Westward-Propagating Synoptic-Scale Disturbances and the ITCZ

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ABSTRACT

In an attempt to understand the dynamics of the intertropical convergence zone (ITCZ), this study explores the extent to which the ITCZ is causally related to zonally propagating synoptic-scale disturbances. The ITCZ, measured by its mean convection, is represented by mean outgoing longwave radiation (OLR). Synoptic-scale disturbances, measured by their deep convective signals, are represented by the spectral power of the OLR that is significantly above its red-noise background. Time-mean spatial distributions as well as annual and interannual variability of the ITCZ are compared with those of synoptic-scale disturbances, which are dominated by westward-propagating signals. In general, they match each other well in their mean distributions and annual cycles. But, in detail, discrepancies between the two fields exist, some of them substantial. The maximum disturbance activity tends to be located at the polar side of the ITCZ. The seasonal cycles of the two share many similarities, but the variations in the intensity and latitudinal locations of the disturbances are greater than those of the ITCZ. On interannual timescales, their relationship is even more limited. Comparisons are also made between the observations and theories relating the ITCZ and westward-propagating synoptic-scale disturbances. The results suggest that the observed ITCZ does not owe its existence to zonally propagating synoptic-scale disturbances, in the sense that it would still exist in the absence of the disturbances. But the similarities in their means and annual cycles imply that the disturbances alone can result in an ITCZ resembling the observed one in many respects. The observations, on the other hand, are consistent with the theories that view the dynamical instability of the ITCZ as a cause of some westward-propagating synoptic-scale disturbances.

1. Introduction

Explaining the intertropical convergence zone (ITCZ) remains a challenge. Various theories of the ITCZ have been proposed. They can generally be categorized into two groups. In the first group, the intensity and location of the ITCZ are thought to be closely related to the sea surface temperature (SST) (e.g., Pike 1971; Xie and Philander 1994; Philander et al. 1996). The second group of theories focus on atmospheric internal dynamics. Some of them emphasize only zonally symmetric processes (e.g., Charney 1971; Lindzen and Hou 1988; Hack et al. 1989; Waliser and Somerville 1994; Tomas and Webster 1997; Raymond 2000). Others consider zonal asymmetries, such as regional monsoon flow (Mitchell and Wallace 1992) and propagating waves/disturbances (e.g., Lindzen 1974).

One particular theory in the second group relates the preferred latitude and perhaps the strength of the ITCZ closely to atmospheric zonally (mostly westward) propagating synoptic-scale disturbances (Holton et al. 1971; Chang 1973). Holton et al. (1971) proposed that maximum convergence in the boundary layer would occur near the (critical) latitude where the Coriolis frequency equals the Doppler-shifted frequency of atmospheric disturbances. They argued that both the ITCZ and maximum synoptic-scale disturbances should be found near this critical latitude. Satellite observations indeed indicate that westward-propagating synoptic-scale disturbances vary with frequencies of 0.2–0.25 cycles day\(^{-1}\) are often seen along the latitudinal belt of the ITCZ (Chang 1970). The predicted ITCZ is about 6°–7° away from the equator, roughly consistent with observations (Waliser and Gautier 1993). This theory was furthered by Chang (1973). It was found that this mechanism was indeed responsible for the existence of the ITCZ in a numerical model, although the ITCZ developed 1°–3° north of the critical latitude. In a GCM study, Hess et al. (1993) examined the role of tropical wave transients in the formation of the ITCZ. They found that zonally propagating equatorial waves and the location of the ITCZ are mutually dependent.

According to the theories of Holton et al. (1971) and Chang (1973), instability in the atmospheric internal dynamics determines the latitude of synoptic-scale disturbances, and a long-term mean of cloudiness or precipitation associated with the disturbances constitutes a narrow, zonally oriented zone of concentrated cloudiness or precipitation north of the equator, which is commonly referred to as the ITCZ. Implicit in the theory...
are influences on the ITCZ from other factors, such as the distributions of SST and topography, because of which the ITCZ is equatorially asymmetric most of the time.

The theories of Holton et al. (1971) and Chang (1973) appear to be very attractive based on the impetus of satellite images that show clear westward-propagating signals along the general latitudes of the ITCZ and the dominance of synoptic-scale signals in the subseasonal variance of ITCZ cloudiness (Chang 1970; Salby et al. 1991). But their theories and the general ideas that the ITCZ can be causally related to synoptic-scale disturbances have yet to be scrutinized quantitatively against observations. Zonal, annual, and interannual variations in both intensity and latitudinal location of the two phenomena can be compared using observations as suggested by Holton et al. (1971). They should follow each other if the theories are correct. The purpose of this study is to explore the coherence between the two in observations, seek evidence validating the theories, and explore the degree to which the ITCZ is causally related to synoptic-scale disturbances.

One major technical difficulty in the task assigned to this study is to identify significant signals of propagating synoptic-scale disturbances that might depend on zonal locations. The conventional wavenumber-frequency spectrum analysis method decomposes a given field into its eastward- and westward-propagating components (Hayashi 1982), but without any local information in longitude. Both the mean ITCZ and its embedded synoptic-scale disturbances, however, vary substantially in longitude (Gu and Zhang 2001, hereafter as GZ01), and therefore their relationships may not be the same in different parts of the Tropics. This difficulty can be overcome by using a new spectrum analysis method developed by GZ01. This method combines the conventional wavenumber-frequency spectrum analysis method with wavelet analysis. By doing so, power spectra of zonally propagating signals can be calculated as a function of longitude at the central latitude of the ITCZ, tracking its seasonal meridional migration. In GZ01, significant signals of zonally propagating synoptic-scale disturbances within the ITCZ have been identified globally as well as regionally; evident longitudinal dependence of the spectral properties of westward-propagating synoptic-scale disturbances are shown to exist not only in their dominant frequencies and wavenumbers, but also in their seasonal and interannual variability.

