
KAZUAKI YASUNAGA
Research Institute for Global Change, Japan Agency for Marine-Earth Science and Technology, Yokosuka, Japan

BRIAN MAPES
Department of Meteorology and Physical Oceanography, Rosenstiel School of Marine and Atmospheric Science, University of Miami, Miami, Florida

(Manuscript received 27 January 2011, in final form 6 August 2011)

ABSTRACT
Precipitation-related differences in different types of convectively coupled equatorial waves are examined here and in a companion paper. Here the authors show spectra and cross-spectra among tropical-belt time sections of satellite-derived surface rain, infrared brightness temperature $T_b$, precipitable water (PW), and Japan Meteorological Agency reanalysis of divergence and PW. Cross-spectra between rain and divergence at 1000- and 200-hPa levels show significant coherence peaks oriented along the dispersion curves of Kelvin, $n = 1$ equatorial Rossby (ER$n_1$), mixed Rossby–gravity (MRG), $n = 0$ eastward inertial gravity (EIG$n_0$), and $n = 1$ and $n = 2$ westward inertial gravity (WIG) waves, as well as the spectral signatures of the Madden–Julian oscillation (MJO) and tropical depression (TD)-type disturbances. Middle-troposphere divergence (indicative of stratiform rain and half-depth convection involvement in the coupling) is coherent with rain for the higher-frequency and more divergent wave types (Kelvin, EIG$n_0$, WIG) but shows little coherence with rain for more rotational disturbance types (ER$n_1$, MRG, TD). These two broad families also exhibit different rain–PW phase lags, a result supportive of the notion that stratiform rain (which occurs in dry conditions after peak PW and rain) is more involved in the more divergent wave types.

1. Introduction

In the tropics, many types of cloudy convection, from isolated to highly organized systems, are observed. These cloud systems are further organized into statistical envelopes that may be advected with prevailing winds or propagate like waves. Comprehensive surveys by Takayabu (1994) and Wheeler and Kiladis (1999) revealed a significant correspondence between frequency–wavenumber spectra of the cloud field and the dispersion relationships of linear equatorial waves in a motionless basic state (Matsumo 1966), indicating that convectively coupled equatorial waves (CCEWs) play significant roles in modulating tropical convection. A recent review of CCEW science can be found in Kiladis et al. (2009).

One-half of CCEW coupling—how a heat source moving like a wave of convection drives large-scale flows—is fairly straightforward and well understood. The other half—how convection is modulated by larger-scale disturbances, via its environmental conditions—is much subtler. The original motivation for this study was to seek clues about the coupling mechanisms between convection and large-scale waves by seeking differences among different types of waves. This paper describes an approach using multivariate cross-spectral analysis, while a companion paper (Yasunaga and Mapes 2012, hereafter Part II) contains results from composites of spaceborne radar data based on wavenumber–frequency filtered gridded rainfall. A broad consistency of results between the two approaches gave rise to the interpretations offered below.

In the time domain, cloud morphology is broadly self-similar across many kinds of CCEWs and tropical variability (Kiladis et al. 2009; Mapes et al. 2006); there is systematic statistical evolution from shallower to deeper...
convection before peak rainfall, merging into increased prevalence of anvils and stratiform precipitation after peak rainfall. This succession process resembles the time evolution of a single cumulonimbus cloud but is played out statistically in larger-scale envelopes from mesoscale convective systems to CCEWs and the Madden–Julian oscillation (MJO; Madden and Julian 1971, 1972). To draw more discriminating conclusions about mechanisms, it will be necessary to look beyond broad similarities and seek small but robust differences among different wave types and scales.

