A Spectrum Analysis of Synoptic-Scale Disturbances in the ITCZ

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ABSTRACT

Spectral characteristics of convective signals in synoptic-scale disturbances along the intertropical convergence zone (ITCZ) are examined using a 20-yr daily outgoing longwave radiation dataset. A new analysis method, which combines conventional wavenumber–frequency spectrum analysis and wavelet analysis, is developed to explore the longitudinal, seasonal, and interannual variations in these disturbances within the ITCZ whose seasonal migration varies in different parts of the Tropics. The analysis focuses on three longitudinal sectors where the ITCZ can be clearly identified: the western-central Pacific, the central-eastern Pacific, and the Atlantic–West Africa. The most striking results are the evident zonal variability in the spectral properties of westward-propagating synoptic-scale disturbances. The zonal variability exists not only in their dominant frequencies and zonal wavenumbers, but also in their seasonal and interannual variations. Eastward-propagating synoptic-scale disturbances in the ITCZ, in contrast, exhibit much less zonal variability. The results suggest that dynamical relationships between the ITCZ and its embedded westward-propagating synoptic-scale disturbances, if they exist as predicted by theories and numerical simulations of the ITCZ, are likely to vary in longitude.

1. Introduction

Westward-propagating synoptic-scale (wavenumber $\geq 6$ and frequency $\geq 0.1$ cycles per day) disturbances have long been known to exist in the tradewind easterlies and in the intertropical convergence zone (ITCZ). In the tropical regions of West Africa, the Atlantic Ocean, and the eastern Pacific Ocean, some of these disturbances are referred to as easterly waves or African waves. They are typically associated with low-level disturbances in winds with periods of 3–8 days, westward-propagating speeds of 5–20 m s$^{-1}$ and wavelengths of 2000–9000 km (Reed and Recker 1971; Nitta et al. 1985; Tai and Ogura 1987; Thorncroft and Hoskins 1994). Often, but not always, organized deep convection is associated with these disturbances. These disturbances are generally observed during boreal summer. They appear to exist also in the Caribbean Sea, the western Pacific, Southeast Asia, and the Indian Ocean, with similar structures and propagation properties, but possibly different physical origins, energetics, and development mechanisms because of differences in the large-scale circulation and surface conditions (e.g., Riehl 1954; Reed and Recker 1971; Zangvil 1975; Reed et al. 1977; Thompson et al. 1979; Nitta et al. 1985; Tai and Ogura 1987; Takayabu and Nitta 1993; Thorncroft and Hoskins 1994). Tables 1 and 2 summarize the basic characteristics of tropical westward-propagating synoptic-scale disturbances diagnosed from observations and model analyses.

Under favorable conditions, some of the westward-propagating disturbances may develop into tropical cyclones (Carlson 1969). Previous studies indicated that nearly half of the Atlantic tropical storms develop from African waves (Frank 1970).

Another interesting aspect of westward-propagating synoptic-scale disturbances is their potential contribution to mean precipitation and cloudiness of the ITCZ. In boreal summer, many of these disturbances are observed at the latitudes of the ITCZ (Gruber 1972; Tai and Ogura 1987). On the other hand, it is well known that the time-mean ITCZ as indicated by a long-term-averaged distribution of outgoing longwave radiation (OLR; e.g., Fig. 1) is composed of isolated, randomly scattered convective clouds and organized synoptic-scale cloud systems. The organizing mechanisms for the latter are commonly synoptic-scale disturbances in the circulation. Many of these disturbances propagate westward (Fig. 2). These facts have motivated theories that tend to explain the existence of the ITCZ in terms of the synoptic-scale disturbances (Holton et al. 1971; Chang 1973; Lindzen 1974; Hess et al. 1993).1 According to these theories, the existence of

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1 Other hypotheses have also been proposed to explain the intensity and location of the ITCZ (e.g., Pike 1971; Philander et al. 1996). Reviews of these theories can be found in Waliser and Somerville (1994) and Tomas and Webster (1997).
the ITCZ and its preferable latitudes are to a large extent determined or at least modulated by the existence and preferable latitudes of the synoptic-scale disturbances. In these theories, the exact mechanisms for the synoptic-scale disturbances may vary, but one common essence is their implication that the ITCZ would be much weaker (if it existed at all), were there no synoptic-scale disturbances.

These theories have been proposed and tested analytically and numerically. But they have seldom been investigated observationally. In fact, empirically relating the synoptic-scale disturbances in the ITCZ and the ITCZ itself using observations can be interesting, challenging, and rewarding. It is interesting because intriguing questions can be addressed observationally:

Do the mean intensity and latitudinal position of the ITCZ vary coherently with the strength of the synoptic-scale disturbances in the ITCZ on seasonal and interannual timescales? Can the zonal variation in the ITCZ itself be related to the zonal variation in the synoptic-scale disturbances at the latitudes of the ITCZ?

How would an observed ITCZ appear if contributions to the mean cloudiness or precipitation by synoptic-scale disturbances were removed from the data? How does the contribution to the mean ITCZ from synoptic-scale disturbances compare to other possible mechanisms for the ITCZ?

Addressing these questions is challenging partially because of some technical difficulties. For example, the ITCZ is well defined in some regions but hardly exists in others (Fig. 1). The seasonal variations in its strength and central latitudes also depend on longitude (Fig. 3). Empirical relationships between synoptic-scale disturbances and the mean ITCZ should, therefore, be established for individual regions. This can hardly be done using conventional methods of wavenumber–frequency spectrum analysis (e.g., Zangvil 1975; Zangvil and Yanai 1980) that only provide global (in the zonal direction) information of disturbances under study. Regional spectral properties can be obtained using part of a global dataset at the cost of losing zonally propagating signals. Also, traditional linear analysis

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A well-defined ITCZ here refers to a latitudinally confined rain or cloud band in monthly mean data that is zonally elongated (i.e., parallel to the equator). This subjective definition helps distinguish an ITCZ from other tropical convective zones that are less spatially oriented.