In the current study, relationships between the ITCZ and synoptic-scale disturbances are further explored in terms of their spatial distributions and temporal (annual and interannual) variability. Both the mean ITCZ and synoptic-scale disturbances are represented by outgoing longwave radiation (OLR) observed from polar-orbiting satellites. The local spectral power of the OLR is taken as a measure of activity of synoptic-scale disturbances. Mean OLR, when exhibiting a zonally elongated, narrow band in the Tropics, is taken as a representation of the ITCZ, with smaller OLR values interpreted as indicating a stronger ITCZ.

The data and method are briefly described in section 2. The results are presented in section 3, which show that the two observed fields are in tandem in a gross sense but exhibit many detailed discrepancies, some of them substantial. The implications of the results are discussed in section 4. The conclusion from the results is that the ITCZ does not owe its existence to, but is substantially affected by, westward-propagating synoptic-scale disturbances. Meanwhile, the observations are consistent with the theories that view the dynamical instability of the ITCZ as a possible cause of the synoptic-scale disturbances.

2. Data and methodology

A 20-yr (1 January 1979–31 December 1998) dataset of daily OLR from the National Oceanic and Atmospheric Administration (NOAA) polar-orbiting satellites (Gruber and Krueger 1984) is used to describe the ITCZ and to obtain spectral signals of deep convection of synoptic-scale disturbances. The data are on 2.5° × 2.5° grid. Missing data have been filled by interpolation (Liebmann and Smith 1996). OLR is commonly used as a proxy of deep convection in studies of large-scale convective activity and convectively coupled wave motions in the Tropics (e.g., Gruber 1974; Nitta et al. 1985; Tai and Ogura 1987; Wheeler and Kiladis 1999). It is recognized that atmospheric synoptic-scale disturbances cannot always be well represented by the OLR data. Disturbances can be present in the wind fields without signals in clouds. But this should not be a concern here because we are trying to compare both convective signals in the mean ITCZ and disturbances.

The mean ITCZ is subjectively defined as a zonally elongated narrow band of low monthly mean OLR in the Tropics. The seasonal migration of the ITCZ thus defined agrees well with that based on a precipitation dataset. This subjective definition of the ITCZ attempts to distinguish the ITCZ from other tropical convective zones that are less spatially oriented. The results to be shown in this study may be subject to modification if other quantities, such as surface wind convergence, are used to define the ITCZ.

The main analysis tool is a two-dimensional (2D) wavelet space–time spectrum analysis. This method, combining a wavelet transform in longitude and conventional wavenumber-frequency spectrum analysis, not only isolates westward/eastward-propagating components of a given field, but also provides their longitudinal and temporal dependence. To obtain spectral signals of zonally propagating synoptic-scale disturbances, this 2D wavelet spectral analysis is applied to a series of daily OLR time series $x(\lambda, \phi, t)$ ($\lambda$ is longitude, $\phi$ is latitude, and $t$ is time), each covering 92 days, centered at the middle of a calendar month and along a
Fig. 1. Wavenumber-frequency spectra of westward-propagating disturbances along the ITCZ during Jun–Nov, divided by corresponding background red spectra (from Gu and Zhang 2001). (a) Global signals, (b–d) regional signals in the (b) west-central Pacific (160°E–160°W), (c) eastern Pacific (160°–90°W), and (d) Atlantic–West Africa (55°W–0°E). Heavy dashed lines denote a significance threshold value of 1.1 for which the spectral signals are significantly above the background at the 95% level based on 500 dof. Unit of the ordinates is cycles day$^{-1}$ and the significance threshold gray scale is given below panel (d). The spectral signals within the boxes are used to represent westward-propagating synoptic-scale disturbances.

3. Results

A critical step in studying synoptic-scale disturbances is to identify their significant signals that are clearly distinguishable from background red noise. In a wavenumber-frequency spectral domain, one way of doing so is to estimate a background “red” spectrum by smoothing an original spectrum and to examine the ratio of the original to the background spectra; the original spectrum is considered significant only if this ratio is greater than a certain threshold (Wheeler and Kiladis 1999). Examples of such significant signals are given in Fig. 1, which shows the ratios of wavenumber-frequency spectra in OLR for westward-propagating disturbances along the latitudinal positions of the ITCZ in different zonal sectors during June–November, the season of maximum activity of the disturbances (GZ01). The significance threshold used here is 1.1 based on 500 degrees of freedom (dof). The strongest global signals are found within a zonal wavenumber range of $k = 6$ to 13 and a frequency range of $f = -0.1$ to $-0.28$ cycles day$^{-1}$ (Fig. 1a). Regional significant signals in the Pacific (Figs. 1b and 1c) are basically located in the same wavenumber and frequency ranges, although those in the west-central Pacific are much stronger than in the eastern Pacific. In the Atlantic (Fig. 1d), the dominant wavenumbers and frequencies of significant signals are shifted approximately to $k = 6–23$ and $f = -0.1$ to $-0.42$ cycles day$^{-1}$. Thus, the global significant signals of westward-propagating synoptic-scale disturbances shown in Fig. 1a are actually dominated by those in the Pacific, especially in the west-central Pacific, but do not accurately represent the spectral characteristics in the Atlantic. To emphasize the locality of westward-propagating synoptic-scale disturbances, two different wavenumber and frequency domains of spectral power will be separately used to represent the significant disturbance activities in the Pacific ($k = 6–15$, $f = -0.1$ to $-0.32$ cycles day$^{-1}$) and Atlantic ($k = 6–20$, $f = -0.1$ to $-0.4$ cycles day$^{-1}$) (see boxes in Figs. 1b–d).

Significant spectral signals in the selected frequency-
zonal wavenumber domains represent easterly waves and perhaps other types of waves. OLR signals corresponding to westward mixed Rossby–gravity waves and equatorial Rossby waves are found in the Tropics (Wheeler and Kiladis 1999). Although the spectra shown in Fig. 1 are along the latitudinal belts of the ITCZ, mostly north of the equator, signals of these waves may exist in our frequency and wavenumber domains. But exactly what types of waves these observed synoptic-scale disturbances represent is not important in this study.

