher range gates due to the range correction term. The nighttime occurrence of low level clouds was higher during precipitation days than during cumulus days. This is broadly consistent with the hypothesis of Anber et al. (2015) regarding presence of low-level fog during precipitating days.

Surface meteorological parameters for cumulus and precipitation days are shown in Figure 9. The surface air temperature from 8 am till midnight was lower during the precipitation days than that during cumulus days. The water vapor mixing ratio at the surface was higher during precipitation days than that during cumulus days from 9 am to 6 pm. The differences in

temperature and mixing ratio are significant even before the occurrence of the precipitation events, suggesting different surface layer thermodynamics on precipitation days than cumulus days. The wind speeds were lower during precipitation days from 8 am to 5 pm than during cumulus days. The winds were southerly during nighttime and westerly during the daytime during each distinction. The polar plot of the thermal wind (not shown) exhibited a clockwise rotation for both distinction, with slightly lower values of the thermal wind during the precipitation days. This is consistent with similar diurnal cycle of surface pressure for the distinction with slightly higher values during the precipitation days. The similar diurnal cycles of winds and pressure for the distinction suggest the effect of river breeze to be similar for the distinction.

The average diurnal cycles of surface turbulent fluxes and friction velocity during cumulus and precipitation days are shown in Figure 10. The SHF and LHF were lower during precipitation days than that during the cumulus days from 8 am to 5 pm. The differences in the LHF and SHF occurred before the occurrence of precipitation, thus, it is difficult to speculate if it is caused by the precipitation and associated cloudiness. The surface friction velocity (U^*) was lower during precipitation, although, a 0.05 m s^{-1} (10%) reduction in the value of U^* seems insignificant to cause substantial changes in the boundary layer turbulence. The average LCL peaked to the value of $\sim 800 \text{ m}$ at 11 am on precipitation days, while it peaked later at 3 pm to a value of $\sim 1300 \text{ m}$ during cumulus days. This is consistent with the presence of lower near-surface air temperature and higher mixing ratio during the precipitation days.

The mean diurnal cycle of radiative fluxes, LWP, and IWV for precipitation and cumulus days is shown in Figure 11. The SWD is substantially lower during precipitation days, due to the higher cloud cover (Figure 8). The higher cloud cover during precipitation days is also consistent

with the higher values of LWD from midnight until local noon. Consistent with lower surface temperature, the LWU was lower during the precipitation days. As expected, due to the higher cloud cover and cloud thickness, the LWP was higher during precipitation days than that during cumulus days. The values of IWV were substantially higher during all hours during precipitation days. The higher statistical significance of these differences along with higher values of relative changes make the IWV to be the most robust parameter to differ for the distinction among the analyzed variables.

The profile of potential temperature at 8 am did not exhibit any significant differences between cumulus and precipitation days (Fig. 12). A mixed layer extending from surface to 1 km is clearly visible in the averaged 2 pm sounding for the cumulus days, while for the precipitation days the profile of potential temperature was statically stable in the entire lower troposphere. The statically stable profile of potential temperature might be due to the evaporatively driven downdrafts part of the precipitation events (Betts et al. 2002). The mixing ratio was higher during precipitation days at all heights in the 8 am and 2 pm sounding, except for near-similar values from 1-2 km in the 2 pm sounding. The lower values of mixing ratio and slightly higher values of temperature during the cumulus days resulted in lower values of equivalent potential temperature and higher values of saturation equivalent potential temperature during cumulus days than during precipitation days. This essentially resulted in lower values of CIN during the precipitation days than during cumulus days. The average CIN at 8 am was -98.94 J kg^{-1} during cumulus days and -76.40 J kg^{-1} during the precipitation days, while the average CAPE was 824.93 J kg^{-1} during cumulus days and 846.52 J kg^{-1} during the precipitation days. In summary, the lower surface air temperature and higher mixing ratio at all levels during precipitation days

than those during cumulus days, collectively result in significantly lower CIN and slightly higher CAPE during precipitation days than those during cumulus days.

The daytime cold air advection below 900 hPa is slightly weaker and shallower during precipitation days than during cumulus days (Fig. 13), consistent with lower temperature over land (AMF-1) reducing the effect of the river breeze. The warm advection around 200 hPa is due to 16 of the daytime precipitation events being preceded by nighttime events or mixed events. This upper level heating is absent in the CuToCu and CuToPrecip comparisons available in the supplemental material. Below 700 hPa level (~3 km), the moisture advection was “spotty” and positive (moist advection) during both cumulus and precipitation days at all hours. While above 700 hPa up to 500 hPa, the moisture advection was negative (dry advection) during cumulus days with the dry layer extending to ~900 hPa during the afternoon hours. The largescale vertical velocity was similar for the distinction except a bit stronger with a higher extent (~400 hPa) during precipitation days than that during the cumulus days (~700 hPa). Broadly synthesizing, the plot suggests the advection of moisture above the boundary layer to be the key difference between the cumulus and precipitation days, which is in agreement with previous studies (e.g. Zhang and Klein, 2008).