### Table 2. Same as Table 1 but for westward-propagating synoptic-scale disturbances in the Caribbean Sea, Atlantic, and West Africa.

<table>
<thead>
<tr>
<th>Region</th>
<th>Data Source</th>
<th>M</th>
<th>T</th>
<th>C</th>
<th>L</th>
<th>References</th>
</tr>
</thead>
<tbody>
<tr>
<td>Western Atlantic and Caribbean Sea</td>
<td>Sea level pressure, u and v (FGGE)</td>
<td>6–9</td>
<td>3–4</td>
<td>6.5</td>
<td>1500–2000</td>
<td>Riehl (1954)</td>
</tr>
<tr>
<td></td>
<td>v (FGGE)</td>
<td>6–8</td>
<td>5.7</td>
<td></td>
<td>4000</td>
<td>Nitta et al. (1985)</td>
</tr>
<tr>
<td></td>
<td>u, v (radiosonde)</td>
<td>7–10</td>
<td>3–5</td>
<td>12</td>
<td>3100–3800</td>
<td>Burpee (1972)</td>
</tr>
<tr>
<td></td>
<td>v (GATE)</td>
<td>8–9</td>
<td>3.5</td>
<td>8</td>
<td>2500</td>
<td>Reed et al. (1977)</td>
</tr>
<tr>
<td></td>
<td>u, v (GATE)</td>
<td>8–9</td>
<td>3.5</td>
<td>8.4</td>
<td>2500</td>
<td>Abignard and Reed (1980)</td>
</tr>
<tr>
<td></td>
<td>v (FGGE)</td>
<td>6–11</td>
<td>3.5</td>
<td></td>
<td>2000–3000</td>
<td>Nitta et al. (1985)</td>
</tr>
<tr>
<td></td>
<td>u, v (ECMWF) and OLR</td>
<td>6–8</td>
<td>3–5</td>
<td></td>
<td>3000</td>
<td>Lau and Lau (1990)</td>
</tr>
<tr>
<td></td>
<td>v (ECMWF)</td>
<td>7–8</td>
<td>3–6</td>
<td></td>
<td>2000–4000</td>
<td>Druyan et al. (1997)</td>
</tr>
<tr>
<td></td>
<td>u, v (ECMWF and NCEP–NCAR)</td>
<td>6–9</td>
<td>3–5</td>
<td>6–9</td>
<td>6000</td>
<td>Diehiou et al. (1999)</td>
</tr>
<tr>
<td></td>
<td>v (radiosonde)</td>
<td>6–10</td>
<td>2.5–6</td>
<td>7–9</td>
<td>2000–4000</td>
<td>Pytharoulis and Thornicroft (1999)</td>
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<tr>
<td></td>
<td>u, v (ECMWF and GCM)</td>
<td>5–10</td>
<td>3–5</td>
<td></td>
<td>2200–4400</td>
<td>Ceron and Gueremy (1999)</td>
</tr>
</tbody>
</table>
methods cannot be used to directly address any rectifying effect of high-frequency phenomena (synoptic-scale disturbances in the present case) on low-frequency variability and the mean (e.g., the ITCZ). Once these and other difficulties are overcome, the reward would be tremendous. We can determine quantitatively whether or not synoptic-scale disturbances are essential components of the ITCZ, and thereby shed light on those ITCZ theories mentioned above.

This study is the first of a series in which the questions posed above are addressed. In this study, spectral characteristics of synoptic-scale disturbances in the ITCZ, especially their zonal, seasonal, and interannual variability, are examined and qualitatively compared to the mean ITCZ. The quantitative relationship between the mean ITCZ and embedded synoptic-scale disturbances will be further examined in its follow-up studies. In these future studies, precipitation associated with the synoptic-scale disturbances in the ITCZ will be identified and removed from the dataset, and mean ITCZs with and without those synoptic-scale disturbances will be compared.

The most unique aspect of this study is its newly developed analysis method. This method allows zonally continuously varying spectra of eastward- and westward-propagating synoptic-scale disturbances to be identified at central latitudes of the ITCZ that undergoes different seasonal migrations in different zonal sectors instead of fixed latitudes (Fig. 3). Details of this method and data used are described in section 2. Using this method, local as well as global features of the synoptic-scale disturbances, including their zonal, seasonal, and interannual variability, are documented (sections 3, 4, and 5, respectively). The implications of this study are discussed in section 6.

2. Data and methodology

A 20-yr dataset of daily OLR from the National Oceanic and Atmospheric Administration (NOAA) polar-orbiting satellites (Gruber and Krueger 1984) is used. Missing data have been filled by interpolation (Liebmann and Smith 1996). The data are on 2.5° × 2.5° grid and extend from 1 January 1979 to 31 December 1998. OLR data have been used in many studies of convective activity and wave motions in the Tropics (e.g., Gruber 1974; Nitta et al. 1985; Hendon and Liebmann 1991; Wheeler and Kiladis 1999). Using OLR to measure tropical synoptic-scale disturbances deeply relies on the assumptions that OLR is a reasonably good representation of tropical deep convective cloudiness and that the disturbances in the dynamic fields are associated with deep convective activity. Although this may not always be the case, westward-propagating synoptic-scale signals in OLR can often be clearly seen along the latitude of the ITCZ in the Atlantic and eastern Pacific, as shown in Fig. 2. As far as the contribution of wave activities to the mean strength of deep convection in the ITCZ is concerned, the OLR data are particularly useful.

