Idealized Simulations of the Intertropical Convergence Zone and Its Multilevel Flows

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ABSTRACT

A mesoscale numerical model with an idealized tropical channel environment is used to study the dynamics of intertropical convergence zones (ITCZs) and the recently identified shallow return flow (SRF) and mid-level inflow (MLI). Four idealized sea surface temperature (SST) distributions are used: a meridionally symmetric SST profile with a sharply peaked SST maximum at the equator, a similar profile with the maximum SST shifted off the equator, a cosine-shaped SST profile with zero gradient at the equator, and an idealized SST profile modeled after the observed SST of the eastern Pacific.

The simulations show that both the SRF and the MLI are robust features of the ITCZ. The prior result that the SRF is a sea-breeze-like response to surface temperature gradients is further supported, whereas the MLI is caused by the upper-level maxima in diabatic heating and vertical motion. Simulations with the SST maximum at the equator generate long-lasting, convectively coupled Kelvin waves. When the SST maximum is off the equator, the meridional circulations become highly asymmetric with strong cross-equatorial flow. Tropical cyclones are frequently generated by dynamic instability of the off-equatorial ITCZs.

The contributions of the multilevel circulations to regional budgets of mass, water, and moist static energy (MSE) are computed. About 10% of the total water transported into the ITCZ region is transported out by the SRF. The water transport of the MLI is minimal, but its mass and MSE transports are significant, accounting for about 1/3 of the mass and MSE entering the ITCZ region. Eddy fluxes are found to be a large component of the net vertically integrated transport of MSE out of the ITCZ.

1. Introduction

a. The intertropical convergence zone and its multilevel flows

The predominant large-scale view of the intertropical convergence zone (ITCZ) describes it as a belt of quasi-steady upward motion between the two large-scale overturnings of the northern and southern Hadley circulations, as illustrated in numerous texts (Peixoto and Oort 1992; Grotjahn 1993; James 1994; Holton 2004; Vallis 2006). However, like so many large-scale features of the atmosphere and ocean circulations, a detailed examination of the ITCZ reveals that its structure and dynamics are more complex than conveyed by this simple view. This more detailed view of the ITCZ has become possible in recent years because of a number of successful field experiments such as the Tropical Ocean Global Atmosphere Coupled Ocean–Atmosphere Research Experiment (TOGA COARE; Webster and Lukas 1992) and the East Pacific Investigation of Climate (EPIC; Raymond et al. 2004) and because of the increasing availability, quality, and resolution of “reanalysis” datasets.

One such feature of the ITCZ that is more complex than previously thought is the vertical structure of the large-scale overturning circulation. By synthesizing observations from the EPIC field project and from previously published studies, Zhang et al. (2004) and McGauley et al. (2004) found a low-level return flow coming southward out of the eastern Pacific ITCZ region, just above the boundary layer, that is completely distinct from the upper-level outflow (ULO) driven by deep convection. By diagnosing the pressure and temperature fields required to drive this so-called shallow return flow (SRF), Nolan et al. (2007, hereafter NZC07) showed that this circulation is in fact a local response to large meridional gradients of surface temperature and pressure, much like a sea-breeze circulation. Using idealized simulations of an ITCZ with a full-physics, mesoscale model, NZC07

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found that the shallow return flow is modulated by deep convection, which is strongest when convection is absent and weaker or nonexistent when deep convection is present.

Zhang et al. (2008) continued investigation of shallow return flows around ITCZ regions by considering their strengths and seasonalities in three different re-analysis datasets and by comparing the reanalyses directly to observations. Although some large discrepancies between the different datasets and the observations were identified, many consistent aspects were found. A robust shallow return flow exists in the Northern Hemisphere summer for the eastern Pacific and eastern Atlantic ITCZs. The strongest shallow return flow is associated with the West African monsoon, where it shows a fundamentally different structure. In that case, the boundary layer flow travels entirely through the ITCZ and into the “heat low” over the hot landmass. The flow rises a few kilometers and then turns southward as a shallow return flow that travels again through the ITCZ and back out over the ocean, again just above the boundary layer [see Fig. 24 of Zhang et al. (2008) for an illustration]. However, in this study we will continue to focus on ITCZ circulations that are not modified by land.