To assess the differences in key parameters for precipitating and non-precipitating days, their averaged values for the entire dry season, for cumulus only days, and for daytime precipitation days are listed in Table 1. The t-value of the statistical significance assuming a two-tailed Gaussian distribution of the differences in the values of cumulus and precipitation days is also listed in Table 1. As anticipated from our analysis, statistically significant differences are present in the surface air temperature, moisture, and the LCL between the cumulus and daytime precipitation days. The CIN was significantly lower during precipitation days, with insignificant

changes in CAPE between cumulus and daytime precipitation days. To further explore the higher IWV observed during the precipitation days than during cumulus days, we calculated the integrated water vapor amount between surface and 1 km (WV_{0-1km}), between 1 and 3 km (WV_{1-3km}), and between 3 to 6 km (WV_{3-6km}) as reported by the soundings. The water vapor in all the three layers was higher during the precipitation days than during cumulus days, suggesting a relatively same amount of increase or decrease in the lower tropospheric water vapor as the total. On average, the horizontal moisture advection was negative during the cumulus days and positive during the precipitation days with large variations. The commonly used indices to predict occurrence of precipitation were calculated from the radiosondes launched at 8 am. The CTP did not exhibit any changes between the cumulus and precipitation days, while the HI_{low} did, further reinforcing the role of moisture in determining deep convection. The inversion Bowen ratio (B_{inv}) was lower during the cumulus days than during the precipitation days, due to the higher lapse rate of moisture during the cumulus days than during precipitation days. The lapse rate of potential temperature did not exhibit any significant differences for the distinction, while the lapse rate of water vapor mixing ratio was significantly lower during precipitation days compared to the cumulus days.

5. Variation of Key Parameters through the dry season

The comparison of the diurnal cycle of surface and profile properties indicated that IWV, horizontal moisture advection above the boundary layer (Q_{ten}), SHF, LCL and CIN to differ substantially between cumulus and precipitation days. In addition to changes in these on precipitation days, their temporal variation through the 2014 and 2015 dry seasons is examined in an attempt to interpret the decrease in precipitation events during the progression of the dry

411 season (Fig. 14). The daily values shown of SHF, LCL, and Q_{ten} in Fig 14 are averages between
 412 8 am to 6 pm and the Q_{ten} is averaged from 700-600 hPa (3-4.5km). The daily averaged column
 413 water vapor varied between 3-6 cm during the dry seasons, with large variations (>2 cm) around
 414 day 15 and 78 in 2014, and day 70 and 100 in 2015. The precipitation events mostly coincided
 415 with the local maximum of IWV throughout the two dry seasons, with some exceptions like day
 416 45 and 90 of 2015. The SHF remained fairly low with low variability in 2014, while it increased
 417 during the progression of the dry season in 2015. Most of the precipitation events coincided with
 418 SHF minimum. Similar to IWV, relatively high ($>30 \text{ Wm}^{-2}$) changes in SHF can be seen to occur
 419 over a span of few days like day 85-90 and 96-102 in the year 2015. The daytime average LCL
 420 varied between ~ 200 m to ~ 2000 m during the course of the dry seasons, with greater variations
 421 in 2015 than in 2014. The variations in LCL were largely due to changes in the surface water
 422 vapor mixing ratio rather than the changes in temperature, and correspond well with the changes
 423 in the IWV. The averaged daytime horizontal advection of water vapor between 600-700 hPa
 424 was weakly negative (dry advection) with several high positive values. Contrary to our
 425 expectations from the diurnal cycle analysis, several precipitation events occurred during dry
 426 advection conditions (e.g. day 105 in 2014), and precipitation not occurring on some of the days
 427 that had high positive values of moisture advection (e.g. day 36 and 90 in 2015). This suggests
 428 that the moisture advection is a secondary parameter determining precipitation. The CIN
 429 calculated from the sounding launched at 8 am varied from about -50 J kg^{-1} at the beginning of
 430 the dry season, to excess of -250 J kg^{-1} towards the end of the dry season. Most of the
 431 precipitation events coincided with local minimum values of CIN. Collectively these
 432 observations suggest the column water vapor to decrease, and the LCL, SHF and CIN to increase
 433 during the progression of the dry season. Most of the precipitation events coincided with the

local maximum of the IWV and minimum of LCL, SHF and CIN. The changes in these parameters are not gradual during the dry season, with large fluctuations occurring over 3-5 day span embedded within the gradual change in them. Most of all, these observations suggest that the changes occurring during the dry season to vary greatly from year-to-year as highlighted by SHF during the two years studied here.

The rain recurrence interval (t_r) was calculated for all precipitation events and for the daytime precipitation events. Similar to Fritzjarrald et al. (2008), we have defined t_r as time interval between the rain events, with a rain event defined as a continuous period when the hourly averaged rain rate was greater than zero. Rain was recorded at the surface on 106 days through 158 rain events (Table 2), denoting 1.49 rain events per rainy day. Although the number of rainy days were almost equal in 2014 and 2015, there were more rain events in 2014 than 2015. The mean t_r for the two dry seasons was 1.42 day, with slightly lower t_r in 2014 (1.36 day) than in 2015 (1.48). About 40% of the rain events had t_r greater than 1 day, with 4.5% having t_r greater than 5 day. The rain recurrence time was ~ 1 day in June and July, and greater than 2 day during August and September. The decrease in t_r from August to September was observed during both years, and is due to multiple precipitation events occurring successively during the first half of September during both years (Figure 1). About 25 rain events occurred in June and July and about 15 in August and September. Although the number of rain events showed the two-month cycle, the total accumulated rain for each month showed a gradual decrease from June to September during both years.

As we have focused on daytime precipitation events, similar to t_r we also calculated recurrence time of daytime precipitation events (t_{dr}). Total of 82 daytime rain events were observed on the 64 days, denoting 1.28 rain events per daytime precipitation day (Table 3). The

average t_{dr} was 2.76 day with slightly higher t_{dr} in 2015 (2.95) than in 2014 (2.62). The number of daytime precipitation events with t_{dr} greater than 1 day, 5 day and 10 day were almost equal during the two years. The t_{dr} increased from June to August during both years and then decreased to a mean value of 2.40 day in September. As mentioned earlier, the decrease in t_{dr} in September is due to multiple daytime precipitation events occurring on consecutive days in the first half of September during both years. The percent of daytime precipitation events of the total rain events together with their share of the accumulated rainfall exhibited an increase each month through the progression of the dry season.