Due to unavailability of other direct observations, only OLR data have been used here. Therefore, related
information about dynamic coupling with the wind field is unknown, neither is the vertical structure of these disturbances inferred. In addition, the coarse latitudinal resolution (2.5°) of OLR cannot describe precisely the seasonal/meridional migration of the ITCZ. All of these hinder us from achieving a complete description of the synoptic-scale disturbances in the ITCZ. The National Centers for Environmental Prediction–National Center for Atmospheric Research (NCEP–NCAR) Reanalysis wind data (Kalnay et al. 1996) may be applied in the future. However, in regions, particularly in the Tropics, where direct observations (such as radiosondes) are rare or absent, the reanalysis data are strongly influenced by the “first guess” field produced by the numerical model. Thus, it is desirable to examine other independent observations, such as OLR, before the reanalysis data are used.

The main analysis tool is a combination of the conventional two-dimensional (2D) spectrum analysis method (in a wavenumber–frequency domain) and the wavelet transform. The conventional 2D wavenumber–frequency spectrum analysis decomposes a field into its eastward- and westward-propagating components (Hayashi 1982). The wavelet transform, on the other hand, is a common and powerful tool for detecting localized spectral signals in time or space (e.g., Torrence and Compo 1998). A combination of the two would be able to isolate local westward- and eastward-propagating components of a field. The resulting 3D (latitude–longitude–time) information can be used to estimate the wavenumbers and frequencies of dominant disturbance signals, their spatial distribution and time evolution. The procedure of combining the two is as follows:

1) For a given field $x(\lambda, t)$, where $\lambda = 0–2\pi$ is longitude and $t = -T/2–T/2$ is time, its wavelet transform in longitude yields the coefficient:

$$\hat{c}(k, \lambda, t) = \int_{-T/2}^{T/2} x(\lambda', t) \frac{1}{\sqrt{a(k)}} \psi^* \left( \frac{\lambda'-\lambda}{a(k)} \right) d\lambda',$$

where $a(k)$ is a function of wavenumber $k$, and $\psi^*$ denotes the complex conjugate of a base function $\psi$. When the Morlet wavelet, a complex wavelet base function commonly used in geosciences (e.g., Torrence and Compo 1998), is applied as in this study, the wavelet transform coefficient $\hat{c}(k, \lambda, t)$ is also complex.

2) A complex Fourier transform in time can then be applied to the wavelet coefficient, leading to the Fourier coefficient:

$$C(k, \lambda, f) = A(k, \lambda, f) + iB(k, \lambda, f) = \frac{1}{T} \int_{-T/2}^{T/2} \hat{c}(k, \lambda, t) e^{-2\pi ift} dt,$$

where $f$ is frequency, whose positive (negative) values indicate a tendency of eastward (westward) propagation. The wavelet power spectrum is $q(k, \lambda, f) = A^2(k, \lambda, f) + B^2(k, \lambda, f)$. Notice the difference between the conventional 2D wavenumber–frequen-
cy power spectrum and this wavelet power spectrum. The latter explicitly contains information regarding the distribution of the power spectrum as a function of longitude.\(^3\)

3) To explore the seasonal variation of synoptic-scale disturbances, the above two steps can be applied to a series of subsets of time series \(x(\lambda, t)\), each subset consisting of \(T = T^*\) days. In this study, \(T^* = 92\) days, centered at the middle of each calendar month. Thus, the wavelet power spectrum, \(q(k, \lambda, f, m)\), is actually for a 92-day running window and is a function of month \(m\).

4) The above three steps can be applied to \(x(\lambda, t)\) at any given latitude, \(\phi\). To effectively track signals in synoptic-scale disturbances along the ITCZ, we choose \(\phi\) as the central latitude of the ITCZ, defined as the latitude of monthly averaged minimum OLR. Because of the regional differences in latitude of the ITCZ (Fig. 3), for a given month (and its 92-day running window), a latitude is chosen for each of the three longitudinal sectors: the western-central Pacific (160°E–160°W), the central-eastern Pacific (160°–90°W), and the Atlantic–West Africa (55°W–0°). These three latitudes may and may not be the same but all they are functions of \(m\). In other words, for a given month \(m\), three wavelet power spectrum calculations were made, all for the 92-day running window centered at that month and for all longitudes, but each at the latitude of the ITCZ in one of the three zonal sectors. This procedure would yield spectra \(q[k, \lambda, f, m, \phi_i(m)]\), where \(i = 1, 2,\) and 3 for the western-central Pacific, the central-eastern Pacific, and the Atlantic–West Africa, respectively. Spectra thus calculated were able to capture the variability of wavenumber–frequency power spectra in time (as functions of month \(m\)) and longitude \(\lambda\) following the seasonal migration of the ITCZ indicated by \(\phi_i(m)\).

5) Finally, to practically display a wavelet power spectrum, averaging is needed for some of the five parameters, \(k, \lambda, f, m, i\). For example, for the central-eastern Pacific, \(i = 2\), averaging \(\overline{P}[k, \lambda, f, m, \phi_2(m)]\) over a certain range of zonal wavenumber \(\Delta k\) and the longitude covering the sector \(\Delta \lambda_3\) (160°–90°W) yields a power spectrum as a function of frequency and month, \(\overline{P}_3(f, m) = P[\Delta k, \Delta \lambda_3, f, m, \phi_3(m)]\).

3. Global spectra

In this section, we examine global spectral features of synoptic-scale disturbances at the general latitudes of the ITCZ. All the wavelet power spectra shown in this section were averaged over the three ITCZ latitudes \((i = 1, 2, 3)\). When such a spectrum is further averaged over all longitudes \((\lambda = 0–2\pi)\) and all the months \((m = \text{February 1979–November 1998})\), it becomes almost the same as the conventional 2D wavenumber–frequency power spectrum (Wheeler and Kiladis 1999). Such a spectrum, \(\overline{P}(k, f)\), shows the global features of disturbances at the general latitudes of the ITCZ.