The spectral signals for eastward-propagating synoptic-scale (6 ≤ k ≤ 15) disturbances are weak. Although these signals can be found along the ITCZ, they are significant only near the equator (not shown). Thus, eastward-propagating synoptic-scale disturbances should not be considered intrinsic to the ITCZ because during the boreal summer the ITCZ has a mean global latitudinal position at 8°N. In the rest of this study, the relationship between only westward-propagating synoptic-scale disturbances and the ITCZ will be emphasized.

a. Spatial distributions

The spatial distribution of total spectral power for westward-propagating synoptic-scale disturbances is plotted in Fig. 2a (contours) for the significant spectral signals within the specified synoptic domain (k = 6 to 20, f = -0.1 to -0.4 cycles day⁻¹) and through the entire data record. In the Pacific, the significant signals are within narrower wavenumber and frequency ranges (k = 6 to 13, f = -0.1 to -0.28 cycles day⁻¹) (Figs. 1b,c). However, their fundamental patterns in this wavenumber and frequency domain are the same as in Fig. 2a, even though the intensity is smaller by about 10%. This distribution is meant to compare to that of mean OLR (shades) representing the mean intensity of deep convection, including that associated with the ITCZ.

An evident band of maximum spectral power is found north of the equator continuously extending from the eastern Indian Ocean to the African continent. Notice that the band of maximum disturbances is not always collocated with regions of maximum mean convection, which underlines the distinction between significant signals of the disturbances and their deep-convective background. Three centers of maximum disturbance activity north of the equator can be identified in the western and eastern Pacific, and the Atlantic, respectively. There is a secondary maxima near the South Pacific convergence zone. These centers of maximum disturbances are located in regions where possible barotropic energy conversion from mean flow exists regarding a typical southwest–northeast tilt of these disturbances, as indicated by a negative meridional shear of mean zonal wind (Lau and Lau 1990).

The ITCZ is well defined as a zonally elongated, narrow band of low mean OLR north of the equator in the

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**Fig. 2.** (a) Time-mean OLR (W m⁻²) (shades) and spectral power (contours) for westward-propagating synoptic-scale disturbances within the wavenumber and frequency domain where the disturbance signals are statistically significant. (b) Time-mean OLR (W m⁻²) (shades) and covariance (contours) between monthly OLR and spectral power of the disturbances (k = 6–20, f = -0.1 to -0.4 cycles day⁻¹). Negative covariance (solid contours) indicates that the interaction between the ITCZ and disturbances are in phase.
central and eastern Pacific, and in the Atlantic. In the western Pacific, the ITCZ seems to be less well defined, even though a band of low OLR quasi-zonally elongated north of the equator is discernible. Based on the given criteria, no ITCZ can be identified at all over the Caribbean Sea and South America, interior Africa, and the Indian Ocean, even though strong convection may still exist.

In general, the distributions of the two fields agree well with each other in regions where the ITCZ is clearly identifiable. Their meridional extents roughly match and so do their zonal distributions. In regions where the disturbances are weak or not concentrated in a zonally oriented band, the ITCZ is less well defined even though mean convection may still be strong (e.g., Indian Ocean, and interior Africa). South of the equator, the ITCZ is in general absent or less well defined, and no band of the disturbances continuously extending in longitude exists. Local maxima are found in the southwestern Pacific and the southern Indian Ocean, but they are weak and zonally confined.

A more careful inspection, however, reveals detailed discrepancies between the two fields. The center of the ITCZ represented by the minimum mean OLR band almost coincides with the center of maximum disturbances in the central Pacific, but the two differ from each other toward both east and west where the center of the ITCZ tends to be on the equatorial side of the maximum disturbances. The discrepancy in the latitudinal locations is 2.5° or larger. This may explain the observations that within the ITCZ, signals of the disturbances are the strongest in the west-central Pacific but they become weaker toward the eastern and far western Pacific (GZ01). Further, the zonally elongated, narrow band of the disturbances extends continuously, fluctuating in intensity and latitude notwithstanding, from the western Pacific to West Africa. The ITCZ, however, appears to be interrupted over the landmass of South and Central America. In the southern Indian Ocean, there is a weak zonal band of spectral signals for the disturbances, whereas no similar distribution in mean OLR exists.

To further quantify the relationship between the ITCZ and disturbances, covariance between monthly disturbance activity (represented by the monthly mean spectral power of synoptic-scale disturbances) and the ITCZ (monthly mean OLR) at each grid point is plotted in Fig. 2b (contours). The time-mean OLR as in Fig. 2a is also plotted (shades) to indicate the mean position and intensity of the ITCZ. Negative covariance values imply the intensities of the ITCZ and disturbances vary in phase. Overall maximum (negative) covariance exists at latitudes north of the equator, near but not collocated with the center of the ITCZ. Very weak covariance is found in the regions close to the equator where maximum mean convection is. This, again, distinguishes the synoptic-scale disturbances from general red-noise background. Maximum (negative) covariance centers are located in the regions of the Southeast Asia, west of Central America, and Sahel of Africa. In between these centers, covariance is generally concentrated within a zonally elongated, narrow band. The axis of this band of large (negative) covariance is always at the polar side of the ITCZ, generally collocating with the band of the maximum intensity of the disturbances (Fig. 2a). However, the western Pacific and Atlantic maximum (negative) covariance in Fig. 2b is farther north of the maximum power of the disturbances in Fig. 2a. In the western Pacific, there is a single center of maximum power (Fig. 2a) but two centers of maximum covariance.

Near and south of the equator, the covariance is generally weaker. The large difference exists between Figs. 2a and 2b in the southwest Pacific and southeast Indian Ocean. Two weak maxima of spectral power are found in the tropical southeast Indian Ocean and southwest Pacific (Fig. 2a), though one center of covariance can be seen over northern Australia. This is a region where large potential instability is found but synoptic-scale disturbances are relatively weak, even in southern summer (Dickinson and Molinari 2000). Large covariance between the two in this region imply that synoptic-scale disturbances contribute to a sizable fraction of total convection there, even though both synoptic-scale disturbances and mean OLR are weak.