Collectively this analysis suggests the recurrence time of daytime precipitation events (t_{dr}) to be higher than that of all precipitation events (t_r). The relative change in t_r through the dry season was higher than that of t_{dr} . The number of total rain events and the total accumulated rainfall exhibited a general decrease each month during the progression of the dry season (Table 2), while the percent of daytime precipitation events and their share of accumulated rainfall exhibited a general increase. This suggests the principal decrease in the rainfall during the dry season to be primarily due to reduction in nighttime precipitation that is a result of passing squall lines.

6. Summary, Discussion and Conclusions

The coverage of the Amazonian rainforest critically depends on the dry season precipitation. The deployment of the AMF-1 for two years in the context of the GO Amazon field campaign provided comprehensive measurements of surface, clouds, radiation and thermodynamic properties for two complete dry seasons. Precipitation was recorded at the surface on 106 days (43%) during the two (2014 and 2015) dry seasons, with some amount of

precipitation recorded at all hours. The precipitation events occurring during the nighttime were associated with northerly advection corresponding to propagating storm systems (non-local effects). On the contrary, the daytime precipitation events were primarily observed between local noon and 4 pm and were a result of local land-atmosphere interactions. The nighttime low-level cloudiness was higher during the precipitations days than during the non-precipitating days. This resulted in lower values of net radiation at the surface that further translated to lower values of surface sensible heat flux. The higher low-level cloudiness was accompanied by higher atmospheric moisture that resulted in lower values of LCL and CIN during the precipitation days as compared to the non-precipitation days. The horizontal advection of moisture above the boundary layer as reported by the ECMWF reanalysis model was weakly positive during precipitation days and weakly negative during non-precipitation days. The tropical South Atlantic is the primary remote source of moisture for the Amazon basin during the austral winter (Drumond et al., 2014). Collectively, these results suggests the increased water vapor in the lower troposphere to be the primary factor responsible for triggering the transition from shallow to deep convection. This is achieved by lowering the LCL, the CIN and reducing the impact of dry entrainment on cloudy air-parcels. The study provides clear indications of a relationship between large-scale moisture advection and precipitation suggesting that accurate representation of these large-scale features could provide improve representation of the precipitation over the Amazon basin. As the conclusions about moisture advection above the boundary layer are based on ECMWF reanalysis model and not on observations, we deem the conclusions to be plausible albeit not exclusive.

Most of the daytime precipitation events occurred during days that exhibited higher moisture content, moist advection above the boundary layer, and lower values of SHF and CIN

before the beginning of the precipitation event. However, few precipitation events did occur during days that did not fit the above criteria. This suggests a complex interplay between the factors responsible for producing precipitation generating clouds, and possible factors other than considered here to have an impact on precipitation generation. The general lack of similarity between the evolutions of the variables during the two dry seasons (Figure 14), strongly warrants for long-term monitoring of meteorology of that region, and suggests the caution needed in drawing scientific conclusion from limited amount of data. Notably, the SHF varied little during the 2014 dry season, while it increased almost gradually during the second half of the 2015 dry season. The analysis of recurrence time of all rain events (t_r) and daytime rain events (t_{dr}) showed the t_r to increase through the dry season (except September) with comparatively lower variations in t_{dr} . The analysis also showed the percent contribution of accumulated rainfall from daytime rain events to the total accumulated rainfall to increase through the progression of the dry season. This together with the prediction by the IPCC GCM (Cook et al., 2012) of the dry season to get longer suggests the rainfall (if any) during this increased length to be primarily from the daytime precipitation events, which critically depend on the model's ability to simulate transition from shallow-to-deep convection and largescale moisture advection.

In addition to the daytime precipitation events, the region also receives precipitation from southward propagating storms during the dry season. 42 precipitation events (39%) occurred entirely or partly during the nighttime (9pm-9am), suggesting significant amount of rainfall from these systems. As these storms develop and propagate due to largescale weather patterns, factors other than local land-atmosphere interactions are also partly responsible for the dry season precipitation. However, the large spread and greater fluctuations over short periods in the column integrated water vapor suggest it to be the primary factor controlling meteorology and

precipitation in that region (Satyamurty et al. 2013; Sherwood et al. 2010). Although it is challenging to track atmospheric rivers, especially in a tropical environment, we anticipate such an effort to have a huge impact on our understanding of the meteorology and climate of the Amazon and other tropical regions (Schiro et al. 2016).

Acknowledgements:

We would like to thank Dr. Anthony D. Del Genio and Dr. Stephen W. Nesbitt for helpful discussions that led to significant improvement in the manuscript. This work was primarily supported by the U.S. Department of Energy's (DOE) Atmospheric System Research (ASR), an Office of Science, Office of Biological and Environmental Research (BER) program, under Contract DE-AC02-06CH11357 awarded to Argonne National Laboratory and Contract DE-SC00112704 awarded to Brookhaven National Laboratory. This research was also supported by the National Science Foundation (NSF) Grant AGS-1445831. All the data used in this study were obtained from the Atmospheric Radiation Measurement (ARM) Program sponsored by the U.S. Department of Energy, Office of Science, Office of Biological and Environmental Research, Climate and Environmental Sciences Division (CESD). We gratefully acknowledge the computing resources provided on Blues, a high-performance computing cluster operated by the Laboratory Computing Resource Center (LCRC) at the Argonne National Laboratory.

