Two-Day Westward-Propagating Inertio-Gravity Waves during GoAmazon

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ABSTRACT: The dynamical and thermodynamical features of Amazonian 2-day westward-propagating inertia–gravity (WIG) waves are examined. On the basis of a linear regression analysis of satellite brightness temperature and data from the 2014–15 Observations and Modeling of the Green Ocean Amazon (GoAmazon) field campaign, it is shown that Amazonian WIG waves exhibit structure and propagation characteristics consistent with the $n = 1$ WIG waves from shallow water theory. These WIG waves exhibit a pronounced seasonality, with peak activity occurring from March to May and a minimum occurring from June to September. Evidence is shown that mesoscale convective systems over the Amazon are frequently organized in 2-day WIG waves. Results suggest that many of the Amazonian WIG waves come from preexisting 2-day waves over the Atlantic, which slow down when coupled with the deeper, more intense convection over tropical South America. In contrast to WIG waves that occur over the ocean, Amazonian 2-day WIG waves exhibit a pronounced signature in surface temperature, moisture, and heat fluxes.

KEYWORDS: Surface fluxes; Gravity waves; Inertia-gravity waves; Tropical variability

1. Introduction

Quasi-2-day waves are tropical disturbances that involve large-scale westward-propagating “cloud clusters” with periods of ~ 2 days. Their main circulation and convective features could be explained by a convectively coupled $n = 1$ westward inertia–gravity (WIG) wave (Matsuno 1966; Takayabu 1994; Kiladis et al. 2009). Observations over the open ocean during the Tropical Ocean and Global Atmosphere Coupled Ocean–Atmosphere Response Experiment (TOGA COARE) in 1992/93 and during the Dynamics of the MJO (DYNAMO) field campaign in 2011/12 revealed the salient features of the quasi-2-day wave (Takayabu 1994; Takayabu et al. 1996; Chen et al. 1996; Chen and Houze 1997; Haertel and Johnson 1998; Haertel and Kiladis 2004; Wheeler et al. 2000; Kiladis et al. 2009; Yu et al. 2018). These studies documented these disturbances as having a phase speed ranging from 20 to 30 m s^{−1} and a zonal wavelength of about 2000–4000 km. These waves are often embedded within the larger-scale convective envelope of the Madden–Julian oscillation (MJO), leading to complex interactions that result in the seemingly chaotic organization of the convection over the Indian Ocean (Chen et al. 1996; Johnson and Ciesielski 2013; Zuluaga and Houze 2013). Two-day waves are also considered to be the manifestation of the diurnal cycle of deep convection (Chen and Houze 1997) and inertio-gravity waves (Takayabu 1994;

Takayabu et al. 1996; Haertel and Kiladis 2004; Kiladis et al. 2009; Yu et al. 2018).

Over the Amazon, a large fraction of precipitation comes from diurnal variability of deep convection, which is coupled to the surface variations in temperature, moisture, and topography, as well as to large-scale tropical disturbances such as the MJO (Mayta et al. 2019) and Kelvin waves (Liebmann et al. 2009; Serra et al. 2020; Mayta et al. 2021). Many previous studies have also documented cloud systems referred to as “Amazon squall lines,” which propagate westward across the Amazon basin from the northern coast of Brazil (Silva Dias and Ferreira 1992; Cohen et al. 1995; Burleyson et al. 2016).

A recent international experiment, the Observations and Modeling of the Green Ocean Amazon (hereafter GoAmazon) (Martin et al. 2016), documented the characteristics of Amazon squall lines (Burleyson et al. 2016; Anselmo et al. 2020). Observations have revealed a pronounced diurnal cycle in these systems (Burleyson et al. 2016), and a strong seasonality, contributing to the frequency of rain events primarily from March to May (Cohen et al. 1995).

Previous studies have hypothesized that many Amazon squall lines organize as WIG waves (Houze 1977; Tulich and Kiladis 2012). However, the characteristics of WIG waves over land are not well understood since previous studies of 2-day WIG waves have focused mainly on the open ocean (Takayabu et al. 1996; Haertel and Kiladis 2004; Yu et al. 2018). The goal of this study is to elucidate the characteristics of these waves over the Amazon. We will assess whether the previously studied Amazon squall lines are organized in 2-day WIG waves, as in the global analysis of Tulich and Kiladis (2012). We will take advantage of high-resolution radiosonde and

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remote sensor retrievals from radar and satellite to examine in detail the dynamical aspects of 2-day convection-wave coupling over the Amazon. Thus, several key research issues relevant to tropical South America, which were not explored in previous studies, are addressed in the present study. The objectives of this study are the following:

- Document the horizontal and vertical structure of 2-day WIG waves over the Amazon, including their origin and propagation characteristics.
- Obtain a comprehensive understanding of how convection evolves in association with 2-day WIG waves.
- Analyze how surface fluxes of heat and moisture are modulated by the passage of WIG waves.
- Compare the structure and evolution of Amazonian WIG waves with the oceanic WIG waves documented during TOGA COARE (Takayabu et al. 1996; Haertel and Johnson 1998; Haertel and Kiladis 2004) and DYNAMO period (Yu et al. 2018).

The paper is organized as follows. Section 2 provides a brief description of the datasets and analysis methods. In section 3, climatological aspects of 2-day waves are discussed. Properties of the convective and atmospheric features of WIG waves are presented in section 4. Section 5 discusses the local dynamics and thermodynamics modulation by 2-day WIG waves. The major findings of the study are summarized in section 6.