As shown by previous studies (e.g., Gruber 1974; Wheeler and Kiladis 1999), the broad nature of the spectrum is “red” in both zonal wavenumber and frequency (Fig. 4a).\(^4\) Discerning significant signals from the red background is needed, even though the dominance of westward-propagating power is very apparent. Adapting the method of Wheeler and Kiladis (1999), we estimated significant signals in OLR by defining a background OLR spectrum (Fig. 4b). The ratio of the OLR spectrum in Fig. 4a to the background spectrum is shown in Figs. 4c and 4d. Power peaks are considered statistically significant at the 95\% confidence level if the corresponding ratios are larger than 1.1, based on a conservative estimate of the degrees of freedom of 500.

Zangvil (1975) proposed to multiply a power spectrum by wavenumber and/or frequency as a way to focus on the power of synoptic-scale disturbances that are known to exist. This method, although adapted by others (e.g., May 1999), cannot guarantee the significance of any resulting spectral peak. In our current case, nevertheless, the peaks of spectral power identified by this method agree very well to the spectral significance based on the method of Wheeler and Kiladis (1999; Figs. 4c and 4d), within the wavenumber and frequency domains of interest. Our main purpose here is not to detect the existence of synoptic-scale disturbances, but rather to examine the zonal and temporal variability of their spectral signals within the ITCZ with the knowledge that they do exist. Hence, spectra are plotted using the method of Zangvil (1975) with their significance highlighted using the method of Wheeler and Kiladis (1999).

Our primary interest is in zonally propagating signals. It is necessary to remove signals of standing waves. The standing components are estimated by calculating the cross-spectrum using the formula \((K^2(C_4, S_4) + \frac{1}{2}[P(C_4) - P(S_4)]^2)^{1/2}\), where \(C_4\) and \(S_4\) are the cosine and sine coefficients of the Fourier harmonics of a time series \(x(\lambda, t); P, K\) are the power and cospectrum of \(C_4\) and \(S_4\), respectively (Pratt 1976; Hayashi 1982; May 1999). The standing components are very small and not statistically significant (not shown). Thus, we consider the spectral power peaks in Figs. 4c and 4d primarily composed of propagating signals.

The significant power peaks in Figs. 4c and 4d are generally similar to those shown by Wheeler and Kiladis (1999). Signals of the Kelvin wave and the Madden–

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\(^3\)The physical meaning of local signals for small zonal wavenumber \((k < 3)\)\), however, is vague.

\(^4\)Two spurious eastward-propagating power peaks at wavenumber 14 and frequencies of about 0.1 and 0.24 cycles per day exist in the original spectrum plots, which are due to problems from daily averages of satellite orbit data (Wheeler and Kiladis 1999). Attempts were made to remove these spurious spectral peaks, which are described in the appendix.
Fig. 4. Mean zonal wavenumber–frequency global wavelet spectra of OLR along the center of the ITCZ during Jun–Nov. (a) Base-10 logarithm of the raw power spectrum; (b) base-10 logarithm of the “background” power calculated by smoothing the raw power spectrum in (a) many times with a 1–2–1 filter in both wavenumber and frequency (Wheeler and Kiladis 1999); (c) power spectrum in (a) divided by the background power in (b) (shades; heavy dashed line denotes a value of 1.1 for which the spectral signals are significantly above the background at the 95% level, based on 500 dof), and power spectrum multiplied by wavenumber and frequency (contours) for $f$, 0 (westward); (d) same as in (c) but for $f > 0$ (eastward). The ordinate is frequency (cycles per day).

Julian oscillation (MJO) (Fig. 4d), and the Rossby wave and mixed Rossby–gravity wave seem to exist, even though the spectra are for the latitudes along the ITCZ. Globally, the westward-propagating synoptic-scale signals are primarily concentrated in a range of frequency $f = -0.1$ to $-0.3$ cycles per day and for zonal wavenumbers $k = 6–13$ (Fig. 4c). The significant spectral signals within this specific wavenumber and frequency domain may represent two kind of disturbances, that is, westward-propagating mixed Rossby–gravity waves and “tropical depression–type” disturbances (e.g., Takayabu and Nitta 1993). These two distinct disturbances may evolve from one to the other while propagating westward, or coexist in the same region. Here we will not try to separate them because of a lack of dynamic fields in our observations. We will hereafter focus on spectral characteristics of synoptic-scale disturbances within the frequency and wavenumber ranges that show statistical significance. As will become clear in the next section, within the ITCZ, westward-propagating disturbances appear to be more interesting than their eastward-propagating counterparts.

4. Longitudinal variability

When the wavelet power spectrum $P[k, \lambda, f, m, \phi_i(m)]$ is averaged over the months $m$ and the three ITCZ latitudes $\phi_i$, it becomes a wavenumber–frequency
The longitudinal variability in the spectra of the westward-propagating signals can be clearly seen in Figs. 5a and 5b, where the spectra are further averaged over zonal wavenumbers $\Delta k = 6-15$ (Fig. 5a) and over frequencies $\Delta f = -0.11$ to $-0.33$ cycles per day (Fig. 5b). These wavenumber and frequency domains were chosen according to the significance of the power peaks in the global spectra (Fig. 4). There is a spectral peak in the central-western Pacific (about 120°E–160°W), the peak is centered roughly at $f = -0.24$ cycles per day and $k = 11$, as seen in Fig. 4. Meanwhile, a second peak is found in the Atlantic–Africa sector (50°W–20°E). This peak is centered at a higher frequency, $f = -0.29$ cycles per day, than in the western Pacific but roughly at the same zonal wavenumber. The longitudinal discrepancy in the dominant frequencies of the westward-propagating signals between these two regions, shown here using power spectra in a continuous longitude/frequency–wavenumber domain, confirms what has been suggested by previous studies (e.g., Nitta et al. 1985; Nitta and Takayabu 1985; Lau and Lau 1990).