549
550
551
552
553
554
555
556
557
558
559
560
561
562
563
564
565
566
567
568
569
570
571

REFERENCES

Anber, U., P. Gentine, S. Wang, and A. H. Sobel, 2015: Fog and rain in the Amazon. *PNAS*, **112**, 11473–11477, doi: 10.1073/pnas.1505077112

Aragao, L., Y. Malhi, N. Barbier, A. Lima, Y. Shimabukuro, L. Anderson, and S. Saatchi, 2008: Interactions between rainfall, deforestation and fires during recent years in the Brazilian Amazonia. *Philos. Trans. Roy. Soc. London*, **B363**, 1779–1785

Betts, A. K., J. D. Fuentes, M. Garstang, and J. H. Ball, 2002: Surface diurnal cycle and boundary layer structure over Rondônia during the rainy season, *J. Geophys. Res.*, **107(D20)**, 8065, doi:10.1029/2001JD000356.

Bolton, D., 1980: The computation of equivalent potential temperature. *Mon. Weather. Rev.*, **108**, 1046-1053.

Burba, G., 2013. Eddy Covariance Method for Scientific, Industrial, Agricultural, and Regulatory Applications: A Field Book on Measuring Ecosystem Gas Exchange and Areal Emission Rates. LI-COR Biosciences, Lincoln, NE, USA, 331 pp

572 Cook, B., N. Zeng, and J-H Yoon, 2012: Will Amazonia Dry Out? Magnitude and Causes of
573 Change from IPCC Climate Model Projections. *Earth Interact.*, **16**, 1–27.

574 de Gonçalves LGG, et al., 2013: Overview of the Large-Scale Biosphere–Atmosphere
575 Experiment in Amazonia Data Model Intercomparison Project (LBA-DMIP). *Agric For*
576 *Meteorol*, **182-183**, 111–127

577 dos Santos, M. J., M. A. F. Silva Dias, and E. D. Freitas, 2014: Influence of local circulations on
578 wind, moisture, and precipitation close to Manaus City, Amazon Region, Brazil, *J.*
579 *Geophys. Res. Atmos.*, **119**, 13,233–13,249, doi:10.1002/2014JD021969.

580 Drumond, A., Marengo, J., Ambrizzi, T., Nieto, R., Moreira, L., and Gimeno, L., 2014: The role
581 of the Amazon Basin moisture in the atmospheric branch of the hydrological cycle: a
582 Lagrangian analysis, *Hydrol. Earth Syst. Sci.*, **18**, 2577-2598.

583 Fernandes, K., A. Giannini, L. Verchot, W. Baethgen, and M. Pinedo-Vasquez, 2015: Decadal
584 covariability of Atlantic SSTs and western Amazon dry-season hydroclimate in
585 observations and CMIP5 simulations, *Geophys. Res. Lett.*, **42**, 6793–6801,

586 Findell, K. L., and E. A. B. Eltahir, 2003: Atmospheric Controls on Soil Moisture–Boundary
587 Layer Interactions. Part I: Framework Development. *J. Hydrometeorol*, **4**, 552-569.

588 Fitzjarrald, D. R., and K. E. Moore, 1990: Mechanisms of nocturnal exchange between the rain
589 forest and the atmosphere, *J. Geophys. Res.*, **95(D10)**, 16839–16850.

590 Fitzjarrald, D. R., R. K. Sakai, O. L. L. Moraes, R. C. Oliveira, O. C. Acevedo, and M. J. C. T.
591 Beldini, 2008: Spatial and temporal rainfall variability near the Amazon-Tapajós
592 confluence, *J. Geophys. Res.*, **113**, G00B11, doi:10.1029/2007JG000596

593 Gentine, P., A. A. M. Holtslag, F. D'Andrea, and M. Ek, 2013: Surface and Atmospheric
594 Controls on the Onset of Moist Convection over Land. *J. Hydrometeorol*, **14**, 1443–1462.

595 Gouveia, D. A., B. Barja, H. M. J. Barbosa, T. Pauliquevis, and P. Artaxo, 2016: Optical and
 596 Geometrical Properties of Cirrus Clouds in Amazonia Derived From 1-year of Ground-
 597 based Lidar Measurements, *Atmos. Chem. Phys. Discuss.*, doi:10.5194/acp-2016-458, in
 598 review.

599 Gloor, M., Gatti, L., Brien, R., Feldpausch, T. R., Phillips, O. L., Miller, J., Ometto, J. P.,
 600 Rocha, H., Baker, T., de Jong, B., Houghton, R. A., Malhi, Y., Aragão, L. E. O. C.,
 601 Guyot, J.-L., Zhao, K., Jackson, R., Peylin, P., Sitch, S., Poulter, B., Lomas, M., Zaehle,
 602 S., Huntingford, C., Levy, P., and Lloyd, J., 2012: The carbon balance of South America:
 603 a review of the status, decadal trends and main determinants, *Biogeosciences*, **9**, 5407-
 604 5430, doi:10.5194/bg-9-5407-2012.

605 Grimm, A. M., 2003: The El Nino impact on the summer monsoon in Brazil: Regional processes
 606 versus remote influences. *J. Climate*, **16**, 263–280.

607 ———, 2004: How do La Niña events disturb the summer monsoon system in Brazil? *Climate*
 608 *Dyn.*, **22**, 123–138

609 Harper, A., I. T. Baker, A. S. Denning, D. A. Randall, D. Dazlich, and M. Branson, 2014: Impact
 610 of Evapotranspiration on Dry Season Climate in the Amazon Forest. *J. Climate*, **27**, 574–
 611 591.

612 Harriss, R. C., et al., 1988: The Amazon Boundary Layer Experiment (ABLE 2A): dry season
 613 1985, *J. Geophys. Res.*, **93(D2)**, 1351–1360, doi:10.1029/JD093iD02p01351.

614 Joetzjer, E., H. Douville, C. Delire, P. Ciais, 2013: Present-day and future Amazonian
 615 precipitation in global climate models: CMIP5 versus CMIP3. *Climate Dynamics*, **41**,
 616 2921-2936.