2. Data and methods

a. CLAUS T_b and TRMM/GPM datasets

Satellite-observed brightness temperature (T_b) data are used as a proxy for tropical convection in this study. The data are obtained from the Cloud Archive User System (CLAUS) satellite data (Hodges et al. 2000), which has eight-times-daily global fields of T_b from July 1983 to June 2009. The data have been extended through 2017 using the merged IR dataset from NOAA. We also make use of precipitation data from the Tropical Rainfall Measuring Mission (TRMM) 3B42 available 3-hourly on a 0.25° grid from 50°N to 50°S and from 1998 to 2014 (Huffman et al. 2007). We also use the GPM Dual-Frequency Precipitation Radar version 6 data to extend the TRMM data to 2015. A detailed discussion of the TRMM/GPM data is documented in Sakaeda et al. (2020).

b. ERA-I

To assess the dynamical and thermodynamical characteristics associated with WIG waves, 6-hourly atmospheric variables from the European Centre for Medium-Range Weather Forecasts interim reanalysis (ERA-Interim, hereafter ERA-I; Dee et al. 2011) are used. The ERA-I data used have a horizontal resolution of $2.5^\circ \times 2.5^\circ$. We make use of the following ERA-I fields: horizontal winds (u and v), and geopotential height (z) at low-level (850 hPa) and upper-level troposphere (200 hPa). The horizontal winds are also used to calculate velocity potential (χ) as the inverse Laplacian of the divergence field following Mayta et al. (2020).

c. GoAmazon campaign dataset

The GoAmazon experiment was carried out in Manacapuru, to the west of Manaus in the central Amazon, Brazil (hereafter referred to “T3” site), for the 2-yr time period from 1 January 2014 through 31 December 2015. The field campaign involved measures of complex interactions among vegetation, atmospheric chemistry, and aerosol production and their connections to clouds and precipitation (Martin et al. 2016). In this study, we used winds, temperature, and specific humidity from radiosonde data at T3 (3.2°S , 60.6°W). Important supplementary products including apparent heating (Q_1), apparent drying (Q_2), and large-scale vertical motion (ω), were obtained from variational analysis (Tang et al. 2016), which are available to the community at the Atmospheric Radiation Measurement (ARM) program archive (https://iop.archive.arm.gov/arm-iop/0eval-data/xie/scm-forcing/iop_at_mao/GOAMAZON/2014-2015/). Twice-daily observations from the global, land-based radiosonde station network are also obtained from the Integrated Global Radiosonde Archive (IGRA) (Durre et al. 2006), available at www.ncdc.noaa.gov.

The radar precipitation used in this study is derived from the System for the Protection of Amazonia (SIPAM) S-band radar operated approximately 70 km to the east of the GoAmazon T3 site. The Joss–Waldvogel disdrometer data from the CHUVA campaign (Machado et al. 2014) were used to generate a rain rate product from the SIPAM radar reflectivities. The rain rate estimation and calibration from SIPAM are detailed in previous studies (Tang et al. 2016; Schumacher and Funk 2018).

Surface data used in section 5 come from diverse observational research sites. Surface radiative, latent, and sensible heat fluxes come from the broadband radiometer and eddy correlation flux measurement system (Atmospheric Radiation Measurement 2014) at the ARM Mobile Facility site. Due to limited number of observations and the large data-void regions outside the core sounding sites, we also use the constrained variational (VARANAL; Tang et al. 2016). This product includes the ECMWF analysis together with top of the atmosphere and surface observations from GoAmazon to constrain vertically integrated mass, water, and dry static energy to facilitate its use at 3-h intervals.

In addition to these datasets, cloud characteristics are available through the Merged Cloud Mask and Cloud Type data product (Feng and Giangrande 2018). This product combines the calibrated radar wind profiler (RWP) reflectivity data and the W-band Cloud Radar Active Remote Sensing of Cloud product (WCR; Kollias et al. 2005). The RWP and WCR products have 20-s temporal resolution, with the latter retaining its 30-m vertical resolution and the RWP vertically interpolated to that same 30-m grid. The estimated cloud types used to characterize WIG waves cloud evolution comes from this merged product, available every 30 min from 19 February 2014 through 30 November 2015. It includes seven cloud types: shallow, congestus, deep convective clouds, altocumulus, altostratus, cirrostratus, and cirrus (Feng and Giangrande 2018).

d. Filter

To isolate the WIG waves signal, we follow a procedure similar to the one described in Wheeler and Kiladis (1999).

FIG. 1. (a) Variance (K^2) of WIG waves-filtered CLAUS T_b anomaly for the October 1998–May 2015 period (retaining only variability with period 1–3 days and westward-propagating zonal wavenumbers 1–15). Shading interval is $1 K^2$. The location of the T3 ARM site at Manacapuru ($3.2^\circ S$, $60.6^\circ W$) is indicated by a red star. Times series of 3-hourly (b) precipitation and (c) WIG- T_b filtered CLAUS satellite data at T3 ARM site. The criteria for the convectively active events are indicated in red dashed lines in satellite filtered data ($WIG-T_b \leq -1.0\sigma$) and precipitation rate (rain rate $\geq 1.0 \text{ mm h}^{-1}$). The nine selected WIG events are indicated with red dots and labeled with the time in UTC at T3 ARM site. The wave envelope is shown as a dashed light-blue line.

Space–time filtering on the T_b time series is carried out by retaining only variability with a period of 1–3 days and westward-propagating zonal wavenumbers 1–15.

e. Linear regression technique

Many of the results shown in this study are obtained from linear regression of ERA-I and total T_b data against 3-hourly WIG-filtered T_b at the base point ($3.2^\circ S$, $60.6^\circ W$). The base point is located at T3 (ARM site) which also coincides with the region of October to May WIG waves T_b maximum variance (Fig. 1). The linear regression results based on this time series are then scaled to $-15 K$ as in Kiladis et al. (2009). This value represents a typical T_b peak anomaly during convection associated with WIG waves passage around T3 (Fig. 1c). The statistical significance of these results is then assessed based on the two-tailed Student's t test. This method takes into account the correlation coefficients and an effective number of independent sample (degrees of freedom) based on the decorrelation time scale, as in Livezey and Chen (1983). More details can be found in Kiladis and Weickmann (1992) and Straub and Kiladis (2002).

f. Identification of WIG events

WIG events over tropical South America are identified following a similar set of criteria to those proposed by Yu et al. (2018) and Serra et al. (2020). First, we define 3-hourly time

series of WIG filtered T_b at T3 ($3.2^\circ S$, $60.6^\circ W$). Then, during the active phase, WIG T_b must be lower than -1.0 standard deviation ($WIG-T_b \leq -1.0\sigma$) for active WIG phases, remaining at this amplitude or larger for at least five consecutive time steps. Analogously, WIG- T_b anomalies must be 1.0 standard deviation for suppressed phases ($WIG-T_b \geq +1.0\sigma$); 1.0σ corresponds to ± 5.15 WIG T_b filtered anomalies. In addition, during the active WIG T_b event rain rates must be higher than 1.0 mm h^{-1} in radar precipitation at Manaus (Fig. 1b). To reliably assess the threshold choice, different values were considered (0.75σ , 1.5σ). The results are not sensitive to the threshold used to select events.

g. Phase and group velocity calculation

We apply the Radon transform (Radon 1917) to calculate the phase speeds of WIG waves from time–longitude diagrams. This method has been used in many applications, including the analysis of equatorial waves (Yang et al. 2007; Dias and Pauluis 2011; Mayta et al. 2021). A detailed description of this method and application over tropical South America is documented in Mayta et al. (2021). The group velocity is calculated by first applying a Hilbert transform to the filtered data in order to obtain its wave envelope. A Radon transform is then applied to the envelope in order to calculate the group velocity. Further discussion of the calculation of group velocity is provided in the appendix.