The covariance map in Fig. 2b includes contributions from both seasonal and interannual variability. The seasonal cycle is, however, the dominant factor. When the seasonal cycles in both fields are removed, their covariance becomes very weak. It is interesting that strong covariance is not always found where the means of the two fields are large. Over northern subtropical Africa and north Australia, in particular, the ITCZ and disturbances there are closely related in their low-frequency (mainly seasonal) variability (see section 3b), even though their means are both very weak (Fig. 2a).

It can be assumed from Fig. 2 that in a hypothetical world where the only convective activities are those associated with westward-propagating synoptic-scale disturbances, there would be an ITCZ. The mean latitudinal location of this ITCZ would be similar to that of the observed. This assumption would be invalid, however, if an existing ITCZ is a necessary condition for the westward-propagating synoptic-scale disturbances (see discussion in section 3d).

b. Seasonal cycle

Seasonal variations have been discussed for both the ITCZ (e.g., Mitchell and Wallace 1992; Waliser and Gautier 1993; Xie and Philander 1994; Wang 1994; Li and Philander 1997) and westward-propagating synoptic-scale disturbances (GZ01). In this section, the seasonal cycles in the ITCZ and disturbances are directly compared separately in four regions where the ITCZ is well defined: The west-central Pacific (150°E–170°W), the central Pacific (170°–130°W), the eastern Pacific
Fig. 3. Time–latitude distribution of OLR (W m$^{-2}$) (shades) and disturbance activity (contours) in (a) the west-central Pacific, (b) the central Pacific, (c) the eastern Pacific, and (d) the Atlantic–Africa. The “0” latitude corresponds to the mean locations of the ITCZ for each year.

(130°–90°W), and the Atlantic–West Africa (55°W–0°E). The Pacific sector is divided into three because of the apparent inhomogeneous relationship between the ITCZ and the disturbances seen in Fig. 2. Their seasonal cycles are plotted in Fig. 3. To limit the effect of interannual variability, the seasonal migrations are plotted relative to the annual mean latitude of the ITCZ. In all these regions, both the ITCZ and disturbances show similar seasonal cycles: Their maximum intensities reach the peaks during July–November (with an exception for the ITCZ in the Atlantic) when they are farthest away from the equator; they are the weakest during January–March when they are the closest to the equator. Correlation coefficients were calculated for the seasonal cycles of the latitude and intensity of maximum disturbance activity ($Y_{DIST}$ and $W_{DIST}$), the intensity of the disturbances along the center of the ITCZ ($W_{ITCZ}$), and the latitude and intensity of the ITCZ ($Y_{ITCZ}$ and $P_{ITCZ}$). Results are listed in Table 1. Correlation between the seasonal cycles in the latitude and intensity is significant in all four regions for the disturbances ($W_{DIST}$ and $Y_{DIST}$). It is, however, significant in none of the regions for the

Table 1. Correlation coefficients for the annual cycles in different variables (see text). From the second to the sixth rows: Maximum disturbance activity ($W_{DIST}$) and its latitudinal location ($Y_{DIST}$); strength of the ITCZ ($P_{ITCZ}$) and latitudinal location of the ITCZ ($Y_{ITCZ}$); latitudes of maximum disturbance activity ($Y_{DIST}$) and the ITCZ ($Y_{ITCZ}$); disturbance activity along the ITCZ ($W_{ITCZ}$) and latitudinal location of the ITCZ ($Y_{ITCZ}$); disturbance activity along the ITCZ ($W_{ITCZ}$) and strength of the ITCZ ($P_{ITCZ}$). Bold type indicates that correlation is significant at the 5% confidence level (±0.44 based on 18 dof).

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<td>$W_{DIST}$ and $Y_{DIST}$</td>
<td>0.66</td>
<td>0.55</td>
<td>0.74</td>
<td>0.83</td>
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<td>$P_{ITCZ}$ and $Y_{ITCZ}$</td>
<td>−0.06</td>
<td>0.11</td>
<td>−0.41</td>
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<td>0.58</td>
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<td>$W_{ITCZ}$ and $Y_{ITCZ}$</td>
<td>0.77</td>
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<td>$W_{ITCZ}$ and $P_{ITCZ}$</td>
<td>−0.23</td>
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ITCZ \((P_{\text{ITCZ}} \text{ and } Y_{\text{ITCZ}})\). The relationship between the ITCZ and disturbances varies in longitude: The seasonal correlation is significant in all four regions between their latitudes \((Y_{\text{DIST}} \text{ and } Y_{\text{ITCZ}})\); it is so only in some of the regions between the ITCZ intensity \((P_{\text{ITCZ}})\) or latitude \((Y_{\text{ITCZ}})\) and disturbance intensity \((W_{\text{DIST}})\). The seasonal relationship between the two fields seems to be the best in the eastern Pacific.

The seasonal cycle in the ITCZ appears to be weaker than that in the disturbances in both intensity and latitudinal displacement, particularly in the Atlantic. This is the main reason why the correlation coefficient between the intensity of the disturbances and the ITCZ is almost zero (Table 1, \(W_{\text{ITCZ}} \text{ and } P_{\text{ITCZ}}\)). The latitudes of maximum disturbances and the ITCZ (mean minimum OLR) are very similar when they are close to the equator but differ significantly when they are farther away from the equator during their peak season (July–November), particularly in the Atlantic. In other words, the seasonal migration in latitude is larger for the disturbances than for the ITCZ, even though they vary in tandem and hold high correlation (Table 1, \(Y_{\text{DIST}} \text{ and } Y_{\text{ITCZ}}\)). Therefore, the seasonal cycle in the ITCZ cannot be fully explained in terms of the seasonal cycle in the disturbances. The possible physical implications will be discussed in section 4.