617 Juarez, R. I. N., M. G. Hodnett, R. Fu, M. Goulden, and C. von Randow 2007: Control of dry
 618 season evapotranspiration over Amazonian forest as inferred from observations at a
 619 southern Amazon forest site, *J. Clim.*, **20**, 2827–2839.

620 Kaufman, Y. J., P. V. Hobbs, V. W. J. H. Kirchhoff, P. Artaxo, L. A. Remer, B. N. Holben, M.
 621 D. King, D. E. Ward, E. M. Prins, K. M. Longo, L. F. Mattos, C. A. Nobre, J. D.
 622 Spinhirne, Q. Ji, A. M. Thompson, J. F. Gleason, S. A. Christopher and S.-C. Tsay, 1998:
 623 Smoke, Clouds, and Radiation-Brazil (SCAR-B) experiment, *J. Geophys. Res.*, **103**, D24,
 624 31783

625 Kollias, P., M. A. Miller, K. L. Johnson, M. P. Jensen, and D. T. Troyan, 2009: Cloud,
 626 thermodynamic, and precipitation observations in West Africa during 2006, *J. Geophys.*
 627 *Res.*, **114**, D00E08, doi:10.1029/2008JD010641.

628 Lewis S. L., P. M. Brando, O. L. Phillips, G. M. Van der Heijden, D. Nepstad, 2011: The 2010
 629 Amazon drought. *Science*, **331(6017)**, 554

630 Li, W., and R. Fu, 2004: Transition of the Large-Scale Atmospheric and Land Surface
 631 Conditions from the Dry to the Wet Season over Amazonia as Diagnosed by the ECMWF
 632 Re-Analysis. *J. Climate*, **17**, 2637–2651.

633 Liebmann, B., and C.R. Mechoso, 2011: The South American Monsoon System. The Global
 634 Monsoon System: Research and Forecast, 2nd Edition. Chang, C. -P., Y. Ding, N.-C.
 635 Lau, R. H. Johnson, B. Wang, and T. Yasunari, Eds., World Scientific Series on Asia-
 636 Pacific Weather and Climate, Vol. 5, World Scientific Publication Company, 608 pp.

637 Luo, Z., and W. B. Rossow, 2004: Characterizing tropical cirrus life cycle, evolution, and
 638 interaction with upper-tropospheric water vapor using Lagrangian trajectory analysis of
 639 satellite observations. *J. Climate*, **17**, 4541-4563, doi:10.1175/3222.1.

Malhi, Y., L. E. O. C. Aragao, D. Galbraith, C. Huntingford, R. Fisher, P. Zelazowski, S. Sitch, C. McSweeney, and P. Meir, 2009: Exploring the likelihood and mechanism of a climate-change-induced dieback of the Amazon rainforest. *Proc. Natl. Acad. Sci.*, **106**, 20610-20615.

Martin, C. L., D. Fitzjarrald, M. Garstang, A. P. Oliveira, S. Greco, and E. Browell, 1988: Structure and growth of the mixing layer over the Amazonian rain forest, *J. Geophys. Res.*, **93(D2)**, 1361–1375.

Mather, J. H., and J. W. Voyles, 2013: The ARM climate research facility: A review of structure and capabilities. *Bull. Amer. Meteorol. Soc.*, **94**, 377-392.

Mehran, A., A. AghaKouchak, and T. J. Phillips, 2014: Evaluation of CMIP5 continental precipitation simulations relative to satellite-based gauge-adjusted observations, *J. Geophys. Res. Atmos.*, **119**, 1695–1707, doi:10.1002/2013JD021152.

Misra, V., 2008: Coupled Air, Sea, and Land Interactions of the South American Monsoon. *J. Climate*, **21**, 6389–6403.

Oliveira A. P., and D. R. Fitzjarrald, 1993: The Amazon river breeze and the local boundary layer: I - observations. *Boundary Layer Meteorology*, **63**, 141–162.

Oliveira A. P., and D. R. Fitzjarrald, 1994: The Amazon river breeze and the local boundary layer: II – Linear analysis and modelling. *Boundary Layer Meteorology*, **67**, 75–96.

Pereira Filho, A. J., M. A. F. Silva Dias, R. I. Albrecht, L. G. P. Pereira, A. W. Gandu, O. Massambani, A. Tokay, and S. Rutledge, 2002: Multisensor analysis of a squall line in the Amazon Region, *J. Geophys. Res.*, **107(D20)**, 8084, doi:10.1029/2000JD000305.

Phillips, O., and coauthors, 2009: Drought sensitivity of the Amazon rainforest. *Science*, **323**, 1344-1347.

663 Raia, A., and I. F. A. Cavalcanti, 2008: The Life Cycle of the South American Monsoon
 664 System. *J. Climate*, **21**, 6227–6246.

665 Rickenbach, T. M., 2004: Nocturnal Cloud Systems and the Diurnal Variation of Clouds and
 666 Rainfall in Southwestern Amazonia. *Mon. Wea. Rev.*, **132**, 1201–1219, doi:
 667 10.1175/1520-0493(2004)132<1201:NCSATD>2.0.CO;2.

668 Saatchi, S. S., N. L. Harris, S. Brown, M. Lefsky, E. T. A. Mitchard, W. Salas, B. R. Zutta, W.
 669 Buermann, S. L. Lewis, S. Hagen, S. Petrova, L. White, M. Silman, and A. Morel, 2011:
 670 Benchmark map of forest carbon stocks in tropical regions across three continents. *Proc.*
 671 *Natl. Acad. Sci. USA*, **108**, 9899–9904.

672 Saatchi, S., S. Asefi-Najafabady, U. Malhi, L. E. O. C. Aragao, L. O. Anderson, R. B. Myneni,
 673 and R. Nemani, 2013: Persistent effects of severe drought on Amazonian forest canopy.
 674 *Proc. Natl. Acad. Sci. USA*, **110**, 565–570.