One wave-type dependence seen in time-domain studies is that lower-frequency variations exhibit relatively larger specific humidity perturbations, relative to the magnitude of temperature $T$ perturbations (e.g., Mapes et al. 2006). With space–time filtering, more refined differences can be observed. For example, Kiladis et al. (2009) present vapor evolutions associated with a $-20$-K perturbation of brightness temperature $T_b$ for each CCEW. For $n = 1$ equatorial Rossby (ERn1) wave, the tilted (in time) structure evident in other waves is obscure. The amplitude of $q$ variations in the middle troposphere is much larger for wave types with more horizontal rotational flow [ERn1 and mixed Rossby–gravity (MRG) waves] than for more purely divergent waves [Kelvin, $n = 1$ westward inertial gravity (WIG), and $n = 0$ eastward inertial gravity (EIGn0) waves]. These results may indicate that coupling mechanisms vary by wave type. Here we attempt to characterize the set of CCEWs using cross-spectra of more variables, in order to raise hypotheses about wave type–dependent coupling mechanisms.

One leading class of theories about convectively coupled waves emphasizes second baroclinic mode dynamics, driven by a vertical dipole component of heating that involves stratiform processes (Houze 1997) and cumulus congestus clouds. Gravest-mode dynamics, driven by deep convective monopole heating, have larger energy but second-mode processes have greater amplitude at low levels, where convection is especially sensitive. There are two schools of thought about why a vertical dipole of heating is essential for the coupling. Early wave–conditional instability of the second kind (CISK) theories (Hayashi 1970; Lindzen 1974) invoked the fact that moisture (and hence moisture convergence) is concentrated in low levels, making the second mode more influential (e.g., Cho and Pendlebury 1997; Matthews and Lander 1999). Later “stratiform instability” ideas (Mapes 2000) reinterpret the key importance of low levels in terms of cloud buoyancy and convective inhibition (CIN). Hybrid control ideas (e.g., “moisture-stratiform instability”) (Khoury 2006; Kuang 2008) are more general and versatile and indeed may imply some frequency dependency as Kuang (2008) demonstrated. Whatever the exact coupling mechanism(s), second baroclinic mode motions [as seen by examining divergence (DIV) in the middle troposphere, not just near surface and tropopause] are thought to be key for understanding wave coupling and so will be examined here, along with humidity data [mainly column integrated vapor or precipitable water (PW)].

There are at least two possible approaches to describe CCEW structures. One involves composite or regression of target fields around a space–time filtered base time series [e.g., $T_b$, outgoing longwave radiation (OLR), rainfall] representing a particular mode (e.g., Haertel and Kiladis 2004; Kiladis et al. 2005, 2006; Wheeler et al. 2000). Our companion paper will take this approach, using nongridged data [Tropical Rainfall Measuring Mission (TRMM) radar-observed precipitation features]. However, since filtering hinges on spectral analysis, we begin here with space–time spectral analysis, whose results fall in a convenient diagrammatic space showing all wave types. Coherence and phase information within that space give us clues, albeit rather abstract ones, about the coupling mechanism across the full range of various CCEWs. Going beyond one-variable power spectra (e.g., Fig. 1), Hayashi (1974) first applied a time–space cross-spectrum analysis to tropical disturbances. Hendon and Wheeler (2008) further demonstrated the utility of examining cross-spectrum coherence between a convection indicator variable (such as OLR or rain) with dynamical fields: in this case there is no need to estimate and remove a background red noise null hypothesis unlike for one-variable power spectra.

A vertical dipole of heating in the troposphere induces the dynamic response of convergence or divergence in the midlevel. Top-heavy or bottom-heavy heating profiles, associated with stratiform precipitation or cumulus congestus–type cloud systems, contain such a dipole component. If dipole heating is crucial for the coupling of a certain type of CCEWs, perhaps via a stratiform instability mechanism, midlevel divergence should have high coherence with rainfall.