Figure 1 shows two plots of temporally and zonally averaged meridional wind $v$ as a function of height and meridional distance (expressed as latitude in the second plot). Figure 1a is reproduced from Fig. 5 of NZC07, derived from their idealized ITCZ simulation in a tropical “half channel” model. The left boundary corresponds to a free-slip vertical wall at the equator. The vertical profile of $v$ near the equator shows four flows: the boundary layer inflow (BLI), the SRF, the midlevel inflow (MLI), and the ULO. The BLI and ULO are well-established features of the tropical circulation, and the existence of the SRF has recently been established. What about the MLI? The influence of midlevel dry intrusions as a sporadic, synoptic-scale phenomenon that influences convection in the ITCZ has been documented in a number of previous studies (Numaguti et al. 1995; Yoneyama and Parsons 1999; Waugh and Polvani 2000; Cau et al. 2005; Zuidema et al. 2006). The MLI as a persistent feature was noted by Mapes (2001) in vertical profiles of meridional wind at the equator derived from reanalyses. Evidence of this feature is apparent in a number of observational analyses of tropical convection, such as in Johnson et al. (1999, their Fig. 13a) and Takayabu et al. (2006, their Fig. 17). A spectacular example of persistent, multilevel flow across the equator was shown by Madden and Zipser (1970) from rawinsonde data from the Line Islands Experiment in the central Pacific. Folkins et al. (2008) also describe multilevel flows similar to the SRF and MLI around large-scale convective complexes in both the Pacific and Caribbean.

As shown by Zhang et al. (2008), the MLI is also evident to varying degrees in three widely used reanalysis datasets. Perhaps the most robust example can be found in the 40-yr European Centre for Medium-Range Weather Forecasts (ECMWF) Re-Analysis (ERA-40) of the meridional wind in the eastern Pacific region (90°–160°W) for the month of July (Fig. 1b). The same four meridional flows are clearly present, in this case flowing across the equator. In comparison to the idealized simulation, the MLI in the reanalysis is less robust than the other circulations; however, between 90° and 120°W, the mean MLI flow speed across the equator does exceed 1 m s$^{-1}$ (see Fig. 9 of Zhang et al. 2008).
b. Current issues and open questions

The results of NZC07 can be summarized as follows: the SRF is primarily a sea-breeze-like response to large meridional temperature gradients; the SRF is modulated by deep convection; and the SRF plays a significant role in the meridional transport of water out of the ITCZ.

Given these results and the only cursory consideration given to the MLI in NZC07 and Zhang et al. (2008), further study is warranted. In particular, to better determine the roles that both the SRF and the MLI play in regulating convection in the ITCZ, it will be helpful to represent and analyze them in a more realistic framework. For this purpose, we have expanded our modeling configuration across the equator to allow for asymmetric convection and cross-equatorial flow. We will also use meridional profiles of SST that are smoothly varying, one of which is modeled closely after the SST profile of the eastern Pacific.

Other features of the idealized ITCZ will be explored with these more realistic configurations. For example, NZC07 found that convection was strongly organized by a single, robust convectively coupled Kelvin wave (CCKW). That convection is organized into a dominant, domain-length CCKW was also found for a global aquaplanet simulation by Frierson (2007). In our case, it will be shown that this wave is not so dominant when the free-slip wall at the equator is removed and that an equatorial wave feature similar to the mixed Rossby–gravity wave will appear.

To consider the potential effects of the multilevel circulations on convection in the ITCZ, NZC07 performed a detailed budget analysis of water (vapor and condensate) transport in and out of the ITCZ region. Although revisiting the water budget in more realistic ITCZ configurations is a logical step (and will be shown below), more can be gained from a budget analysis of the moist static energy (MSE). Since the seminal paper below), more can be gained from a budget analysis of the moist static energy (MSE). Since the seminal paper by Neelin and Held (1987), MSE budgets have been advocated as an approach to understanding both convecting and nonconvecting regions in the tropics (see also Emanuel et al. 1994; Raymond et al. 2009).