3. WIG climatology over the Amazon

Figure 1 shows the climatological variance of WIG-filtered T_b computed for the October–May period. The maximum variance associated with WIG waves is seen over the west and central Pacific and over Africa. A secondary maximum is seen over South America centered near the equator (10°N – 10°S). The maximum extending into the Atlantic Ocean is placed off-equator along the mean position of the ITCZ.

a. Seasonal variability

To further analyze the seasonality of WIG waves over tropical South America, WIG events are identified for the entire record (1984–2015) and compared to 61 events that were identified during the 2014–15 period. The distribution of these events per calendar month is shown in Fig. 2a. It is clear that the maximum WIG activity occurs from October to May, with peaks in November and April also matching with the seasonal precipitation over the region (Tanaka et al. 2014). Peaks during March–to May are also in agreement with the maximum activity of the squall lines along the Amazon (Cohen et al. 1995). A minimum in WIG occurrence is seen from June to September, even during 2014–15 period, consistent with the northward migration of precipitation during these months. Low activity recorded in WIG events during December of the GoAmazon 2014–15 period is associated with the strong El Niño that occurred in 2015, which usually suppress convection over the Amazon. The seasonality of the WIG waves closely resembles those other equatorial waves over the Amazon such as Kelvin waves (Liebmann et al. 2009; Serra et al. 2020; Mayta et al. 2021). Because the maximum WIG activity occurs from October to May, our analysis will be focused on this period henceforth.

b. 2-day wave maximum amplitude at T3

Studies by Tang et al. (2016) and Giangrande et al. (2017) showed that the large-scale circulation (horizontal scale ~ 2000 – 4000 km) and precipitation over the GoAmazon domain have large diurnal variations. Due to high temporal resolution of the T_b and precipitation data, we investigate at what time of day a WIG wave event amplitude tends to maximize at the T3 ARM site. The times in UTC of the maximum T_b amplitude of WIG events for April 2014 are shown in Fig. 1c, while the composite diurnal cycle of T_b and rain rate of all October–May 2014–15 WIG events are shown in Fig. 2b. The diurnal cycle of radar and TRMM/GPM precipitation is also plotted for the 2 years, excluding the period from June to September.

Precipitation at T3 increases its intensity from 1200 UTC (0800 LT) to 1800 UTC (1400 LT). Both radar and satellite data show a similar diurnal variation, although TRMM/GPM underestimates the total precipitation. This variation is consistent with Tanaka et al. (2014), who showed a similar diurnal cycle of precipitation over Manaus using rain gauge data. Another distinct feature of the WIG events at T3 is that their maximum amplitude coincides with the time of strongest diurnal rainfall. The maximum amplitude of the 2-day waves at T3 (bars in Fig. 2b) is also consistent with the evolution of cloud clusters over this point (Anselmo et al. 2020). These results

FIG. 2. (a) Climatological percent of occurrences during each month (bars) of the all 2-day WIG events filtered T_b anomalies larger than 1.0 standard deviations at T3 (3.2°S , 60.6°W). Red line is also showing the number of WIG events identified during each month of the 2014–15 GoAmazon campaign. (b) Diurnal variation of the peaks of the 2-day WIG events at T3 (bars). The red solid and dashed line depicts the diurnal cycle of precipitation at Manaus from radar and TRMM/GPM data, respectively.

suggest that the mesoscale convective system (MCSs) over the Amazon are often associated with WIG waves.

4. Horizontal and vertical structure of WIG waves over tropical South America

a. Horizontal structure

Figure 3 shows the regression horizontal structure $n = 1$ WIG wave for a base point within the GoAmazon study region at 3.2°S , 60.6°W . Convection propagates westward from the northeast coast of Brazil (day -1 ; Fig. 3a), reaching its maximum intensity at T3 on day 0 (Fig. 3b). After passing the base point (Fig. 3c), the convective area spreads and begins to exhibit an arch-shaped structure in the eastern slope of the Andes, as documented in Takayabu (1994). The evolution of the convective systems resembles the squall lines propagating into the Amazon documented in Silva Dias and Ferreira (1992), Cohen et al. (1995), and the inland propagation of the cloud clusters documented recently in Anselmo et al. (2020).

On the other hand, the maximum divergence at 850 hPa is located in the east of the region of maximum convection. Similar low-level features of 2-day waves are documented in Tulich and Kiladis (2012). Convection, as observed in Fig. 3b, promotes strong easterlies resulting in low-level jet occurrences along with the wave passage (Fig. 3a). This mechanism affects the propagation and life cycle of the convective systems over central Amazon. These results also yield some clues about the organization of these convective systems into WIG waves.

At the upper levels (200 hPa), in contrast, the observed structure is dominated by upper-level convergence into the positive T_b anomalies (Fig. 3d), and divergence from the negative

FIG. 3. Maps of anomalous CLAUS brightness temperature T_b (shading), geopotential height (contours), and winds (vectors) based on 2-day WIG-filtered T_b at T3. Composite maps are shown for (top) day -1 , (middle) day 0, and (bottom) day $+1$ and for (left) 850 and (right) 200 hPa. Anomalies are scaled to a -15 K. Shaded T_b (K) are shown by the legend. The geopotential height contour interval is 4 m, with negative contours dashed. Wind vectors are plotted only where either the u or v component is significant at the 95% level or greater. The location of the T3 ARM site (3.2°S , 60.6°W) is indicated by a green star.