In our hypothetical world where all convective activities were associated with westward-propagating synoptic-scale disturbances, the assumed ITCZ would have an annual cycle with a similar phase, but stronger amplitude and meridional migration in comparison to those observed.

c. Interannual variability

Interannual fluctuations in the ITCZ and disturbances are examined also in terms of their latitudinal positions and intensity. It would not be surprising if the ITCZ and disturbances vary in concert interannually, either because they are causally related or because they respond similarly to the influence of El Niño–Southern Oscillation (ENSO). Previous studies do show the impact of the ENSO events on the ITCZ and tropical convection (e.g., Waliser and Gautier 1993; Matthews and Kiladis 1999). It would be intriguing and perhaps more interesting if the ITCZ and disturbances show different interannual characteristics.

To exclude the annual signals as seen from the previous subsection, interannual time series of the latitudinal positions of the ITCZ and maximum disturbances are compared in seasonal means for June–November, the peak season of the disturbance activity, and December–May, the season of minimum disturbances (Fig. 4). The three strongest warm events of ENSO during the period are those in 1982–83, 1986–87, and 1997–98. The interannual displacements in latitude for both ITCZ and disturbances show strong seasonal and regional dependence. In the Pacific, both ITCZ and disturbances tend to move toward the equator near the peak of warm phases of ENSO (December–May) when the equatorial cold tongue is the weakest or disappears (Figs. 4b,d,f). This ENSO-related interannual variability is weaker during June–November for the ITCZ and almost does not exist for the disturbances (Figs. 4a,c,e). The interannual variations in the Atlantic are weaker than in the Pacific and are not apparently related to ENSO (Figs. 4g,h).

The intensities of the ITCZ and disturbances also show interannual variability that is seasonally and regionally dependent (Fig. 5). The variability in the ITCZ is stronger during December–May than June–November. The phase seems to be opposite in the Pacific and Atlantic. During warm phases of ENSO, the ITCZ tends to be stronger in the Pacific but weaker in the Atlantic. It is not obvious how the interannual variability of the disturbance intensity is related to ENSO.

To further quantify the relationships between the interannual variations in the ITCZ and disturbances and between them and ENSO, interannual correlation is calculated for these quantities for each month (Fig. 6). A Southern Oscillation index (SOI) is used to describe the ENSO variability. ITCZ amplitudes are represented by changing the signs of OLR anomalies (i.e., positive OLR anomalies correspond to stronger ITCZs). The interannual latitudinal displacement of the disturbances is significantly correlated with the SOI only during boreal winter and in the Pacific (Fig. 6a). For the ITCZ, significant correlation can be found in both winter and summer. The positive correlation indicates that the latitudinal locations for both the ITCZ and disturbances tend to be farther away from the equator during cold phases of ENSO (positive anomalies in the SOI), as seen in Fig. 4. If there is any significant correlation between the latitudinal location of the ITCZ and ENSO in the Atlantic, it is in the opposite sign, as in the Pacific. Figures 6a and 6b suggest that the latitudinal location of the disturbances is not as susceptible to influences of the ENSO-related variability in SST as that of the ITCZ.

The seasonality in the intensity of the disturbances is more complicated (Fig. 6c). During warm phases of ENSO, the intensity appears to be greater (negative correlation) in boreal summer for the eastern Pacific disturbances, but less (positive correlation) in boreal winter for the western Pacific disturbances and in boreal summer for the Atlantic disturbances. A stronger ITCZ can be found in the Pacific during warm phases of ENSO, most in boreal summer for the western Pacific ITCZ and in boreal winter for the eastern Pacific ITCZ, but without much seasonality for the central Pacific ITCZ (Fig. 6d). In the Atlantic, the ITCZ tends to be weaker during warm phases mostly in boreal winter.

The interannual displacements in the latitudinal locations of the ITCZ and disturbances are always positively related to each other, most noticeably in boreal winter (Fig. 6e). The interannual variations in the intensity of the ITCZ and disturbances are also positively
related but only during boreal winter (Fig. 6f). The relationship between the interannual variations in the latitudinal locations of the ITCZ and disturbances can be interpreted as a manifestation of the similar influences on them by ENSO (Figs. 6a,b). But it is less so for the relationship between their intensity. For boreal winter, for example, while the eastern Pacific ITCZ tends to be stronger during warm phases of ENSO (Fig. 6d), there are no such signals at all for eastern Pacific disturbances (Fig. 6c). Therefore, the positive correlation between the intensities of the ITCZ and disturbances during boreal winter can be taken as an indication of their causal relationship. A general impression one might get from Figs. 6e and 6d would be, however, that the interannual relationships between the ITCZ and disturbances are much weaker than their annual relationships (Table 1). The assumed ITCZ solely due to the synoptic-scale disturbances in our hypothetical world would, therefore, have a quite different interannual variability from the observed.

d. Comparison with theories

As mentioned in section 1, Holton et al. (1971) proposed a theory relating the ITCZ to zonally propagating atmospheric transients. In this theory, the latitudinal location of the ITCZ is determined by a critical latitude where the Coriolis frequency equals the Doppler-shifted frequency of zonally propagating waves and boundary layer convergence reaches its peak. Using a numerical model, Chang (1973) modified the theory by including the efficiency of atmospheric response to diabatic heating. This modification shifts the ITCZ to the north of the critical latitude by $1^\circ$-$3^\circ$. Holton et al. (1971) suggested that a large number of simultaneous observations of Doppler-shifted frequencies of the waves and the latitude of the ITCZ, in different regions and on different timescales, are necessary to verify the theories. Such observations were not available at that time. The theories can now be validated against our 20-yr observations of the zonally propagating wave activities.