Another characteristic of stratiform precipitation (top-heavy heating) is that it occurs after the more intense convective rainfall in mesoscale convective systems (MCSs) and induces a thermally indirect (warm yet descending) mesoscale downdraft (e.g., Houze 2004). Such a downdraft makes the lower troposphere relatively dryer. This suggests that another signature of stratiform rainfall processes may be a phase offset between rainfall and column water vapor (PW), with precipitation shifted into a lagging phase relative to PW. On the other hand, such a phase offset cannot be expected in “scattered convection,” which is characterized by dominant convective rainfall (e.g., Rickenbach and Rutledge 1998).
FIG. 1. Signal strength of (left) symmetric and (right) antisymmetric components of (top) $T_b$, (middle) rain, and (bottom) PW over the latitudes of 15°S–15°N. Shading interval is 0.1 with first level at 0.2, which is significant at the 99% level with 214 DOF. Positive and negative wavenumbers correspond to the eastward and westward propagation, respectively. Dispersion curves are also plotted for Kelvin ($n = 1$ equatorial Rossby (ERn1); $n = 1$ and $n = 2$ westward inertia–gravity (WIGn1 and WIGn2); $n = 0$, $n = 1$, and $n = 2$ eastward inertia–gravity (EI Gn0, EI Gn1, and EI Gn2); and mixed Rossby–gravity (MRG) waves with equivalent depths of 8, 12, 25, 50, and 90 m, respectively. Three dashed lines for the symmetric components indicate constant phase speeds of 7.0, 9.0 and 11.0 m s$^{-1}$. 
For this reason, we will show the phase spectrum between precipitation and water vapor. The remainder of the paper is organized as follows. In section 2, the data used here are briefly described. Power spectra of $T_b$, rain, and PW are also shown in section 2. Cross-spectra (coherence-squared and phase lag information) are shown in section 3. The precipitation characteristics associated with CCEWs are discussed in section 4. The results are summarized and interpretations are offered in section 5.

2. Data and methods

a. Data description

In this section, we will briefly describe the data of precipitation, $T_b$, PW, and divergence used in the present analysis.

1) GSMAP: PRECIPITATION (RAIN)

In the Global Satellite Mapping of Precipitation (GSMaP) project, global precipitation (rain) is estimated by a retrieval algorithm for satellite microwave radiometer measurements. The GSMaP algorithm is based on Aonashi and Liu (2000) and Aonashi et al. (2009). The horizontal resolution is 0.25° in the longitude and latitude directions, covering the whole globe within the tropics about every 12 h. We utilized the GSMaP data spanning the period 1 January 1998–31 December 2006.

2) CLAUS: BRIGHTNESS TEMPERATURE

In the Cloud Archive User Services (CLAUS) project, a long time series of global grid of $T_b$ is produced with 3-h time interval and 1/3° grid spacing. The algorithm for the generation of gridded window $T_b$ values is presented in Hodges et al. (2000). The CLAUS data archive currently spans the period 1 July 1983–30 June 2006.

3) JCDAS: DIV AND PRECIPITABLE WATER

For the divergence and PW, we use reanalysis data offered by the Japan Meteorological Agency (JMA) known as JMA Climate Data Assimilation System (JCDAS) products. JCDAS takes over the same system as the Japanese 25-yr Reanalysis (JRA-25; Onogi et al. 2007), and the data assimilation cycle is extended up to the present. The time resolution of JCDAS is 6 h, and horizontal grid intervals are 1.25° for the divergence and about 1.125° (Gaussian grid to the latitude direction) for the PW. We acquire the data spanning from 1 January 1998 to 31 December 2006 to match the temporal coverage of GSMaP.

4) NVAP: PRECIPITABLE WATER

The National Aeronautics and Space Administration (NASA) Water Vapor Project (NVAP) provides one of the most reliable global PW data sets, and the NVAP Next Generation datasets are the latest phase of the NVAP (hereafter simply called NVAP herein). The NVAP covers the whole globe at 0.5° and twice-daily resolution in years 2000 and 2001. The temporal range is too short to get the significant results in the spectral analyses. On the other hand, the NVAP provides 12-yr PW data (1988–99) at 1°. However, time resolution is low (daily), and the data range does not match that of the GSMaP data. Therefore, we only utilize the NVAP data to check the robustness of the results obtained from the reanalyzed PW data.