T_b anomalies (Fig. 3e). This structure resembles that predicted by shallow-water theory (e.g., Fig. 3a in Kiladis et al. 2009).

b. Origin and propagation of 2-day waves

Previous studies have discussed the WIG phase speed, which exhibits a wide range of values and is generally slower than the values predicted by dry shallow-water theory. Figure 4 shows a time–longitude diagram of circulation and convective fields considering two different base points. The base point for 2-day waves over the Amazon is 3.2°S , 60.6°W and 3.2°N , 30.3°W for the waves over tropical Atlantic.

Figure 4 shows that the upper-level velocity potential (χ_{200}) is almost in phase with the convective field, with its divergence (negative χ_{200} anomalies) being in the region of the intensified convection for both WIG waves in the Amazon (Fig. 4a) and tropical Atlantic (Fig. 4b). This structure has an average wavelength of 36° (~ 4000 km) and moves westward at a phase speed of 27.3 m s^{-1} . A similar phase speed can be inferred from the cloud clusters tracked by Anselmo et al. (2020, see their Fig. 2). Two-day WIG waves over tropical Atlantic, on the other hand, appear less coupled with convection and propagate at a faster phase speed of 30.6 m s^{-1} . This phase speed is slightly different

from those estimated over oceanic regions. During TOGA COARE, for instance, several authors found a phase speed for 2-day waves ranging from 10 to 20 m s^{-1} (Takayabu 1994; Haertel and Johnson 1998).

The group velocity exhibits a wider range of values. The most characteristic values estimated by the Radon transform are 9.4 and 13.5 m s^{-1} (Fig. 4b). A modeling/observational study of the Amazonian squall lines during Global Tropospheric Experiment/Atmospheric Boundary Layer Experiment (GTE/ABLE) 2B also documented different group velocities, which were found to depend on the wave's zonal scale (Silva Dias and Ferreira 1992). A faster group velocity of 13.5 m s^{-1} is in agreement with the observed mean velocity of propagation of the systems in the Amazon region documented in Cohen et al. (1995), and theoretically estimated for days with squall-line formation cases in Silva Dias and Ferreira (1992).

Figure 4b also provides evidence that the Amazonian WIG waves can be triggered by 2-day waves coming from Africa. This interaction is more evident when the analysis is restricted to only include the months of March to May (Fig. A3). During this period, the Atlantic ITCZ usually comes closer to the equator in the western Atlantic. Observational and modeling studies (e.g., Silva Dias et al. 1984; Silva Dias and Ferreira

FIG. 4. Time-longitude diagram of T_b (shading) and 200-hPa velocity potential (χ_{200} , contours) with T_b variations of WIG ($n = 1$) at (a) T3 ARM, and (b) tropical Atlantic base point. Each field is averaged for the latitudes of 5°S – 5°N . T_b is shaded at intervals of 1 K with red (blue) shading representing positive (negative) values. The contour interval is $2 \times 10^5 \text{ m}^2 \text{ s}^{-1}$. Solid (dashed) contours represent positive (negative) anomalies. The solid arrow represents phase speed estimated by the Radon transform considering different domains (gray dashed lines) and a 6-day window. The red dashed lines represent the longitudinal position of the T3 ARM site.

1992; Cohen et al. 1995) also suggested that squall lines need at least necessary dynamical conditions (e.g., strong low-level winds) to start convection first, and then propagate inland. This low-level jet is an important energy transfer mechanism due to the vertical shear of the basic state wind activity. Vertical wind profiles of 2-day WIG waves at Belem (Fig. A3b), closely resemble the formation of squall lines documented during GTE/ABLE 2B (Silva Dias and Ferreira 1992). We hypothesize that these initial “pulses” are triggered by 2-day waves traveling over the Atlantic region.

c. Vertical structure

1) TEMPERATURE, MOISTURE, AND WINDS

Figure 5a shows a time-height lag regression of temperature during the passage of a WIG over T3. As observed in previous studies, the temperature shows maximum amplitude near the surface. A secondary peak is located in the lower free troposphere (~ 500 – 850 hPa), where tropical rainfall is highly influenced by fluctuations in temperature (e.g., Haertel and Kiladis 2004; Ahmed et al. 2020). Finally, a third peak is located near the tropopause. This third peak is weak when compared to WIG occurring in other regions of the world, such as those seen in the western Pacific (see Fig. 10 in Kiladis et al. 2009) or in the Indian Ocean (see Fig. 12 in Yu et al. 2018). The temperature anomalies tilt westward with height within the boundary layer, eastward from 700 up to 250 hPa, and tilt westward again in the upper troposphere. The evolution of temperature over time shows warm anomalies beginning at low levels near lag -24 h, then during the

intensifying stage, with cold surface from 6 h before and 15 h after the deepest convection.

Before the deepest convection (from lag -24 h to lag -3 h), the troposphere is anomalously warm with enhanced moisture predominantly located over the boundary layer (maximum from -18 to -12 h as shown in Fig. 5b). As the convection develops, the positive moisture anomalies deepen and become more elevated. While moisture anomalies in the midtroposphere (~ 500 hPa) become maximum, the lower troposphere cools and dries.

Figures 5c and 5d show a time-height regression of zonal and meridional winds. The largest amplitude in zonal and meridional winds is located around 200 hPa. An eastward tilt in height of zonal wind, from near-surface to 600 hPa is observed, switching to a westward tilt above this level. This vertical tilt structure suggests downward energy propagation from the upper troposphere into the lower free troposphere (Haertel and Kiladis 2004; Kiladis et al. 2009). Stronger zonal and meridional winds near surface are observed during GoAmazon than in the WIG waves observed over the ocean (e.g., Takayabu et al. 1996; Haertel and Kiladis 2004; Kiladis et al. 2009).