675 Satyamurty, P., C. P. W. da Costa, and A. O. Manzi, 2013: Moisture source for the Amazon
 676 basin: a study of contrasting years. *Theor. Appl. Climatol.*, **111**, 195–209.

677 Schiro, K. A., J. D. Neelin, D. K. Adams, and B. R. Lintner, 2016: Deep convection and column
 678 water vapor over tropical land vs tropical ocean: a comparison between the Amazon and
 679 the tropical western Pacific. *J. Atmos. Sci.*, doi:10.1175/JAS-D-16-0119.1, in press.

680 Silva Dias, M. A. F., et al. 2002: A case study of convective organization into precipitating lines
 681 in the southwest Amazon during the WETAMC and TRMM-LBA, *J. Geophys. Res.*, **107**,
 682 8078, doi:10.1029/2001JD000375.

683 Silva Dias, M. A. F., 2002: Cloud and rain processes in a biosphere-atmosphere interaction
 684 context in the Amazon Region, *J. of Geophys. Res.*, **107**, D20

685 Sherwood, S. C., R. Roca, T. M. Weckwerth, and N. G. Andronova, 2010: Tropospheric water
 686 vapor, convection, and climate, *Rev. Geophys.*, **48**, RG2001,
 687 doi:10.1029/2009RG000301.

688 Spracklen, D. V., S. R. Arnold, and C. M. Taylor, 2012: Observations of increased tropical
 689 rainfall preceded by air passage over forests. *Nature*, **489**, 282-185.

690 Stull, R. B., 1988: *An Introduction to Boundary Layer Meteorology*. Kluwer Academic, 666 pp

691 Tanaka, L. M. d. S., P. Satyamurty, and L. A. T. Machado, 2014: Diurnal variations of
 692 precipitation in central Amazon basin. *Int. J. Climatol.*, **34**, 3574-3584.

693 Toomey, M., D. A. Roberts, C. Still, M. L. Goulden, and J. P. McFadden, 2011: Remotely
 694 sensed heat anomalies linked with Amazonian forest biomass declines, *Geophys. Res.*
 695 *Lett.*, **38**, L19704, doi:10.1029/2011GL049041.

696 Werth, D., and R. Avissar, 2002: The local and global effects of Amazon deforestation, *J.*
 697 *Geophys. Res.*, **107(D20)**, 8087, doi:10.1029/2001JD000717.

698 Wohlfahrt, G., & Widmoser, P., 2013: Can an energy balance model provide additional
 699 constraints on how to close the energy imbalance? *Agricultural and Forest Meteorology*,
 700 **169**, 85–91.

701 Yoon. J-H., 2016: Multi-model analysis of the Atlantic influence on Southern Amazon rainfall.
 702 *Atmos. Sci. Let.* 17:10.1002/asl.2016.17.issue-2, 122-127.

703 Zhang, Y. and S. A. Klein, 2010: Mechanisms Affecting the Transition from Shallow to Deep
 704 Convection over Land: Inferences from Observations of the Diurnal Cycle Collected at
 705 the ARM Southern Great Plains Site. *J. Atmos. Sci.*, **67**, 2943–2959.

Zhang Y., and Stephen A. Klein, 2013: Factors Controlling the Vertical Extent of Fair-Weather Shallow Cumulus Clouds over Land: Investigation of Diurnal-Cycle Observations Collected at the ARM Southern Great Plains Site. *J. Atmos. Sci.*, **70**, 1297–1315.

Table 1: Mean and standard deviation of key parameters for the entire dry season, cumulus only days, and daytime precipitation days. The t-value denoting the statistical significance between the values cumulus and daytime precipitation days is reported. The average values of LCL and Q_{ten} were calculated for daytime hours only. Values of thermodynamic variables are from the sounding launched at 8 am local time. All other values were calculated from hourly averages.

Parameter	Dry Season		Cumulus Days		Daytime Precipitation Days		t
Number of Days	244		138		64		
	Mean	Std	Mean	Std	Mean	Std	

Surface air Temperature (°C)	27.51	3.89	28.09	4.22	26.82	3.48	11.02
Surface Mixing Ratio (g kg ⁻¹)	19.29	1.55	19.04	1.63	19.51	1.44	10.13
Lifting Condensation Level (m)	819.11	491.20	1010.4	493.65	608.61	373.08	21.22
SHF (W m ⁻²)	23.93	51.69	28.44	58.96	18.47	41.44	6.77
CAPE (J kg ⁻¹)	806.11	565.68	824.93	624.53	846.52	512.38	0.25
CIN (J kg ⁻¹)	-88.82	53.18	-98.94	58.09	-76.40	37.09	3.32
IWV (cm)	4.71	0.67	4.41	0.61	4.98	0.53	6.77
WV _{0-1km} (cm)	1.47	0.12	1.43	0.13	1.50	0.09	4.43
WV _{1-3km} (cm)	2.01	0.27	1.90	0.28	2.09	0.18	5.79
WV _{3-6km} (cm)	1.08	0.42	0.90	0.38	1.20	0.32	5.83
$Q_{ten} _{600-700hPa}$ (g kg ⁻¹ day ⁻¹)	-0.27	22.59	-2.60	20.14	2.48	25.48	4.65
CTP (J kg ⁻¹)	427.87	59.15	435.92	58.47	432.37	48.51	0.45
HI_{low} (°C)	9.91	4.75	11.57	4.99	8.64	3.35	4.91
$\frac{\partial q}{\partial z} _{1-3km}$ (g kg ⁻¹ km ⁻¹)	-3.58	1.05	-3.86	1.12	-3.21	0.84	4.58
$\frac{\partial \theta}{\partial z} _{1-3km}$	4.01	0.74	4.00	0.76	3.91	0.75	0.79

(K km ⁻¹)							
B_{inv}	0.47	0.13	0.43	0.12	0.50	0.11	4.08

Table 2: Rain recurrence time for all rain events (t_r) during 2014 and 2015 dry season. The percent of t_r values greater than 1, 5, and 10 day are reported in parenthesis.