b. Calculation procedure

We computed space–time power and cross-power spectra, making use of fast Fourier transform (FFT) algorithms. The calculation procedure is similar to those used by previous studies (e.g., Hendon and Wheeler 2008; Masunaga et al. 2006; Roundy and Frank 2004; Wheeler and Kiladis 1999) and will be briefly described here. First, each variable is partitioned into equatorially symmetric and antisymmetric components; they are defined as a mean from 15°S to 15°N, and $[\Phi(NH) - \Phi(SH)]/2$, respectively. Here, $\Phi(NH)$ and $\Phi(SH)$ denote the quantity in the Northern and Southern Hemispheres, respectively (mean from 15°N/S to the equator). This procedure makes it easy to identify the theoretical equatorial wave modes. Second, the climatological annual cycle, which is defined by the mean and first three harmonics of the annual cycle, is removed to get the anomalies for each component. Third, time series of the anomaly data are divided into 92-day segments that overlap by about 2 months. Then, each segment is detrended and tapered to zero over the first and last 9 days. Finally, the power and cross-spectra are computed for each segment, and they are averaged over all segments.

For the better visualization, power spectra are smoothed by applying the 1–2–1 filter in frequency only, while the 1–2–1 filter to the wavenumber direction is further applied for the cross-spectrum. After the above procedure, the degrees of freedom (DOF) for power and cross-spectra can be calculated as follows:

$$\text{DOF} = 2(\text{amplitude and phase}) \times \text{(years)} \times 365 \times 3(\text{for 1–2–1 filter})/92(\text{days per segment}).$$

The resulting DOF for each product is listed in Table 1. (The DOF for coherence and phase spectra is 3 times as large as that for power spectra.)
c. Power spectra of $T_b$, rain, and PW

Figure 1 shows the symmetric and antisymmetric power spectra of $T_b$ (CLAUS), rain (GSMaP), and PW (JCDAS) in the frequency–wavenumber domain. Since the power spectrum peaks are largely concealed by the red background spectrum, the “signal strength,” which is defined as the strength relative to the background spectrum (Hendon and Wheeler 2008), is used for the display. Although there are several ways to estimate the background spectrum, a red noise process is assumed for each wavenumber in the present study (e.g., Hendon and Wheeler 2008; Masunaga et al. 2006). DOF is estimated to be 214 (the smallest value except NVAP; see Table 1) in the evaluation of the peak significance. In Figs. 1–3, the signal strength greater than 0.2 is shaded, which is significant at the 99% level with 214 DOF.

In the $T_b$ power spectrum (Figs. 1a,b), prominent peaks are apparent along the dispersion curves for theoretical equatorial wave modes: Kelvin, ERn1, MRG, EIGn0, and WIG waves. Moreover, the peaks associated with MJO and tropical depression (TD)-type disturbances (Takayabu and Nitta 1993) are also evident. These features have been repeatedly confirmed since Wheeler and Kiladis (1999).

In the rain power spectrum (Figs. 1c,d), peaks are also concentrated along the dispersion curves, and the patterns resemble that in the $T_b$ power spectrum. However, peaks associated with ERn1 wave and TD are modest. Moreover, the peak shifts to the shallower equivalent-depth part at higher wavenumber and frequencies for Kelvin wave, and the peak separation between Kelvin wave and MJO is more distinct than the $T_b$ power spectrum.