In the western Pacific, the ITCZ is not as well defined as in the eastern Pacific or the Atlantic, but the mean convection in general appears to be stronger off the equator than at the equator during June–November (Fig. 1a). The longitude of the spectral peak (150°E) indeed coincides with the longitude of a local maximum in mean deep convection just north of the equator. The orientation of the mean off-equatorial convective maximum is west-northwestward instead of parallel to the equator. At these longitudes, synoptic-scale disturbances propagate west-northwestward during summer (e.g., Lau and Lau 1990). It is interesting that farther to the west (80°–120°E) and at the similar latitudes, the mean strength of deep convection is sustained or slightly increased whereas the spectral power of the westward-propagating synoptic-scale disturbances decreases. No sign of any well-defined ITCZ can be detected in that region; a mean maximum of deep convection resides in a large region over the western Indonesian Archipelago. This is a region where deep convection is closely associated with both the summer and winter Asian monsoons.

A similar situation can be found over the Atlantic and Africa. Over the Atlantic and West Africa (40°W–0°), the ITCZ is well defined and the power spectra of westward-propagating disturbances shows a distinct peak. To the east and over the African continent (0°–30°E), while mean convection increases, the ITCZ becomes less identifiable and the signals of westward-propagating disturbances become weaker. It is interesting that the spectral signals of westward-propagating synoptic-scale disturbances appear to be more closely related to the identity of the ITCZ than to the mean strength of deep convection.

There are two regions of relatively weak westward-propagating signals. One is over East Africa and the western Indian Ocean (30°–60°E) and the other is over the Caribbean and Central America (90°–60°W). Mean convection is very weak in the region of the western Indian Ocean but relatively strong in the region of Central America (Fig. 1). Weak mean convection and weak westward-propagating disturbances in the western Indian Ocean can both be attributed to the relatively cold sea surface due to strong oceanic upwelling along the Somalian coast driven by the monsoon flow. Strong mean convection and weak westward-propagating disturbances over Central America suggest that the landmass there has different effects on convection associated with the synoptic-scale westward-propagating disturbances and other types of convection.

There is an obvious discrepancy between the dominant frequencies of the westward-propagating spectral power in the Atlantic and in the Pacific: They are systematically higher in the Atlantic than in the Pacific. As will be shown in the next section, the spectral signals in the Atlantic within the frequency range of $f = -0.3$ to $-0.4$ cycles per day are significant locally, even though they are not globally. The discrepancy in the dominant frequencies cannot be completely explained in terms of the Doppler-shift effect of the mean zonal wind (see section 5a). The implication is that either the characteristics of the synoptic-scale disturbances of the
Atlantic are modified by the topography when they propagate across Central America (Zehrnder and Reeder 1997) or most of them in the eastern Pacific hold different origins and development mechanisms (Tai and Ogura 1987).

The relatively weak spectral signals over the eastern Pacific, where the ITCZ is well defined and a perpetual feature, are unexpected. From the western to eastern Pacific along the ITCZ, signals of westward-propagating disturbances become weaker whereas the convective strength along the ITCZ for (a) $\Delta k = 3-8$ and (b) $\Delta f = -0.11$ to $-0.25$ cycles per day. Unit of the ordinate in (a) is cycles per day.

Fig. 6. Mean spectra of eastward-propagating disturbance signals along the center of the ITCZ for (a) $\Delta k = 3-8$ and (b) $\Delta f = -0.11$ to $-0.25$ cycles per day. Unit of the ordinate in (a) is cycles per day.

5. Regional spectra

Described in this section are more detailed regional spectral characteristics of the westward-propagating synoptic-scale disturbances in three regions along the ITCZ: the western-central Pacific (160°E–160°W), the central-eastern Pacific (160°–90°W) and the Atlantic–West Africa (55°W–0°). These regions are selected because the ITCZ is well defined there.

a. General characteristics

Local properties of the westward-propagating synoptic-scale disturbances along the ITCZ were examined through their 2D wavenumber–frequency spectra for the three longitudinal sectors. In each longitudinal sector, a regional 2D wavelet spectrum $\mathcal{P}(k, f)$, $i = 1, 2, 3$ can be obtained by averaging the wavelet spectrum $\{P[k, \lambda, f, m, \phi_i(m)]\}$ over time $m$ and longitude $\lambda$ at the latitude $\phi_i$ of the central ITCZ there. As in section 3, a background spectrum was estimated using the method of Wheeler and Kiladis (1999) so as to describe the local significance of spectral peaks. The results are shown in Fig. 7.

In the western-central Pacific (Fig. 7a), the most significant spectral power is at $f = -0.2$ to $-0.3$ cycles per day and $k = 6-15$. The same spectral peak has been found in the global spectrum (Fig. 4c). Two other peaks are also found. One is at $f = -0.08$ to $-0.15$ cycles per day and $k = 6-9$, and the other is at $f = -0.36$ to $-0.42$ cycles per day and $k = 16-21$. Notice that the latter peak is missing from the global spectrum (Fig. 4c). The three significant spectral peaks may suggest the existence of different types of waves in the region. The frequency and wavenumber ranges of regional dominant signals identified here (i.e., $f = -0.2$ to $-0.3$ cycles per day, $k = 6-15$) are quite consistent with the

weakest near the eastern coast of Africa. These signals in the relatively low-frequency range could be related to the signatures of the MJO (Fig. 6a), which is usually initiated in the Indian Ocean, becomes mature in the western Pacific, and starts decaying as it propagates over the cold water of the central and eastern Pacific (e.g., Rui and Wang 1990). The MJO is a complicated multiscale phenomenon. What we have seen here are probably part of its components. The signals in the relatively large wavenumber and high-frequency ranges could be associated with convectively coupled Kelvin waves. Interestingly MJO or Kelvin wave signatures notwithstaning, there is no apparent correspondence between the eastward-propagating signals and the ITCZ as seen in Fig. 1. From Fig. 6 one can hardly argue that the eastward-propagating synoptic-scale disturbances contribute directly to the zonal variability of the ITCZ. In the next section, we will mainly focus on westward-propagating disturbances when regional spectra are compared for three longitudinal sectors.
results from other studies (e.g., Dunkerton and Baldwin 1995).