2) VERTICAL VELOCITY (ω), APPARENT HEATING (Q_1), AND APPARENT DRYING (Q_2)

The composite divergence and ω is shown in Fig. 6a. Fifteen hours before the minimum in T_b , shallow ascent is observed. The ascent quickly deepens and becomes maximum from -6 to -2 h. On the other hand, strong near-surface convergence

FIG. 5. Time–height sections of regressed (a) temperature, (b) specific humidity, and (c) zonal and (d) meridional winds at the ARM T3 GoAmazon base point (3.2°S , 60.6°W). Anomalies are scaled to a -15-K perturbation in WIG waves filtered T_b at the base point on lag 0. Contour interval is 0.1 K for temperature, 0.1 g kg^{-1} for specific humidity, and 0.5 m s^{-1} for winds. Red (blue) shadings indicate positive (negative) anomalies. (top) The associated T_b and TRMM/GPM anomalies.

lead the maximum upward motion, with convergence peaking at hour -12 to -6 . The maximum ascent is initially found near 600 hPa but quickly expands upward to 300 hPa , where the divergence field changes sign. The strongest divergence is observed just above the 200 hPa level at the time when the highest rain rate occurs (-3 h before; Fig. 7c). Upper-level convergence after the wave, and near-surface divergence made up the lower-tropospheric subsidence (downward motion in Fig. 6a). Dynamically, the tilt observed in the ω vertical structure has an important role in the maintenance and propagation of the 2-day waves because the downdraft is supported by precipitation from the updraft which, in turn, is fed by the surface convergence induced by the downdraft ahead of the wave.

Figures 6b and 6c show the large-scale vertical structure of the apparent heating Q_1 and apparent drying Q_2 . Starting from -12 h , Q_1 vertical structure show positive anomalies aloft, with a maximum around 500 hPa (from -6 to -3 h). At the same time, opposite anomalies are barely observed near-surface. A transition to cooling with a similar structure to the heating is observed by lag $+9\text{ h}$. Some differences in Q_1 features were also observed compared to those documented

during TOGA COARE (Haertel and Kiladis 2004) and DYNAMO (Yu et al. 2018). For instance, no significant tilts in Q_1 are observed, indicating that it is governed mainly by changes in deep convection. The other difference is the nature of the diurnal cycle, in view of the large difference between the Bowen ratio (much smaller over the ocean compared to land). Thus, the surface forcing is remarkable different due to the large sensible heating during the day whereas over the ocean sensible heating is relatively small during whole day. As consequence, it is reasonable to expect significant differences between the Q_1 vertical profile over land and ocean that lead to the energy being projected to smaller equivalent depth over land than in the ocean (Fig. 6c), which slow down the WIG's in the Amazon compared to typical oceanic speeds (DeMaria 1985). The apparent drying Q_2 , also shows a nearly upright structure with the same polarity as Q_1 (Fig. 6c). The Q_2 peak, however, occurs between the $900\text{--}700\text{ hPa}$ level and from -9 to -3 h (Fig. 6c) when strong precipitation occurs at T3 (later in Figs. 7a and 9d). After lag day 0, the anomalies in Q_2 become shallower, perhaps indicating weaker anomalies in convection. The weaker drying (Q_2 positive anomalies) are also consistent with the upward shift of positive omega at lag $+1$ to about $+4\text{ h}$,

FIG. 6. As in Fig. 5, but for (a) divergence (shaded) and vertical velocity (solid and dashed contours), (b) diabatic heating Q_1 , (c) apparent drying Q_2 , and (d) $Q_1 - Q_2$. Contour interval is $5 \times 10^{-2} \text{ K h}^{-1}$ for diabatic heating and apparent drying, $1 \times 10^{-6} \text{ s}^{-1}$ for divergence, and 1 hPa h^{-1} for vertical velocity.

which indicates a shift from deep to stratiform convection and corresponding weaker rainfall.

Following Yanai et al. (1973), we can show that the difference between Q_1 and Q_2 describes the sum of radiative heating and the eddy flux of moist static energy (MSE):

$$Q_1 - Q_2 = Q_R - \frac{\partial \omega' m'}{\partial p}, \quad (1)$$

where $m \equiv s + L_v q$ is MSE, $s = \Phi + c_p T$ is dry static energy (DSE), L_v represents the latent heat of vaporization ($2.5 \times 10^6 \text{ J kg}^{-1}$), q the specific humidity, and c_p the specific heat capacity of dry air ($1004 \text{ J kg}^{-1} \text{ K}^{-1}$). Masunaga and L'Ecuyer (2014) estimated that the MSE eddy flux exceeds the radiative heating in tropical convection with time scales of 1–2 days. Thus, from Q_1 and Q_2 in Eq. (1) could be used as a proxy of eddy flux of MSE, which is commonly used to measure the activity of convection (Yanai et al. 1973). The vertical profile of MSE flux associated with the passage of 2-day WIG waves is shown in Fig. 6d. The MSE flux builds up near surface during the suppressed phase (before -15 h). Then there is an upward shift in the MSE flux during the early and active convective stages (from -12 to $+6 \text{ h}$), with a peak between 450 and 300 hPa. It is important to stand out that the release is in the

mid- to upper troposphere consistent with the tropics deriving energy from the surface latent heat fluxes and releasing that energy aloft.

5. Local dynamic and thermodynamic modulation by 2-day WIG waves

a. Surface precipitation, wind speed, pressure, and surface fluxes

It is known that convectively driven circulations have pronounced effects at the surface. We use surface data from the VARANAL analysis product for GoAmazon to investigate the influence of the 2-day WIG waves in surface processes. Figure 7 depicts the time series evolution of precipitation rate, surface zonal wind, wind speed, and pressure evolution as the wave pass around the ARM T3 station. In addition, surface heat flux components are considered for the analysis: Sensible heat flux (F_{SH} , upward positive), latent heat flux (F_{LH} , upward positive), net longwave radiative flux (F_{LW} , upward positive), net shortwave radiative flux (F_{SW} , downward positive), and net surface heat flux ($F_{net} = F_{SW} + F_{LW} + F_{LH} + F_{SH}$).

Figure 7a shows the evolution of the surface precipitation rate from radar and the TRMM/GPM estimates (3 hourly).

FIG. 7. (a) Composite time series of surface precipitation from radar and TRMM/GPM computed at the T3 ARM site (3.2°S, 60.6°W). The gray solid line represents the evolution of the filtered T_b WIG waves at T3 during GoAmazon. (b) Surface zonal wind (green line), meridional wind (purple), wind speed (blue), and pressure (dark orange). (c) Time series evolution of surface heat flux components observed at T3: sensible heat flux (F_{SH}), latent heat flux (F_{LH}), net longwave radiative flux (F_{LW}), net shortwave radiative flux (F_{SW}), and net surface heat flux (F_{net}). All flux components are shown by the legend. Anomalies are scaled to a -15-K perturbation in WIG filtered T_b at the base point at lag 0.