According to Holton et al. (1971), the key factor to a well-defined ITCZ, mostly over the ocean, is the existence of zonally propagating waves that must exhibit significant spectral peaks. Thus, a crucial step in the validation is to identify such significant spectral peaks. The red nature of the OLR spectra in both wavenumber and frequency has been previously shown (e.g., Gruber 1974). In this study, as described before, the background

![Fig. 4. Latitudes of maximum spectral power (solid lines) and minimum OLR (dashed lines) of the disturbances as a function of year in the (a) and (b) west-central Pacific, (c) and (d) the central Pacific, (e) and (f) the eastern Pacific, and (g) and (h) the Atlantic–West Africa region.](image-url)
red spectra are estimated by smoothing the original power spectrum, and the statistically significant spectral signals in the wavenumber-frequency space are determined by comparing the original spectral power against its corresponding background red noise (Fig. 1), as in Wheeler and Kiladis (1999). The disturbance signals are considered significant only if the ratio of the two is larger than 1.1 (based on 500 dof).

Among all significant OLR signals, only those that are of westward propagation and synoptic scales show maximum power at the similar latitudes of the ITCZ and thus can be considered possibly intrinsic to the ITCZ. Others either have too low frequencies or maximum power only near the equator. The theory is therefore validated only against westward-propagating synoptic-scale disturbances. Their frequencies and the Coriolis frequency are used to determine the “observed critical latitude” ($Y_H$). A “predicted latitude of the ITCZ” ($Y_C$) is one that is $1^\circ$ to $3^\circ$ north of $Y_H$, based on Chang (1973). Here $Y_H$ and $Y_C$ are then compared to the latitudes of observed maximum disturbances ($Y_{DIST}$) and the latitude of the observed mean ITCZ ($Y_{ITCZ}$) in Table 2.

Significant OLR signals of westward-propagating synoptic-scale disturbances are observed within a frequency range of 0.1–0.4 cycles day$^{-1}$ (Fig. 1). The corresponding critical latitudes ($Y_H$) are between $3^\circ$ and $11.4^\circ$N. This is roughly the same range for the seasonal migration of the ITCZs in the Pacific and Atlantic, where the ITCZs can be clearly identified (Fig. 2). However, the observed critical latitudes $Y_H$ are generally different from the observed ITCZ latitude $Y_{ITCZ}$ and the latitudes of maximum disturbances $Y_{DIST}$ by up to $5^\circ$–$6^\circ$ in the Pacific (Table 2). In the Atlantic, the observed critical latitude $Y_H$ is similar to the observed ITCZ latitude $Y_{ITCZ}$. The modification by Chang (1973) makes the predicted ITCZ latitude $Y_C$ closer to the latitudes of maximum disturbances $Y_{DIST}$. Generally, observed critical latitudes $Y_H$ and ITCZ latitudes $Y_C$ are apart from the observed latitudes of the the maximum disturbances $Y_{DIST}$ during June–November, the season of the strongest disturbance activity (Fig. 4).

In a parallel way of evaluating the theories, frequencies of the disturbances that are predicted to be related to the ITCZ by Holton et al. (1971) are estimated using the Coriolis frequencies at the observed latitudes of the ITCZs ($F_H$). A $2^\circ$ latitudinal adjustment suggested by Chang (1973) is included to obtain another estimate

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**Fig. 5.** Spectral power of the disturbances (solid lines) and OLR (W m$^{-2}$) (dashed lines) along the center of the ITCZ as function of year in the (a) and (b) west-central Pacific, (c) and (d) the central Pacific, (e) and (f) the eastern Pacific, and (g) and (h) the Atlantic–West Africa region.
Fig. 6. Correlation between different variables on interannual timescale: "O" in the west-central Pacific, "×" in the central Pacific, "+" in the eastern Pacific, "*" in the Atlantic–West Africa region. ITCZ amplitudes are represented by changing the signs of OLR anomalies. Positive OLR anomalies correspond to stronger ITCZs. The 5% confidence level of correlation coefficients (18 dof) is ±0.44 (dashed lines).

These predicted frequencies of disturbances are then compared to the observed \( F_o \) in Table 3. The predicted frequencies \( F_H \) based on Holton et al. (1971) are generally higher than the observed \( F_o \). The predicted frequencies \( F_C \) based on Chang (1973) are much closer to the observed. According to the theories, a higher critical latitude should correspond to a higher frequency of disturbances. It is not the case in the observations. The latitudes of maximum disturbances \( \bar{\gamma}_{\text{DIST}} \) are lower in the Atlantic than in the Pacific (Table 2), but the observed frequencies of synoptic-scale disturbances in the Atlantic is higher (Fig. 1). Also, according to the theory, an annual meridional migration in the disturbances and therefore in the ITCZ should correspond to an annual cycle in the dominant frequencies of the disturbances. While an annual variation in the frequencies of the disturbances has been observed (GZ01), it is not sufficiently large to account for the observed annual meridional migration of either the disturbances or the ITCZ.

Table 2. Observed dominant frequencies \( (F_o) \) of westward-propagating synoptic-scale disturbances, latitudinal locations of the ITCZ predicted by Holton et al. (1971) \( (Y_H) \) and modified by Chang (1973) \( (Y_C) \) \( 1°-3° \) are added, and observed annual mean (Jun–Nov mean in italics) latitudes of the ITCZ \( \bar{Y}_{\text{ITCZ}} \) and the maximum disturbance activity \( \bar{F}_{\text{DIST}} \).