The spectrum in the central-eastern Pacific (Fig. 7b) differs from that in the western Pacific. It is generally much weaker as seen from Fig. 5. The primary peak is at $f = -0.08$ to $-0.28$ cycles per day and $k = 6$ to $14$. The secondary peak seen for the western Pacific at $f = -0.36$ to $-0.42$ cycles per day and $k = 16$ to $21$ is absent.

In the Atlantic-West Africa (Fig. 7c), although the spectral power is weaker than in the western Pacific, its peak is highly significant locally. The primary spectral peak is at higher frequencies ($-0.22$ to $-0.4$ cycles per day) and larger wavenumbers ($10$ to $21$) than in the Pacific, consistent with the previous observations (Tables 1 and 2). In this region, westward-propagating synoptic-scale disturbances are often called African waves or easterly waves, which are very important as they frequently develop into Atlantic tropical cyclones (Frank 1970), and even could move into the eastern Pacific (e.g., Zehnder and Reeder 1997). These disturbances over land and the adjacent ocean in the eastern Atlantic-West Africa, however, hold very similar spectral properties (not shown here), even though their intensity can be different (Carlson 1969; Thorncroft and Hoskins 1994). In the global analysis, the westward-propagating disturbance signals within high-frequency ($|f| > 0.3$ cycles per day) and high-wavenumber domain ($k > 13$) are not statistically significant (Fig. 4c). However, they become very strong and are much above their local red background in the Atlantic-West Africa region (Fig. 7c).

The zonal variability in the spectra is intriguing. It may partially result from the Doppler-shift effect of the mean zonal wind and partially from the dynamics of the disturbances in the different regions. It is not obvious that the Doppler shift of deep convective signals such as OLR should be determined by mean zonal wind at low levels where deep convection is originated or at high levels where the signals are observed. For the sake of demonstration, we consider the mean zonal wind at the midlevel (500 hPa). The difference of the mean zonal wind in the Pacific and the Atlantic is about $3$ m s$^{-1}$ (in a range of $1$ to $4$ m s$^{-1}$, see Fig. 1 of Wheeler et al. 2000). The uncertainty in the observed frequency due to the Doppler-shift effect of the longitudinally varying zonal wind would be roughly $0.098$ day$^{-1}$ for $k = 15$. As shown in Fig. 7, the significant regional signals for westward-propagating synoptic-scale disturbances are
approximately at wavenumbers 6–20 in these three regions, and the frequency difference between the significant signals in the Pacific and Atlantic is about 0.1 day\(^{-1}\). The Doppler-shift effect thus appears to be responsible for the differences in frequency between the two regions. However, the differences in zonal wavenumber for the significant spectral peaks between the two regions, also clearly shown in Fig. 7, cannot be explained by the Doppler-shift effect. Therefore, there must be other reasons for the longitudinal variability in the spectra of the synoptic-scale disturbances.

Previous observations and studies indicate that westward-propagating synoptic-scale disturbances originating from West Africa hold their own distinct characteristics and dynamic mechanism that are quite different from those in the tropical Pacific (Burpee 1972, 1974; Reed et al. 1977; Norquist et al. 1977; Thompson et al. 1979). Different energy sources could partially explain this. Dynamic instability due to the African jet is the main energy source for the development of synoptic-scale disturbances over the West African continent (e.g., Norquist et al. 1977; Thorncroft and Hoskins 1994). Latent heat release associated with cumulus convection (e.g., Reed and Recker 1971) and Rossby wave accumulation (Sobel and Bretherton 1999) are proposed to be two candidates for the development of synoptic-scale disturbances in the western Pacific. The barotropic instability of the ITCZ (Ferreira and Schubert 1997) and the propagating of synoptic-scale waves from the Caribbean Sea (Molinari et al. 1997) have been suggested to explain the synoptic-scale disturbances over the eastern Pacific. In addition, other factors, for example, spatial distribution of sea surface temperature (e.g., Chang and Miller 1977) and a non-Doppler effect of the zonal mean wind (Zhang and Webster 1989) may also affect these synoptic-scale disturbances.

b. Seasonal variations

Seasonal variations in the spectra are examined for each of the longitudinal sectors. Three spectra, \(P_{\lambda f M i}(k, f, M, i)\) for \(i = 1, 2, 3\), are obtained by averaging \(P(k, \lambda, f, m, \phi(m))\) over longitudes in each sector and over the same calendar months through the entire data record. To efficiently display the seasonal variability, they are further averaged over different frequency and wavenumber ranges. These frequency and wavenumber ranges are chosen separately for the three longitudinal sectors according to the significant power peaks estimated in Fig. 7. The results are shown in Fig. 8.

The seasonality of the westward-propagating disturbances is remarkable. These disturbances are active during June–November. Most previous studies on tropical westward-propagating synoptic-scale disturbances focused on part of this period (Tables 1 and 2). The disturbances are almost absent during January–April. In addition to this common seasonal contrast, there exist detailed differences in the seasonal cycle of the disturbances among the three regions. For example, the maximum spectral peaks appear to occur a little later in the western Pacific than in the Atlantic. In the Atlantic, spectral peaks shift gradually from slightly higher to lower frequencies and smaller to larger zonal wavenumbers during the active season of May–October. The implication of this systematic transition in the spectra is unclear. It may reflect a response of the disturbances to their seasonally changing environment, such as SST and mean winds.