Precipitation evolution is centered at lag hour 0, when the WIG-related T_b reaches its minimum value at the base point. During the suppressed stage of the wave (~ -24 h prior the wave), both precipitation estimates show their lowest anomalies. At lag -12 h the rainfall anomalies become positive, indicating that the convection is intensifying. This stage was called “initial convective tower stage” in Takayabu et al. (1996). Precipitation rate peaks from -6 to -3 h in radar observations. This result was expected since strong precipitation occurs when deep convective clouds are dominant in the convective events prior to their maturity stage (Houze and Betts 1981). In addition, there is a spatial offset between the radar located in Manaus and the T3 ARM site (Fig. 1 in Tang et al. 2016), giving some differences in the peak of precipitation. Thus, from Fig. 7a, two differences are seen between the two precipitation datasets: (i) rainfall rates from radar peaks earlier than those from TRMM/GPM (3-h lag); (ii) TRMM/GPM underestimates rainfall rates during maximum WIG-related convection. These results also suggest that a significant fraction of precipitation in the Amazon comes from shallow warm clouds as discussed in previous studies (e.g., Liu and Zipser 2005; Calheiros and Machado 2014). Moreover, the

interpolation algorithm in TRMM, which also uses OLR, apparently does not properly identify the warm cloud precipitation (Cifelli et al. 2002; Betts and Silva Dias 2010). A similar phase shift between radar and TRMM precipitation was also observed during DYNAMO (Yu et al. 2018).

In the theoretical shallow water WIG waves, pressure and zonal wind are spatially anticorrelated (Matsuno 1966; Wheeler et al. 2000), which is observed in Fig. 7b, with anomalously high pressure during easterlies. Zonal wind anomalies at the lower troposphere during the passage of the WIG are stronger compared to oceans (e.g., Kiladis et al. 2009). This is likely in response to the strong precipitation signal over the Amazon. Surface wind speed (Fig. 7b), on the other hand, gradually increases from the initial stages (12 h before the wave), and peaks around the T_b minimum. This peak also reveals the presence of a low-level jet observed in Fig. A3a.

Figure 7c shows the 3-hourly surface heat fluxes during 2-day WIG passage around T3. The surface F_{SH} is largely controlled by the surface winds (e.g., Saxen and Rutledge 1998). However, it exhibits a peak prior to the peak in wind speed likely because the surface temperature drops as cloudiness, rainfall, and wind increase (Figs. 5a, 7b,a). The F_{SH} peaks at the time of maximum convection at the base point (around 3 h before the wave), timing with maximum radar precipitation (Fig. 7a).

The F_{LH} (blue line in Fig. 7c) evolves in phase with the wind speed (Fig. 7b) and precipitation. It exhibits a similar peak amplitude as F_{SH} , although it lags it by ~ 7 h. The comparable magnitude between them observed during GoAmazon is, different to those observed over the ocean, where F_{LH} exceeds F_{SH} by nearly an order of magnitude (e.g., Saxen and Rutledge 1998).

When considering the radiative fluxes, it is clear that the F_{SW} is the largest contributor (Fig. 7c). These fluctuations are likely a result of clouds reducing the amount of incoming surface radiation during WIG convection (Figs. 9 and 10). The F_{LW} (dark purple, Fig. 7c) is the second largest term in the net flux budget at T3. Its magnitude, as observed in Fig. 7c, is slightly larger than F_{LH} .

Finally, F_{net} at the peak of convection results in anomalous net heating at the surface (Fig. 7c). As discussed above, the F_{net} is strongly modulated by a large reduction in the F_{SW} (i.e., F_{SW} is controlled by clouds). However, the contribution of surface heat fluxes (F_{SH} , F_{LH}) and F_{LW} , cancels the contribution of F_{SW} . The relationship between F_{net} , precipitation, and circulation has been explored in several papers (e.g., Biasutti et al. 2018; D’Agostino et al. 2020). Serra et al. (2020) also found similar relations at T3 associated with the Kelvin wave passage using ERA-I surface fluxes.

b. GOES-13 IR imagery and vertical profiler

Figure 8 shows GOES-13 IR images and provides an overview of the westward propagation of the cloudiness field associated with the 2-day waves. Figure 9 presents profiler data for the 24-h window, which covers the period of significant rainfall at T3 associated with the 2-day wave passage that is shown in Fig. 8. Profile values of reflectivity, vertical velocity, cloud types, and precipitation rate come from multisensor ground instruments at the T3 site (more details in Giangrande et al. 2017; Feng and Giangrande 2018).

FIG. 8. *GOES-13* IR highlight temperature for 6 Nov 2015 WIG event. Highlighted temperature ($^{\circ}\text{C}$) is shown by the legend. The location of the T3 ARM site is indicated by a cross circle in each panel. The vertical dashed line indicates the longitudinal location of the T3 ARM site.

The westward propagation of large-scale cloudiness for 0900–2100 UTC is observed in Fig. 8. At T3, strong convection becomes a maximum following the diurnal cycle observed in Fig. 2b. It is worth pointing out that days classified as highly active, cloud clusters tracked in Anselmo et al. (2020, see their Fig. 1 and Table 1) also match with the active phase of 2-day WIG waves at T3.

As shown in Fig. 8, convection around T3 builds rapidly on 6 November. The first rainfall signal associated with the WIG passage (Fig. 9d) occurs at 0900 UTC (lag hour -6), and is associated with congestus convection (Fig. 9c). By 1100 UTC

(lag hour -4), convection near T3 intensifies and the first initial tower of convection is observed by the profiler with echo top reflectivity heights reaching 7 km. The most intense rainfall of the 24-h period begins at 1200 UTC, 3 h before the strong WIG signal at T3 based on T_b . This result captures the results of the composite shown in Fig. 7a, where a 3-h lag was observed between WIG T_b and radar precipitation. At the same time, the first deep convective signal is observed in the merged WACR-RWP profiler with echo top heights reaching at least 14 km.