In the PW power spectrum (Figs. 1e,f), the westward-propagating signal is more significant, and the correspondence with the dispersion curves is obscure except for Kelvin and ERn1 waves with smaller wavenumber and longer period. The lack of clear wave signatures in PW was also obtained by Roundy and Frank (2004), making use of the NVAP data. The peaks for the westward wavenumber are oriented along a line corresponding to westward motion at a speed of about 9 m s$^{-1}$ (middle dashed line in Fig. 1e), and the center (around a wavenumber of 10 and a frequency of 0.2) coincides with that of the TD part in the $T_b$ power spectrum. If tropical depressions consist of moisture and vorticity (which can be thought of as short Rossby waves with negligible propagation speeds) being advected by low-level easterly winds, then a part of the TD branch near the dashed lines can be thought of as Doppler-shifted short Rossby waves in the lower troposphere (otherwise known as easterly waves). For the eastward wavenumbers less than 10, two peaks are found: one is positioned along the Kelvin wave dispersion band with equivalent depth $H \sim 50$ m or phase speed $c = (gH)^{1/2} \sim 22$ m s$^{-1}$, and the other extends from the MJO part with a gentle slope (the phase speed corresponds to about 5–9 m s$^{-1}$).

3. Cross-spectral results

a. Coherence of rain and divergence

Figure 2 represents the coherence between rain and divergence at 200-, 500-, and 1000-hPa levels. Coherence squared greater than 0.02 is shaded, which is significant at the 99% level based on the assumed 642 DOF (Table 1).

The coherence spectrum at the 200-hPa level shows clear peaks in almost all modes: MJO, ERn1, Kelvin, MRG, EIGn0, and WIG waves and the TD band. Coherence at the 200-hPa level is slightly weaker than at the 1000-hPa level at high frequencies, and stronger at low frequencies. At the 500-hPa level, however, peaks are seen only for the higher-frequency modes (Kelvin, EIGn0, and WIG waves), with values similar to 200 hPa.

Figures 2 and 3 show that the coherence between rain and divergence has a minimum around the middle troposphere (600–300-hPa level) for ERn1 and MRG waves.
FIG. 2. Space–time coherence-squared spectrum for (left) symmetric and (right) antisymmetric components of rain and divergence at (top) 200-, (middle) 500-, and (bottom) 1000-hPa levels. Shading interval is 0.1 with first contour at 0.02, which is significant at the 99% level with 642 DOF. Dispersion curves are as in Fig. 1.
FIG. 3. As in Fig. 2, but for the coherence-squared spectrum of rain and divergence at (top) 300-, (middle) 400-, and (bottom) 600-hPa levels.
and TDs, while a midlevel maximum is distinct around the 500–400-hPa level for Kelvin, EIGn0, and WIG waves (Figs. 2 and 3). These results are robust to seasonal and geographical subsampling (not shown), lending confidence in their statistical significance. The analysis of midlevel divergence is probably somewhat sensitive to model physics in the data-poor tropics, but Part II also adds independent support based on convective versus stratiform rainfall distinctions from TRMM’s spaceborne radar data.

b. Coherence and phase between rain and PW

Figure 4 shows the coherence of rain and PW. The spectrum shows peaks along the dispersion curves for the Kelvin, EIGn0, and WIG waves as well as for the ERn1 and MRG waves, TD, and MJO, although the peak of PW power is modest for Kelvin waves (especially for the higher-frequency part) and inertia–gravity waves. Figure 5 shows phase lags between PW and rain. Positive and negative values mean that PW variations predate and lag rain variations, respectively. The area with coherence squared larger than 0.02, which is significant at the 99% level, is shaded. Phase spectrum uncertainty is estimated to be about ±2.1° at 99% confidence limit where coherence squared is 0.1 and DOF is 642 (11.5° where coherence squared is 0.02).