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<tr>
<td>( F_o ) (cycles day(^{-1}))</td>
<td>-0.23</td>
<td>-0.24</td>
<td>-0.24</td>
<td>-0.29</td>
</tr>
<tr>
<td>( Y_H ) (°N)</td>
<td>6.6</td>
<td>6.9</td>
<td>6.9</td>
<td>8.3</td>
</tr>
<tr>
<td>( Y_C ) (°N)</td>
<td>7.6–9.6</td>
<td>7.9–9.9</td>
<td>7.9–9.9</td>
<td>9.3–11.3</td>
</tr>
<tr>
<td>( \bar{Y}_{\text{ITCZ}} ) (°N)</td>
<td>7.6</td>
<td>9.5</td>
<td>10.5</td>
<td>7.8</td>
</tr>
<tr>
<td>( \bar{F}_{\text{DIST}} ) (°N)</td>
<td>10.1</td>
<td>10.0</td>
<td>11.4</td>
<td>9.3</td>
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According to the theories of Holton et al. (1971) and Chang (1973), the frequencies of these disturbances should be relative to the mean zonal wind in the boundary layer (Doppler-shifted or intrinsic), \( f_{\text{doppler}} = f_{ob} - \bar{U}k \) (where \( f_{ob} \) is observed frequency, \( \bar{U} \) mean zonal wind in the boundary layer, and \( k \) wavenumber). In the Pacific and Atlantic where we are primarily concentrated on, zonal easterly flow \( (\bar{U} < 0 \text{ and } |\bar{U}| \leq 6 \text{ m/s}) \).
s\(^{-1}\)) in the boundary layer generally dominates during boreal summer, the most intense disturbance season, except near and over Central America and West African continent where weak westerly flow exists due to the impact of regional monsoon circulations. Thus, the magnitude of Doppler-shifted frequencies will be smaller than the (observed) ones with respect to the surface, for example, \(\sqrt{\langle U \rangle} = 5\) m s\(^{-1}\) and \(k = 9\) can make a difference about 0.1 cycles day\(^{-1}\) between \(f_{\text{doppler}}\) and \(f_{\text{obs}}\), which means that the predicted latitudes of the ITCZ by Holton et al. (1971) and Chang (1973) decrease by 2°–3°, much closer to the equator. Therefore, the observed and predicted ITCZs will be farther apart when the Doppler-shifted effect of the mean wind is considered.

It should be pointed out here that the critical latitude theory of the ITCZ was built upon the wind field of zonally propagating synoptic-scale disturbances. In the above observation–theory comparisons, it has to be assumed that the detected westward-propagating synoptic-scale signals in OLR (i.e., their frequencies and zonal wavenumbers) do represent westward-propagating synoptic-scale signals in the wind field. In fact, maximum westward-propagating synoptic-scale signals in both winds and cloud (OLR) are collocated in the western and eastern Pacific (e.g., Lau and Lau 1990). In the Atlantic, maximum signals in winds are north of those in cloud, possibly because the dry air advected from the Sahara desert and the disturbances not coupling with moist convection (e.g., Duvel 1990).

The discrepancies between the ITCZ and westward-propagating synoptic-scale disturbances warrant a brief review of other theories that attempt to explain one in terms of the other. It has been suggested that westward-propagating disturbances along the latitudinal belt of the ITCZ could be caused by the (barotropic and/or baroclinic) instability of the ITCZ (Bates 1970, 1972; Kuo 1978; Hack et al. 1989; Schubert et al. 1991; Ferreira and Schubert 1997). An unstable environment for the disturbances can be produced by moist convection within the ITCZ (Ferreira and Schubert 1997). Through a zonal balance model, Hack et al. (1989) and Schubert et al. (1991) showed that an unstable lower-tropospheric region (with a reversal of the meridional gradient of potential vorticity) could be formed poleward of the ITCZ where synoptic-scale disturbances are generated. This model result is generally supported by observations (Molinari et al. 1997, 2000). Locations of the disturbances predicted by this theory actually coincide with the observed separation of latitudes of the ITCZ and maximum disturbance activity in this study (Figs. 2 and 3). The interannual variations in the disturbances are perhaps more directly related to the ITCZ than ENSO (Fig. 6). Thus, the results here are consistent with the self-destruction and formation theory of the ITCZ, suggesting that the ITCZ breakdown mechanism is a cause for some of the synoptic-scale disturbances.

### Table 3. Observed annual mean latitudes (and seasonal variation range in italics) of the maximum disturbance activity (\(\overline{F}_{\text{max}}\)) and the ITCZ (\(\overline{F}_{\text{ITCZ}}\)), and the dominant frequencies predicted according to Holton et al. (1971) (\(F_p\)) and Chang (1973) (\(F_s\)) (2° latitudinal difference is assumed), observed frequency (\(F_s\)) (and range) corresponding to the significant signals of westward-propagating synoptic-scale disturbances.

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<tr>
<td>(\overline{F}_{\text{max}}) (N)</td>
<td>10.1</td>
<td>10.0</td>
<td>11.4</td>
<td>9.3</td>
</tr>
<tr>
<td>(F_p) (cycles day(^{-1}))</td>
<td>(-0.26) to (-0.35)</td>
<td>(-0.28) to (-0.45)</td>
<td>(-0.28) to (-0.52)</td>
<td>(-0.4) to (-0.33)</td>
</tr>
<tr>
<td>(\overline{F}_{\text{ITCZ}}) (N)</td>
<td>7.6</td>
<td>9.5</td>
<td>10.5</td>
<td>7.8</td>
</tr>
<tr>
<td>(F_s) (cycles day(^{-1}))</td>
<td>(-0.11) to (-0.33)</td>
<td>(-0.22) to (-0.32)</td>
<td>(-0.2) to (-0.4)</td>
<td>(-0.08) to (-0.32)</td>
</tr>
<tr>
<td>(F_s) (cycles day(^{-1}))</td>
<td>(-0.23)</td>
<td>(-0.24)</td>
<td>(-0.24)</td>
<td>(-0.29)</td>
</tr>
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### 4. Summary and discussion

This study examines the observed relationship between the ITCZ and tropical synoptic-scale disturbances. The dominant components of synoptic-scale disturbances in the latitude belt of the ITCZ (0°–20°N) are those propagating westward and within ranges of zonal wavenumber \(k = 6\)–15, frequency \(f = -0.1\) to \(-0.32\) cycles day\(^{-1}\) in the Pacific and \(k = 6\)–20, \(f = -0.1\) to \(-0.4\) cycles day\(^{-1}\) in the Atlantic. These disturbances are represented by significant spectral power of daily OLR calculated using a new spectrum method. This method combines the traditional frequency-wavenumber spectrum analysis and wavelet analysis to resolve longitudinal distributions of spectra for zonally propagating disturbances (GZ01). The significance of a spectrum is evaluated by comparing the total spectrum to its smoothed background red spectrum, as in Wheeler and Kiladis (1999). The ITCZ is represented by a zonally elongated, narrow band of low mean OLR in the Tropics. Comparisons were made between the ITCZ and the disturbances for their intensities and latitudinal locations. The observations were also compared to the theories that attribute the ITCZ to those synoptic-scale disturbances. The purpose of these comparisons is to
explore the degree to which the ITCZ is causally related to the disturbances. The main results are the following:

- Distributions and variability of the ITCZ and maximum activity of the disturbances agree generally well with each other in certain aspects. They are both located within the same latitudinal range (5°–15°N). The disturbances are strong primarily at longitudes where the ITCZ is well defined (Fig. 2). Their seasonal migrations in latitude are in phase (Fig. 3).
- Detailed, quantitative discrepancies, however, exist between the two fields, some of them substantial (Figs. 2 and 3). The most marked discrepancies are (i) the center of the ITCZ tends to be located about 2°–4° apart from the disturbance maximum on its equatorial side (Figs. 2 and 3); (ii) the annual cycle of the ITCZ is much weaker than the disturbances; and (iii) the seasonal and interannual variations in the strength of the ITCZ and the disturbances are not closely related to each other (Table 1 and Fig. 6f).
- The relationships between the ITCZ and synoptic-scale disturbances vary zonally (Table 1, Fig. 6). Their correlation for the annual cycle is the best in the eastern Pacific.
- The theory of Holton et al. (1971), modified by Chang (1973), can explain the approximate mean latitudinal location but not the zonal and temporal variability of the ITCZ (section 3d).

These results suggest that the ITCZ and synoptic-scale disturbances may be causally related to each other but only to a limited degree. Their detailed discrepancies indicate that the main features of the ITCZ (e.g., the preferred latitude, annual, and interannual variations) cannot be satisfactorily explained in terms of the disturbances alone. Other factors must also be important. The theories of Holton et al. (1971) and Chang (1973) are correct in the sense that, in the absence of all other factors, an ITCZ would still exist solely because of the westward-propagating synoptic-scale disturbances. This “wave-ITCZ,” however, would be weaker, located farther away from the equator, and undergo a stronger seasonal cycle and a different interannual variation in comparison to the observed ITCZ. On the other hand, the observations in this study suggest that an ITCZ would also exist in the absence of any synoptic-scale disturbances. This “non-wave-ITCZ” would resemble more the observed than would the wave-ITCZ. We conclude, therefore, that in reality the ITCZ does not owe its existence to the synoptic-scale disturbances but can be substantially modulated by them.

This conclusion does not contradict the theory that the synoptic-scale disturbances are generated by the ITCZ breakdown (see section 3d). The long-term effect of such ITCZ-induced disturbances would be to weaken the ITCZ and make it slightly widened northward. This mechanism cannot, however, explain all the observed features of the synoptic-scale disturbances. The stronger annual cycle in the synoptic-scale disturbances than in the ITCZ (Fig. 4) and the regions where obvious signals in the disturbances exist without any clear sign of an ITCZ (Fig. 2) are cases in point.

Our conclusion raises an intriguing question: What are the main convective constituents of the ITCZ if it is not the westward-propagating synoptic-scale disturbances? To quantify the contribution to the ITCZ from the westward-propagating synoptic-scale disturbances relative to other disturbances, the total subseasonal (within a three-month running window) convective perturbation signals (in OLR) are divided into three groups. The first includes only the westward-propagating synoptic-scale disturbances that are significant based on the criterion described in section 3. The second represents all low-frequency, subseasonal ($|f| = 0.022–0.2 \text{cycles day}^{-1}$; $k \leq 5$) signals, including both westward- and eastward-propagating ones. In this group, significant signals are mainly those of planetary scales ($k \leq 3$). The third represents all high-frequency ($|f| \geq 0.1 \text{cycles day}^{-1}$) signals, including eastward-propagating ones excluding those in the first group. Most signals in this group are not significant and can thus be categorized as random. Significant eastward-propagating synoptic-scale signals do exist but are much weaker than those westward-propagating and low-frequency ones (GZ01). The significant planetary-scale signals excluded, the low- and high-frequency groups constitute the red background noise. The total subseasonal variance and the variance of the three groups of disturbances are plotted in Fig. 7. A rough estimate indicates that of the total subseasonal variance of ITCZ convection, the westward-propagating synoptic-scale disturbances may contribute 15%, the low-frequency disturbances up to 5%, and the high-frequency disturbances about 80%. Salby et al. (1991) reported that the total variance of brightness temperature along the ITCZ is dominated by its synoptic-scale components; the low-frequency (period $> 10$ days) contribution is quite small. This is consistent with our analysis. Most (84%) of the dominant synoptic-scale signals in cloudiness or convection, measured by brightness temperature in Salby et al. (1991) and by OLR in this study, are random, according to our analysis. Convection in the mean ITCZ is mainly constituted by these random signals. They must be associated with the large-scale mean circulation. The large-scale environment for this random convection could be maintained by dynamic and thermodynamic fields that vary at much lower frequencies. The zonal asymmetry of the ITCZ is thus determined by the zonal variations in these large-scale fields. The dominance of these random disturbances in total convection of the mean ITCZ suggests that it is crucial to understand the mechanisms for their large-scale environment in order to explain the ITCZ. It is unclear to what degree the westward-propagating synoptic-scale disturbances in convection are also influenced by the same large-scale environment.
FIG. 7. Mean OLR (W m$^{-2}$) (shades) and variances (contours). (a) Total subseasonal variance, (b) variance of westward-propagating synoptic-scale disturbances ($k = 6-20, f = -0.1$ to $-0.4$ cycles day$^{-1}$), (c) variance of low-frequency disturbances ($k = 1-5, |f| = 0.022$ to 0.2 cycles day$^{-1}$), (d) variance of high-frequency disturbances. (See text.)

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