FIG. 9. WIG hourly evolution from the merged WACR-ARSCL-RWP dataset for a 6 Nov 2015 WIG event: (a) merged WACR-RWP reflectivity, (b) vertical velocity, (c) cloud-type classification, and (d) rain rate and WIG-related T_b anomalies.

On the other hand, strong updrafts accompany the onset of deep convection (Fig. 9b). This rapid transition in the radar indicates that buoyant convective updrafts produced these cells, as also observed during TOGA COARE (Houze 1977; Houze and Betts 1981). Immediately after the strong updraft there is a strong convective-scale downdraft. These deep-convective downdrafts (Figs. 9b,c), are observed near the time of maximum precipitation (Fig. 9d). The high-temporal-resolution data also confirm our results observed in previous Fig. 6a, where strong updraft occurs before the minimum in T_b .

Following the passage of the deep convection at T3, after 1600 UTC (lag hour 1), rainfall decreases and is primarily related to stratiform precipitation, as depicted by the reflectivity and also inferred from the satellite images (Fig. 8d). At this time, the convective signal is propagating westward (Figs. 8c,d). At this time, vertical velocity primarily shows weak downdrafts, also indicative of stratiform precipitation. As the suppressed phase of the WIG approaches T3, the cloud population becomes primarily composed of cirrus and altocumulus. In this final stage,

the convective envelope continues to shift westward (Fig. 8d).

c. Composite features of clouds

Figure 10 shows the percentage of occurrence of cloud-type categories with respect to the total number of cloud profiles available at each hour for a 24-h window, as in Serra et al. (2020). The relative hour considered for the analysis represents the WIG waves phase evolution around T3 (for the 61 WIG events recorded during 2014–15 campaign) as in Figs. 9c and 9d.

As observed in Fig. 9c, the occurrence of the different types of clouds are in agreement with Takayabu et al. (1996). The shallow convection stage (before -9 h) is characterized by a preponderance of shallow clouds (about of 25%). A comparable fraction the cloud population is composed of cirrus (gray bars) or by clear sky (white bar). During the onset of deep convection (from -9 to -2 h), shallow clouds have the highest percentage of the cloud population. Altocumulus and/or cirrus are on par with congestus clouds from -9 to -5 h with congestus finally becoming the second most observed cloud type

FIG. 10. Hourly cloud-type evolution associated with WIG waves convection at T3. Cloud-type occurrence relative to all available cloud profiles from the Merged Cloud Mask and Cloud Type P1 product. Lag 0 h indicates the maximum WIG signal at T3. Different cloud types separated into shallow, congestus, deep, altocumulus, altostratus, anvil, and cirrus are shown by colors according to the legend and clear sky is shown as white.

at -4 h. These initial convective towers produce considerable precipitation rates at T3 as observed in Fig. 9d. Altocumulus (yellow bar in Fig. 10) are commonly observed between lag hours -10 to -5 , in agreement with Fig. 9c. The mature stage (from -3 to $+3$ h) is characterized by deep convection and heavy rainfall (Figs. 9d and 10). Deep convection at 0 h (light blue bar in Fig. 10) accounts for more than 50% of the total cloud population at this time. The presence of congestus and shallow convection around this stage could be related to the short duration of convection. Finally, at the decaying stage ($+4$ h onward; Figs. 9c, 10) the cloud population is primarily composed by cirrus, anvil, and shallow convection. The presence of altocumulus also characterizes this stage. Overall, the evolution of the cloud population resembles the life cycles found for other convectively coupled waves (Mapes et al. 2006; Kiladis et al. 2009).

6. Summary and conclusions

Properties of the convective features and the atmospheric structures of 2-day westward inertio-gravity (WIG) waves, have been investigated using satellite-observed brightness temperature (T_b), reanalysis, and data collected during the GoAmazon 2014–15 campaign. Climatological aspects of these disturbances demonstrate that they are more prominent during the transition seasons (austral spring and autumn), with peaks in November and April (Fig. 2a). The peaks from March to May also coincide with the maximum activity of squall lines along the central Amazon (Silva Dias and Ferreira 1992; Cohen et al. 1995). We provide some evidence that many of these squall lines are organized into 2-day WIG waves.

When considering the wet season only (October to May), convection associated with the 2-day WIG waves is in phase with the diurnal cycle of deep convection at T3 (Fig. 2b), in agreement with many previous studies (Burleyson et al. 2016; Tang et al. 2016).

The 2-day window composite (-1 to $+1$ day) of T_b and upper- and lower-level circulation at T3 could be summarized as follows: (i) westward propagation of convection starts over the northeast coast of Brazil (day -1 ; Fig. 3a), (ii) reach their maximum intensity at T3 on day 0 (Fig. 3b), (iii) finally at day $+1$, 2-day WIG convection area spreads to exhibit an

arch-shaped structure in the eastern slope of the Andes, as documented in Takayabu (1994). The evolution of the convective systems resembles the propagation of squall lines into the Amazon documented in Silva Dias and Ferreira (1992) and Cohen et al. (1995). Indeed, from Fig. A3 it is inferred that 2-day waves promote the vertical shear in the low-level jet to support the long-lasting squall lines. In addition, many of the Amazon 2-day waves (Fig. 4a) come from preexisting WIG waves in the Atlantic Ocean (Fig. 4b), and when reaching the Amazonian region, triggers the appropriate basic state to couple with convection and propagate inland. Preexisting waves in the Atlantic Ocean have been attributed for promoting the right basic state and dynamical condition for the development of the Amazonian convection in other equatorial waves (Raupp and Dias 2005).

The WIG vertical structure over central Amazon is documented using high-temporal-resolution sounding data. Thermodynamically (Figs. 5 and 6), we can infer that cloudiness evolves from shallow to deep and then, stratiform as in other large-scale equatorial waves, the MJO, and even in mesoscale convective systems (MCSS) (Lin and Johnson 1996; Kiladis et al. 2005; Straub and Kiladis 2003; Kiladis et al. 2009; Takayabu et al. 1996; Haertel and Kiladis 2004; Zipser et al. 1981; Houze and Betts 1981; Mapes et al. 2006).