Blue colors in Fig. 5 indicate that for the more rotational wave types (ERn1, MRG, TD), PW lags rain. On the other hand, for the more divergent Kelvin, EIGn0, and WIG waves, phase lags vary with the equivalent depth; negative (positive) values appear in the deeper (shallower) equivalent-depth part. Similar features are confirmed when NVAP is used as PW data (Fig. 6), although the pattern is noisy given the small sample number and the significance is too low to evaluate WIG waves. Still, it seems clear enough that the results are not artifacts of the reanalysis data (JCDAS).

c. Implications for possible coupling mechanisms

The significant coherence between rain and midlevel divergence (Fig. 2) suggests that second baroclinic (vertical dipole) heating components likely play an especially important role in the coupling process for higher-frequency or more divergent wave types (Kelvin, EIGn0, and WIG), in contrast with the more rotational ERn1, MRG, and TD disturbances. Still, as Kuang (2008) and Khouider and Majda (2006) point out, moisture preconditioning is crucial for the convection–wave coupling even in pure gravity waves. In fact, high coherence and precedence of PW to rain are found, implying the importance of moisture buildup for the MJO and for the Kelvin, EIGn0, and WIG waves (especially for shallower equivalent-depth parts of these). Moreover, the phase spectrum between lower-level convergence and upper-level divergence from reanalysis data shows the vertically tilted structure in these waves, with low-level convergence leading upper-level divergence (not shown), which is consistent with the results in Kuang (2008). However, opposite phase lags are found in the deeper equivalent-depth components for Kelvin, EIGn0, and WIG waves in Figs. 5 and 6. In the phase spectrum between PW and negative anomaly of $T_b$ (the phase sign is switched), which can roughly be considered as lower- and upper-level water contents respectively, the tilting
structures (or precedence of lower-level moisture) are only seen in the shallower equivalent-depth components (Fig. 7). These results might imply that moisture buildup is less crucial in the coupling for Kelvin, El Niño, and WIG waves, although the significant coherence suggests that moisture likely contributes somewhat to coupling.

In terms of the more rotational waves (ERn1, MRG, and TD), other factors may also play a critical role in the coupling [e.g., vertical shear of the zonal wind, or frictional convergence as in Wang and Xie (1996)]. It is likely that these differences result in the different precipitation characteristics associated with the wave disturbances, since precipitation characteristics are deeply related with heating profile. We will discuss this in the next section.

4. Discussion

a. Precipitation characteristics associated with CCEWs

It is helpful to refer to aspects of mesoscale organization to explain the features in the coherence and phase spectra seen above, foreshadowing some results to be further illustrated in Yasunaga and Mapes (2011). Observational studies reveal that scattered convection has limited stratiform rainfall area (e.g., Rickenbach and Rutledge...
1998), while MCSs are accompanied by extensive stratiform rainfall (e.g., Houze 2004). Convective features have strong near-surface convergence, while stratiform precipitation areas have divergence near the surface and convergence around the melting level (e.g., Mapes and Houze 1995; Mapes and Lin 2005). Therefore, strong midlevel convergence would be absent from scattered convection, in contrast with an MCS. These features are depicted schematically in Fig. 8.

Perhaps a degree of the convective organization can also be linked to phase relationships between PW and rain. As discussed in the introduction, stratiform rainfall follows convective precipitation in MCSs and the induced mesoscale downdraft makes a lower troposphere relatively dryer (i.e., phase shifted into a lagging phase relative to PW). For example, Holloway and Neelin (2010, hereafter HN10) composited PW around rain and found that PW gradually increases before a local high precipitation event in a humid condition, whereas the corresponding decline afterward is steeper. For dryer cases, on the other hand, PW steeply rises before rain and then slowly declines to a higher value than the preconvective state. MCSs occur in a more humid environment, while scattered convection is found in a dryer condition (e.g., Bretherton et al. 2005; Tompkins 2001; Zelinka and Hartmann 2009). Moreover, the mean convective rain rate is much higher in MCSs than in scattered convection (e.g., Nesbitt et al. 2006; Rickenbach and Rutledge 1998). This suggests that the “high precipitation event” in HN10 would reflect the evolution of an MCS accompanied with extensive stratiform precipitation, while their drier case represents the evolution of scattered convection with little amounts of stratiform rainfall. Similar differences in phase relationships between PW and rain are also confirmed in the satellite data–based investigation (e.g., Fig. 8 in Zelinka and Hartmann 2009), and the contrastive roles of convective and stratiform precipitation for PW are discussed in Raymond et al. (2009). The PW asymmetry before and after the precipitation peak corresponds to the lead and lag of PW to rain in the phase spectrum, as schematized in Fig. 8.