The weak or absent signals during January–April are intriguing. Recall that the power spectra were calculated at central latitudes of the ITCZ, which migrates seasonally and is the closest to the equator during January–April (Fig. 3). Distribution of SST, increasing in the western Pacific but decreasing in the eastern Pacific when moving toward the equator, seems not to be a main factor. Barotropic instability and ITCZ breakdown may generate synoptic-scale disturbances (Ferreira and Schubert 1997). The resulting disturbances become weaker when the ITCZ is closer to the equator. But the seasonal displacement of the ITCZ is drastically different in the western Pacific and Atlantic Ocean (Fig. 3), yet the seasonal contrast in the synoptic-scale disturbance embedded in the ITCZ is remarkably similar in the two regions. Whether the seasonal variation in the westward-propagating disturbances is related to the latitudinal position and intensity of the ITCZ or to the seasonal cycle of other environmental factors, or both, is currently under investigation.

c. Interannual variability

The interannual variability of the westward-propagating synoptic-scale disturbances along the ITCZ were investigated by averaging the wavelet power spectrum \(P[k, \lambda, f, M, \phi(m)]\) over longitude in each of the three longitudinal sectors and over normal, warm, and cold years of the ENSO cycle,\(^*\) respectively. The spectra were further summed over a range of zonal wavenumbers and frequencies corresponding to the regional significant signals (Fig. 7).

The most interesting result is that the interannual variability is different in these three longitudinal sectors, especially between the Pacific and Atlantic. Having different influences in different regions, the warm (cold) ENSO events generally correspond to stronger (weaker) signals of westward-propagating disturbances in the Pacific, particularly in the eastern Pacific (Fig. 9), but apparently show opposite effects on the disturbances in

\(^*\) The years 1982, 1986, 1987, 1991, 1993, 1994, and 1997 are considered to be the years of the warm ENSO events (about six continuous warm months during May–Dec existed in these years); 1984, 1988, 1989, 1995, and 1998 the years of the cold ENSO events (about six continuous cold months during May–Dec existed in these years); and the rest the normal years, based on Trenberth (1997) and the Southern Oscillation index (SOI) from NCEP (Fig. 9b).
Fig. 8. Spectra of westward-propagating disturbances along the center of the ITCZ as function of month and wavenumber \([P(\Delta f, k, M)]\), and month and frequency \([P(\Delta k, f, M)]\). Here \(\Delta f = -0.11\) to \(-0.33\) cycles per day in (a) and (c), \(-0.21\) to \(-0.4\) cycles per day in (e); \(\Delta k = 6\) to 15 in (b) and (d), 9 to 18 in (f). (a) and (b) In the western-central Pacific (160°E–160°W); (c) and (d) in the central-eastern Pacific (160°–90°W); (e) and (f) in the Atlantic–West Africa (55°W–0°). Unit of the ordinates in (b), (d), and (f) is cycles per day. Shading indicates that regional signals are significant.

The correlation coefficients between June and November mean spectral power (Fig. 9a) and the SOI (Fig. 9b) are \(-0.47\), \(-0.68\), and 0.33 for the western-central Pacific, eastern Pacific, and Atlantic–West African region, respectively. At the 95% confidence level for 18 degrees of freedom (df), the correlation would be significant if its coefficient is above 0.44. Even though it is unclear whether each year (or season, i.e., June–November mean) can be taken as an independent sample as treated here, the main point is that ENSO has the opposite effects on westward-propagating synoptic-scale disturbances in the Pacific and the Atlantic. Thorncroft and Rowell (1998) have shown a marked interannual variability in African easterly wave activity in a GCM. But there have been few, if any, observational studies relating the westward-propagating synoptic-scale disturbances to the ENSO cycle.

The cold ENSO events obviously suppress the activity of eastward-propagating synoptic-scale disturbances (possibly MJO–Kelvin wave modes) in these three regions. The evident intensification of these disturbances during the years of the warm ENSO events, however, only happened in the Atlantic–West Africa (not shown here).

Figure 9 also indicates a similarity between the interannual variability of westward-propagating synoptic-scale disturbances in the Atlantic–West African region and that of the annual number of Atlantic hurricanes (dash–dot line in Fig. 9a; see Fig. 4 in Landsea et al. 1999). This confirms the previous results that more easterly waves over Africa could be associated with more
synoptic-scale waves in the Atlantic, and then implying more tropical Atlantic storms (Thorncroft and Rowell 1998). Previous studies indicate that the ENSO influence on tropical cyclone activity is quite different in different geographical locations (Gray 1984; Lander 1994; Henderson-Sellers et al. 1998). There are reduced numbers of tropical cyclone genesis west of 160°E, but increased cyclogenesis events east of 160°E during the warm ENSO events. In the Atlantic, tropical cyclone genesis and intensification are discouraged during the warm ENSO events. This is generally consistent with our results about ENSO effects on the westward-propagating synoptic-scale disturbances along the ITCZ in the region.

The results from this study clearly show the strong regional dependence of the impact of ENSO on the tropical disturbances. However, the disturbance signals do not closely follow the signature of ENSO every year. To further explore the mechanisms for the interannual variability of westward-propagating synoptic-scale disturbances along the ITCZ, only one or even two more factors are apparently not enough. The phase of stratospheric quasi-biennial oscillation (QBO) may be an important factor for tropical disturbances (Gray et al. 1992; Dunkerton 1993; Landsea et al. 1999). Atlantic hurricane activity is generally enhanced during the westerly phase of the QBO and suppressed in the easterly phase. For westward-propagating synoptic-scale disturbances in the Atlantic and Pacific, however, this relationship is not very obvious. For example, in 1988, the westerly phase of the stratospheric QBO and the cold ENSO event effectively increased the activity of tropical disturbances in the Atlantic–Africa, and obviously suppressed their spectral power in the Pacific (Fig. 9). In 1994, the warm ENSO event and the easterly QBO decreased the synoptic-scale disturbances in the Atlantic, but not in the Pacific. However, in 1984, the relatively weak cold ENSO event enhanced the appearance of tropical disturbances in the Atlantic–Africa and the Pacific, even though the QBO was in the easterly phase.