Parameter	All	2014	2015
Number of Rainy days	106	54	52
Number of Rain Events	158	86	72
Mean t_r (day)	1.42	1.36	1.48
Events with $t_r > 1$	63 (40.65)	32 (37.64%)	31 (44.28%)
Events with $t_r > 5$	7 (4.52%)	4 (4.70%)	3 (4.28%)

Events with $t_r > 10$	2 (1.29%)	1 (1.17%)	1 (1.42%)
Mean June t_r (day)	1.04	0.99	1.10
Mean July t_r (day)	1.01	1.09	0.93
Mean August t_r (day)	2.50	2.25	2.79
Mean September t_r (day)	2.03	1.73	2.82
Events in June	51	26	25
Events in July	54	26	28
Events in August	26	14	12
Events in September	22	16	6
Accum. in June (mm)	77.79	42.08	35.71
Accum. in July (mm)	59.89	30.58	29.31
Accum. in August (mm)	36.53	22.89	13.64
Accum. in September (mm)	27.02	21.79	5.23

Table 3: Rain recurrence time for daytime precipitation events (t_{dr}) during 2014 and 2015 dry season. The percent of t_{dr} values greater than 1, 5, and 10 day are reported in parenthesis. The percent contribution of daytime precipitation events to the total precipitation events and to the total accumulated rainfall are also reported in parenthesis.

Parameter	All	2014	2015
Number of daytime rain days	64 (60.37%)	34 (62.96%)	30 (57.69%)
Number of Rain Events	82 (51.89%)	45 (52.32%)	37 (51.38%)
Mean t_{dr} (day)	2.76	2.62	2.95

Events with $t_{dr}>1$	52 (65%)	29 (67.44%)	22 (61.11%)
Events with $t_{dr}>5$	13 (16.25%)	6 (13.95%)	7 (19.44%)
Events with $t_{dr}>10$	4 (5%)	2 (4.56%)	2 (5.56%)
Mean June t_{dr} (day)	2.53	2.69	2.44
Mean July t_{dr} (day)	2.76	2.81	2.69
Mean August t_{dr} (day)	3.44	2.63	4.82
Mean September t_{dr} (day)	2.40	2.39	2.42
Events in June	20 (39.21%)	8 (30.76%)	12 (48%)
Events in July	22 (40.74%)	12 (46.15%)	10 (35.71%)
Events in August	19 (73.07%)	12 (85.71%)	7 (58.33%)
Events in September	18 (81.81%)	12 (75%)	6 (100%)
Accum. in June (mm)	27.57 (35.44%)	15.36 (36.50%)	12.21 (34.19%)
Accum. in July (mm)	17.90 (29.88%)	12.20 (39.89%)	5.70 (19.44%)
Accum. in August (mm)	27.16 (74.34%)	19.57 (85.49%)	7.59 (55.64%)
Accum. in September (mm)	24.36 (90.15%)	19.13 (87.79%)	5.23 (100%)

Figure Captions

Figure 1: Daily accumulated precipitation during the dry season of 2014 (top) and 2015 (bottom). Precipitation was observed on 106 days (43%) from the total 244 days, 54 days in 2014 and 52 days in 2015. The rain events were classified as daytime if it occurred between 9 am and 9 pm, as nighttime if it occurred between 9 pm and 9 am, and mixed if it occurred between the both hours. 64 rain events were identified as daytime, 16 as nighttime, and 26 as mixed.

Figure 2: Mean diurnal cycle of the surface air temperature, mixing ratio, wind speed, wind direction and rain rate during the dry season. The bars denote one standard deviation values.

754 Figure 3: Mean diurnal cycle of ECOR surface sensible heat flux, ECOR surface latent heat flux,
 755 and ECOR surface friction velocity, and SEBS surface latent heat flux during the dry season.

756 Figure 4: Diurnal cycle of upwelling and downwelling shortwave radiation at the surface (top
 757 left), upwelling and downwelling longwave radiation at the surface (top right), liquid water path
 758 (bottom left), and column integrated water vapor (bottom right).

759 Figure 5: Mean profile of potential temperature (θ), water vapor mixing ratio (r), wind speed
 760 (wspd) and wind direction (wdir) as reported by the soundings launched four times a day during
 761 the dry season. The bottom panels show the average profiles of equivalent potential temperature
 762 (θ_e) and saturation equivalent potential temperature (θ_{es}) from the 8 am and 2 pm soundings.

763 Figure 6: Averaged diurnal cycle of hydrometeor fraction as reported by the ARSCL. The black
 764 line shows the mean cloud base height from the ceilometer while the magenta line indicates the
 765 LCL calculated from the surface measurements.

766 Figure 7: Averaged diurnal cycle of horizontal advection of temperature, horizontal advection of
 767 moisture, and large-scale vertical velocity in the lower troposphere during the dry season as
 768 reported by the ECMWF model. The black lines denote the zero value contour in each panel.

769 Figure 8: Diurnal cycle of hydrometeor fraction, first cloud base height (black) and LCL
 770 (magenta) for cumulus days (top) and daytime precipitation days (bottom). The vertical bars in
 771 LCL and cloud base height denote the one standard deviation values.

772 Figure 9: Diurnal cycle of surface temperature, water vapor mixing ratio, wind speed and wind
 773 direction for cumulus (blue) and daytime precipitation (red) days. The vertical bars indicate the
 774 one standard deviation values of the parameters. The gray boxes indicate times when the
 775 difference between the two diurnal cycles were statistically significant.