The evolution similar to an MCS has been reported for various types of wave-related disturbances (e.g., Lin and Johnson 1996; Straub and Kiladis 2003; Takayabu et al. 1996), and this fact encourages us to apply above mesoscale viewpoint to the results obtained in the spectral analysis, although the scale is different.

From the significant coherence between the precipitation and midlevel divergence, together with PW leading rain, MCSs with extended stratiform precipitation areas are expected in Kelvin and inertia–gravity (EIGn0 and WIG) wave disturbances (especially for the shallower equivalent-depth part). On the other hand, weak coherence of divergence at the midlevel implies that scattered convection with limited area of stratiform precipitation would be most effectively modulated by the ERn1, MRG, and TD disturbances. Based on the preceding paragraphs, the lag of PW to rain also supports this speculation. Recent investigations, making use of radar echoes, pointed out that Kelvin wave disturbances likely contain a larger fraction of stratiform-to-convective area compared to MRG wave disturbances (Holder et al. 2008; Swann et al. 2006). Therefore, our results are partly consistent with previous studies. On the other hand, we cannot rule out other

FIG. 7. As in Fig. 5, but for the phase spectrum of PW and negative anomaly of $T_b$ ($-T_b$). Positive and negative values (shaded with warm and cool colors) indicate that the negative anomaly of $T_b$ predates and lags PW, respectively.
possible interpretations of the coherence and phase spectra. Especially for the phase spectrum, one might consider that the analogy with mesoscale viewpoint is too forced, since a given disturbance has its signal spread throughout a given spectral space. Part II further explores these issues and demonstrates that the phase spectrum reflects the real phase relationships between PW and rain.

From the lead of PW to rain, an MCS (scattered convection) might prevail in slower (faster) parts with shallower (deeper) equivalent-depth components of Kelvin, EIGn0, and WIG waves, although the coherence between rain and divergence has a peak in the midlevel. Moreover, phase spectrum between PW and negative anomaly of $T_b$ indicates that tilting structures like MCSs are only found in the shallower equivalent-depth components of Kelvin, EIGn0, and WIG waves (Fig. 7). The heating associated with deep convection has deeper (monopole) vertical profiles than stratiform (dipole) heating, so the gravity wave phase speed corresponding to convective heating is faster than that to the stratiform heating (e.g., Fulton and Schubert 1985; Mapes 1998; Mapes and Houze 1995). Some schools of thought attribute the vertical mode of the heating profile to the apparent equivalent depth linked with the convective disturbances in the tropics (e.g., Mapes 2000). Our speculation on an MCS is consistent with this hypothesis.

b. Rainfall in reanalysis data

Cumulus parameterization has been recognized as one of the most uncertain parts in a general circulation model. Lin et al. (2006) evaluates the fidelity of CCEWs in climate models, focusing on the peaks of precipitation, and points out that CCEWs are inconsistently simulated by global models with parameterized convection. Therefore, it would be interesting to examine the relationships between observed and analyzed rainfall in the spectral space.

Figure 9 shows phase spectra of observed and reanalyzed rain. Positive and negative values mean that reanalysis rain predates and lags observed rain, respectively. It is clear that phase of the reanalysis rainfall has a systematic bias; reanalysis rainfall leads the observed precipitation for the MJO and Kelvin, EIGn0, and WIG waves, while model lags observation for ERn1 and MRG waves and TDs. Moreover, the lead of the reanalysis rainfall is more prominent in the shallower equivalent-depth parts of Kelvin, EIGn0, and WIG waves. From the discussion in previous section, MCSs with extensive stratiform precipitation can be expected in more divergent waves (especially those with the shallower equivalent depth). Therefore, more appropriate representation of stratiform precipitation might be crucial to reduce the early-rain bias in the reanalysis.