The interannual variability of westward-propagating synoptic-scale disturbances and its regional dependence could also be closely related with various dynamic mechanisms for the formation of westward-propagating synoptic-scale disturbances and their embedded large-scale environment, for example, zonal wind field and sea surface temperature. For instance, the African easterly wave activity could be positively correlated to seasonal mean rainfall in the Guinea coastal region and the intensity of the African easterly jet (Thorncroft and Rowell 1998), and western Sahelian monsoon rainfall variability is also suggested to be strongly linked to variability of tropical Atlantic storms (Landsea and Gray 1992). To further address these issues, distribution and variation of other large-scale factors (e.g., SST, surface wind fields, etc.) should also be considered. The relationships between synoptic-scale disturbances and global-scale phenomena will be quantified in the future.

6. Summary and discussion

Spectral characteristics of westward-propagating synoptic-scale disturbances along the ITCZ were explored. A new analysis tool was developed which combines conventional wavenumber–frequency spectrum analysis and wavelet analysis to make possible the calculation of spectra for zonally propagating disturbances along the observed zonally asymmetric ITCZ whose seasonal migration varies in different parts of the Tropics (Fig. 3). The desire of capturing signals of propagating disturbances along the zonally asymmetric ITCZ is motivated by a lack of detailed observational clarification of any possible relationship between the two. Such a relationship has been predicted by theories and numerical simulations of the ITCZ (e.g., Holton et al. 1971; Hess et al. 1993). Daily mean OLR data were used to represent deep convective signals of synoptic-scale disturbances. Regional spectra of westward-propagating disturbances along the ITCZ and their seasonal and interannual variability were focused. The main results are as follows.

- The variability of the spectral characteristics in longitude is striking. Not only are the dominant zonal wavenumbers and frequencies different (Figs. 5 and 8), but so are the seasonal cycles (Fig. 8) and interannual variability (Fig. 9). Based on the conventional method, a global spectra and its seasonal and inter-
annual variability would be dominated by the strongest signals in the western-central Pacific.

- The strength of the spectral signals varies zonally with the identity of the ITCZ instead of total convection. It becomes weaker where the ITCZ is less identifiable even though total convection may be stronger (Figs. 1 and 5). This suggests a possible dynamic relationship between the disturbances and the ITCZ.

- Westward-propagating synoptic-scale disturbances are in general the weakest during January–April and the strongest during July–October (Fig. 8). The most interesting seasonal cycle is found in the Atlantic–West African sector, where there is a tendency that the dominant frequency (and zonal wavenumber) of westward-propagating disturbances varies from the early to late season in the year.

- The zonal dependence of the interannual variability in westward-propagating synoptic-scale disturbances is evident. Warm (cold) ENSO events tend to increase (suppress) the disturbance activities in the Pacific (east of 160°E), but suppress (increase) them in the Atlantic (Fig. 9).

- Eastward-propagating synoptic-scale disturbances along the ITCZ show maximum spectral signals in the western Pacific and a monotonical decay moving eastward (Fig. 6).

These results strongly suggest that the dynamics of the disturbances in different longitudinal sectors might be different. Any relationship between the ITCZ and its embedded disturbances, therefore, must be quantified separately for different parts of the Tropics.

Several aspects of this study should be clarified. First, all (zonal, seasonal, and interannual) variability of the propagating disturbances discussed in this study should actually be the variability of convective signals of those disturbances in the ITCZ. They are by no means the variability of the disturbances in the dynamic field. The spectral characteristics of the dynamic field of the disturbances may not be the same as those of deep convection. Second, because of the missing dynamic field from the present analysis, no attempt was made to categorize the dominant disturbances observed in this study. In the Atlantic–West Africa, these disturbances are likely what are commonly referred to as easterly waves or African waves. In the Pacific, some of them are possibly similar to those in the Atlantic, others might be completely different (such as the equatorial mixed Rossby–gravity waves, e.g., Liebmann and Hendon 1990; Hendon and Liebmann 1991). We leave the task of dynamically categorizing these disturbances to the future when reliable observations of dynamic fields or equivalence are available. Third, the results from this study are rather qualitative. The data record length (20 yr) does not allow a further quantitative description of relationships between the propagating disturbances in the ITCZ and other phenomena, such as ENSO and QBO. Possible relationships between the ITCZ and propagating disturbances wherein for their seasonal and interannual variations have been deliberately excluded. Those relationships need to be quantified in comparison to possible relationships between the ITCZ and zonally propagating disturbances in general whose central latitudes may not be within the ITCZ. Such a quantitative analysis will be reported in the near future.

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Fig. A1. (a) Unsmoothed and (b) smoothed spectra ($f > 0$) along the center of the ITCZ. Unit of the ordinates is cycles per day.
APPENDIX

Removal of Erroneous Spectral Peaks

Because global daily gridded OLR data were assembled from 14 data swaths corresponding to 14 individual satellite orbital paths, two obviously spurious peaks exist in its spectrum (Wheeler and Kiladis 1999). One is at eastward wavenumber 14 and frequency 0.1 cycles per day; another at eastward wavenumber 14 and frequency 0.24 cycles per day (Fig. A1a). Obviously, there is no effect of these two spurious peaks on the significant power spectral peaks in the frequency and wavenumber range of interest (i.e., westward-propagating components). In spite of this, we smoothed the wavelet power spectra between wavenumbers 12–18 and frequencies 0.08–0.13 cycles per day about 60 times to limit their contaminations in the eastward-propagating part of the spectra. A smoothed spectrum is shown in Fig. A1b. The first spurious peak is almost removed.

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