Figure 10: Diurnal cycle of sensible heat flux, latent heat flux, and friction velocity for cumulus (blue) and daytime precipitation (red) days. The vertical bars indicate the one standard deviation values of the parameters. The gray boxes indicate times when the difference between the two diurnal cycles were statistically significant.

Figure 11: Diurnal cycle of shortwave radiation, longwave radiation, LWP and IWV for cumulus (blue) and daytime precipitation (red) days. The solid and dashed lines in the upper panels indicate downwelling and upwelling radiation respectively. The vertical bars indicate the one standard deviation values of the parameters. The gray boxes indicate times when the difference between the two diurnal cycles were statistically significant.

Figure 12: Averaged profiles of potential temperature (top), mixing ratio (middle), equivalent potential temperature and saturation equivalent potential temperature (bottom) for cumulus and precipitation days from the soundings launched at 0800 LT (left) and 1400 LT (right). The solid lines in the bottom panels denote the equivalent potential temperature and dashed lines denote the saturation equivalent potential temperature.

Figure 13: Average diurnal cycle of the horizontal advection of temperature (left), horizontal advection of moisture (middle), and largescale vertical velocity (right) as reported by the ECMWF model for cumulus (top panels) and daytime precipitation (bottom panels) days. The black lines denote the zero value contour in each panel.

Figure 14: From top to bottom the daily averaged values of IWV, SHF, daytime LCL, daytime moisture tendency between 600-700 hPa (Q_{ten}), and CIN from the 8 am sounding during the 2014 (left) and 2015 (right) dry season. The red squares denote the days with daytime precipitation.

Figure 1: Daily accumulated precipitation during the dry season of 2014 (top) and 2015 (bottom). Precipitation was observed on 106 days (43%) from the total 244 days, 54 days in 2014 and 52 days in 2015. The rain events were classified as daytime if it occurred between 9 am and 9 pm, as nighttime if it occurred between 9 pm and 9 am, and mixed if it occurred in the both hours. 64 rain events were identified as daytime, 16 as nighttime, and 26 as mixed.

Figure 2: Mean diurnal cycle of the surface air temperature, water vapor mixing ratio, wind speed, wind direction, rain rate and pressure during the dry season. The bars denote one standard deviation values.

Figure 3: Mean diurnal cycle of ECOR surface sensible heat flux, ECOR surface latent heat flux, and ECOR surface friction velocity, and SEBS surface latent heat flux during the dry season.

Figure 4: Diurnal cycle of upwelling and downwelling shortwave radiation at the surface (top left), upwelling and downwelling longwave radiation at the surface (top right), liquid water path (bottom left), and column integrated water vapor (bottom right).

843
844
845
846
847
848

849

850 Figure 5: Mean profile of potential temperature (θ), water vapor mixing ratio (r), wind speed
851 ($wspd$) and wind direction ($wdir$) as reported by the soundings launched four times a day during
852 the dry season. The bottom panels show the average profiles of equivalent potential temperature
853 (θ_e) and saturation equivalent potential temperature (θ_{es}) from the 8 am and 2 pm soundings.

Figure 6: Averaged diurnal cycle of hydrometeor fraction as reported by the ARSCL. The black line shows the mean cloud base height from the ceilometer while the magenta line indicates the LCL calculated from the surface measurements.

Figure 7: Averaged diurnal cycle of horizontal advection of temperature, horizontal advection of moisture, and large-scale vertical velocity in the lower troposphere during the dry season as reported by the ECMWF model. The black lines denote the zero value contour in each panel.

Figure 8: Diurnal cycle of hydrometeor fraction, first cloud base height (black) and LCL (magenta) for cumulus days (top) and daytime precipitation days (bottom). The vertical bars in LCL and cloud base height denote the one standard deviation values.

Figure 9: Diurnal cycle of surface temperature, water vapor mixing ratio, wind speed, wind direction and pressure for cumulus (blue) and daytime precipitation (red) days. The vertical bars indicate the one standard deviation values of the parameters. The gray boxes indicate times when the difference between the two diurnal cycles were statistically significant.

Figure 10: Diurnal cycle of sensible heat flux, latent heat flux, and friction velocity for cumulus (blue) and daytime precipitation (red) days. The vertical bars indicate the one standard deviation values of the parameters. The gray boxes indicate times when the difference between the two diurnal cycles were statistically significant.

Figure 11: Diurnal cycle of shortwave radiation, longwave radiation, LWP and IWV for cumulus (blue) and daytime precipitation (red) days. The solid and dashed lines in the upper panels indicate downwelling and upwelling radiation respectively. The vertical bars indicate the one standard deviation values of the parameters. The gray boxes indicate times when the difference between the two diurnal cycles were statistically significant.

Figure 12: Averaged profiles of potential temperature (top), mixing ratio (middle), equivalent potential temperature and saturation equivalent potential temperature (bottom) for cumulus and precipitation days from the soundings launched at 0800 LT (left) and 1400 LT (right). The solid lines in the bottom panels denote the equivalent potential temperature and dashed lines denote the saturation equivalent potential temperature.

931

932

933

934

936

937 Figure 13: Average diurnal cycle of the horizontal advection of temperature (left), horizontal
938 advection of moisture (middle), and largescale vertical velocity (right) as reported by the
939 ECMWF model for cumulus (top panels) and daytime precipitation (bottom panels) days. The
940 black lines denote the zero value contour in each panel.

941

942

944 Figure 14: From top to bottom the daily averaged values of IWV, SHF, daytime LCL, daytime
945 moisture tendency between 600-700 hPa (Q_{ten}), and CIN from the 8 am sounding during the
946 2014 (left) and 2015 (right) dry season. The red squares denote the days with daytime
947 precipitation.

948

949