5. Summary and conclusions

In the present study, we have explored the utility of the space–time cross-power spectrum to characterize the convectively coupled equatorial waves (CCEWs). In the brightness temperature $T_b$ power spectrum, significant peaks are positioned along the dispersion curves for Kelvin wave, $n = 1$ equatorial Rossby (ERn1) wave, $n = 1$ and $n = 2$ westward inertial gravity (WIG) waves, $n = 0$ eastward inertial gravity (EIGn0) wave, and mixed Rossby gravity (MRG) wave, as well as MJO and tropical depression (TD)-type disturbances. Similar patterns are
found in the precipitation (rain) power spectrum, but more significant signals are located for the westward-propagating part in the PW.

The coherence-squared spectra between rain and divergence at 1000- and 200-hPa levels show peaks in all modes along the dispersion curves. However, the peaks of the coherent-squared spectrum at the 500-hPa level are significantly weaker than those at the 200-hPa level for the MJO, ERn1 and MRG waves, and TD, although they are almost comparable in Kelvin, EIGn0, and WIG waves. From the comparison to other levels, the coherent-squared spectrum has a minimum in the middle troposphere for ERn1, MRG, and TD, although a midlevel maximum is found for Kelvin, EIGn0, and WIG waves (around 500–400-hPa level). Therefore, it is unlikely that vertical dipole heating components play a particular role in the convection–wave coupling for more rotational waves (ERn1, MRG, and TD).

The phase spectrum indicates that PW variations predate (lag) rain variations for the MJO and Kelvin, EIGn0, and WIG waves (ERn1, MRG, and TD), and precedence of PW peak is more (less) prominent in the shallower (deeper) equivalent-depth components for Kelvin, EIGn0, and WIG waves.

The 500–400-hPa level is close to the melting level, and strong convergence is expected when stratiform precipitation prevails (Mapes and Houze 1995). Recent observational study indicates that lead of PW to rain can be expected in an MCS accompanied with extensive stratiform precipitation, while lag of PW to rain can be accounted by the scattered convection with limited stratiform area (e.g., HN10). When these mesoscale situations are applied to the current result, it is suggested that 1) stratiform precipitation (and MCS) is especially strongly modulated in Kelvin, EIGn0, and WIG wave disturbances; 2) convective rainfall (and scattered convection) is modulated preferentially in ERn1, MRG, and TD disturbances; and 3) an MCS (scattered convection) would more predominantly be coupled with slower (faster) component [i.e., with a shallower (deeper) equivalent depth] of Kelvin, EIGn0, and WIG waves.

The results obtained in the present investigation are partly consistent with previous works (e.g., Holder et al. 2008; Kiladis et al. 2009; Swann et al. 2006) and with the explanation that wave phase speed corresponding to deep-convection heating is faster than that corresponding to stratiform heating (e.g., Fulton and Schubert 1985; Mapes 1998; Mapes and Houze 1995). Nonetheless, the evidence presented above is indirect and more direct evidence is needed. Moreover, the linearity premised in the present study is not fully guaranteed. The work of Yasunaga and Mapes (2011) reinforces the sense of these findings using partitioned rainfall and cloud system size statistics observed by the TRMM Precipitation Radar (PR). While the case for this overall interpretive story is far from airtight, it at least functions as a self-consistent mnemonic of our results, so its mechanistic underpinnings stand as hypotheses worth testing.

Acknowledgments. The results in the present study were obtained using various data sets (GSMaP, CLAUS, JCDAS, and NVAP). The GSMaP project was promoted for a study “Production of a high-precision, high-resolution global precipitation map using satellite data,” sponsored

FIG. 9. As in Fig. 5, but for the phase spectrum of observed and reanalyzed rain. Positive and negative values (shaded with warm and cool colors) indicate that the reanalysis rain predates and lags observed rain, respectively. Contour interval is 5°.
REFERENCES