Strong modulation of radiative, sensible, and latent heat fluxes is also observed in response to the 2-day wave (Fig. 7c). We found that the surface latent and sensible heat fluxes are comparable in magnitude, in contrast to those observed over the ocean, where the latent heat flux is nearly an order of magnitude larger than the sensible heat flux (Saxen and Rutledge 1998). These results are expected since WIG waves over the equatorial continental regions exhibit marked differences from their oceanic counterparts. One of them is the nature of the diurnal cycle in these regions (Sakaeda et al. 2020). Over land, the Bowen ratio is much larger than over the ocean. Therefore, WIG waves over land exhibit much larger fluctuations in sensible heating when compared to oceanic WIG waves. Finally, the net shortwave radiative flux is the largest contributor to the surface energy fluxes, with anomalous heating (cooling) during the suppressed (active) WIG phase (Fig. 7c). However, the sum surface heat fluxes and the

longwave radiative flux exceed the contribution of shortwave heat fluxes, resulting in positive anomalies in the net surface heat flux.

The cloud population (Figs. 9c and 10) during the passage of WIG waves was found to evolve in four stages: (i) a shallow convection stage characterized by a preponderance of shallow convection and cirrus; (ii) an initial convection stage mainly composed of cumulus congestus, altocumulus, and shallow convection producing considerable precipitation rates at T3; (iii) a mature stage characterized by deep convection and heavy rainfall; (iv) finally, a decay stage where the cloud population is composed by cirrus, anvil, and shallow convection.

The results presented in this work well describe the main characteristics of one of the most important convective-coupled waves over the tropics. Aspects of convection such as vertical velocity, mass flux, and heating rates are crucial for validating cloud-resolving models and convective parameterizations in the era that weather and climate models move toward higher resolutions. However, several questions remain regarding the WIG waves. The processes that lead to the evolution of the cloud population (Figs. 9c and 10) are not well understood. The relative importance of moisture and temperature in the thermodynamics of WIG waves has not been characterized. Future work will seek to answer these outstanding questions by examining the moist static energy and buoyancy budget applied to these waves (Adames et al. 2021).

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APPENDIX

Propagation Characteristics of 2-Day Waves

a. 2-day WIG waves during March to May

Figure A1 shows the time–longitude diagram of T_b and χ_{200} at T3 ARM site base point, but considering the March to May season. During this period, the Atlantic ITCZ reaches its southernmost position, allowing the formation of the organized systems such as squall lines. From Fig. A3, we see that many of the χ_{200} anomalies begin near the prime meridian and quickly propagate westward. As they approach South America they slow down and become coupled to convection. The coupled system then propagates westward, passing through the T3 region as indicated by the red dashed line.

b. 2-day WIG waves sounding profiles

Figure A2 shows the vertical profiles of zonal wind and wind speed associated with the passage of 2-day waves at the T3

FIG. A1. As in Fig. 4, but for March–May season.

ARM site. In addition, we included sounding profiles at Belem base point (1.48°S, 48.8°W) to compare itself with previous findings such as those documented during GTE/ABLE 2B (Silva Dias and Ferreira 1992). The vertical profile of zonal wind shows strong low-level easterlies with maximum at ~ 900 hPa (Fig. A2) at both stations. Low-level easterlies are accompanied by strong wind speed, stronger around T3 ARM site (Fig. A2a), which confirm the presence of the low-level jets. Vertical wind profile at Belem (Fig. A2b), also resembles those structures observed for the squall lines cases propagating inland during GTE/ABLE 2B (Silva Dias and Ferreira 1992).

c. Calculation of the group velocity of 2-day waves

To study the group velocity of 2-day waves, the wave envelopes of the WIG convection are calculated by using the Hilbert transform, as in Chen and Wang (2018). The Hilbert transform of any time series $f(t)$ is given by the following formula:

$$H(t) = -\frac{1}{\pi} \int_{-\infty}^{\infty} \frac{f(t')}{t - t'} dt'. \quad (\text{A1})$$

For a real signal, there is an associated complex signal called the analytical signal, defined by

$$f(t) - iH(t) = A(t) \exp[i\omega(t)]. \quad (\text{A2})$$

While the real part of the analytic signal is the original time series of the wave, the imaginary is a copy of the original time series with each of its Fourier components shifted in phase by 90° . Finally, the amplitude $A(t)$ of this signal contains the

FIG. A2. Vertical sounding profiles of zonal wind and wind speed (m s^{-1}) at (a) T3 ARM site and (b) Belem (1.48°S , 48.8°W). The vertical sounding profile at Belem comes from the IGRA dataset. The shaded region represents maximum wind speeds in the lower troposphere.

information about the envelope. Once the wave envelope is defined (see Fig. 1c), we used its time series to calculate the group velocity using the Radon transform.

Briefly, the Radon transform (RT) of $g(x, y)$ is the integral of g along the line L oriented at angle θ , with a range of angles from 0° to 180° (see Fig. A1 in Mayta et al. 2021). Thus, the RT can be defined as a projection of $g(x, y)$ on L as follows:

$$P(s, \theta) = \int_u g(x, y) du, \quad (\text{A3})$$

where u is the direction orthogonal to L , and s is the coordinate on L . Therefore, for a given θ , the RT is a function of the line coordinate s . Rewriting Eq. (A3) above in terms of coordinates x and y ,

FIG. A3. Distribution of the sum of squares of the Radon transform vs θ for the (a) phase velocity (C) and (b) group velocity (C_g), normalized such that the maximum value equals 1. The red dashed lines represent the θ_{max} of Eq. (A3) in Mayta et al. (2021).

$$p(x', \theta) = \int_{y'} f(x, y) \left| \begin{pmatrix} x = x' \cos \theta - y' \sin \theta \\ x = x' \sin \theta - y' \cos \theta \end{pmatrix} \right| dy'. \quad (\text{A4})$$

Figure A3 shows the normalized sum of squares of the RT as a function of θ . The distributions in Fig A3 show that the phase speed of the 2-day waves is about $C = 30.6 \text{ m s}^{-1}$ over the tropical Atlantic, and about $C = 27.3 \text{ m s}^{-1}$ over tropical South America. However, θ for C_g varies widely, describing a bimodal distribution in group velocity (Fig. A3b). Examination of the two peaks in the distribution in Fig. A3b yields two group velocities of $C_g = 9.4 \text{ m s}^{-1}$ and $C_g = 13.5 \text{ m s}^{-1}$, respectively.

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