Section 2 describes the updated model, domains, and meridional profiles of SST used in the simulations. Section 3 documents the simulated ITCZs and the structure of their multilevel flows for SST profiles that are symmetric about the equator. Section 4 does the same for simulations with SST profiles that are asymmetric about the equator. Section 5 discusses the causes of the MLI and the implications of the multilevel flows for budgets of mass, water, and MSE in the ITCZ. Section 6 provides a summarizing discussion.

2. The model, domain, and boundary conditions

a. The numerical model and parameterizations

The model used for all simulations in this study was version 2.1.2 of the Weather Research and Forecast model (WRF). WRF is a mesoscale model that simulates fully compressible atmospheric dynamics with high-order advection schemes, using $\eta = p_x/p_y$ as the vertical coordinate, where $p_x$ and $p_y$ are the hydrostatic pressure and surface hydrostatic pressure, respectively (Skamarock et al. 2005).

For the best comparison to previous results, the horizontal and vertical resolutions are equal to those used in NZC07, with $\Delta x = \Delta y = 20.87\text{ km}$ and 40 vertical levels equally spaced in the WRF model coordinates $\eta$ between the surface and approximately 20-km altitude. Also used are the same WRF single moment 5-class microphysics scheme (WSM5, Hong et al. 2004), the Yonsei University planetary boundary layer scheme (YSU; Noh et al. 2003; Hong et al. 2006), and the rapid radiative transfer model longwave radiation scheme (RRTM; Mlawer et al. 1997; Iacono et al. 2000). Although NZC07 used an idealized upper-level thermal relaxation scheme to generate the tropopause, in this study we use the Goddard shortwave radiation scheme (Chou et al. 1998) with the diurnal cycle present and solar parameters fixed to a permanent equinox. Except where noted, the simulations also use the Grell–Dévényi ensemble cumulus parameterization scheme (Grell 1993; Grell and Dévényi 2002).

b. Domain and boundary conditions

For most of the simulations, the model domain is $200 \times 300$ points in the zonal and meridional directions, respectively, corresponding to a tropical channel 4174 km in length and 6261 km in width. The domain is centered at the equator, with the north and south walls corresponding to 28°S and 28°N, respectively.

With the model domain symmetric about the equator, four meridional profiles of SST are used. The first is a cross-equatorial equivalent to that used in NZC07, which had the SST maximum of 30°C decreasing linearly away from the equator at the same rate of $1/3$ °C per degree latitude. When extended across the equator, this makes a sharply peaked SST profile as shown in Fig. 2a. Hess et al. (1993) used a similarly peaked SST distribution in their evaluation of ITCZs simulated in a global aquaplanet model, and a similar profile has been used as one of the test cases for an intercomparison of aquaplanet simulations (Neale and Hoskins 2001a,b). It is used here to evaluate the extent to which the results in NZC07 hold for simulations that allow for cross-equatorial flow. The simulation with this SST profile is...
referred to as PKD; a list of all the simulations and their parameters are presented in Table 1.

Most of the simulations use smoother SST profiles. A convenient functional form that, with variation of one parameter, produces a smooth SST profile around the equator that still asymptotes to a linear profile at large distance is the hyperbola,

\[
\text{SST}(\phi) = \frac{\text{SST}_{\text{max}}}{1 + \frac{\phi - \phi_{\text{max}}}{d}},
\]

where \(\phi\) is the latitude, \(\phi_{\text{max}}\) is the latitude where the SST reaches its peak value \(\text{SST}_{\text{max}}\), and the distance \(d\) controls the shape of the hyperbola. The constant \(A\) is chosen so that, for \(\phi_{\text{max}} = 0\), the SST profile intercepts 20°C at \(\phi = \pm 30^\circ\), just like for PKD. As \(d \to 0\), (2.1) approaches the PKD profile. Examples with \(d = 2.5^\circ\) (HYP2.5) and \(d = 5^\circ\) (HYP5.0) are shown in Fig. 2a.

As an opposite extreme to PKD, an SST profile that varies as a cosine function with latitude (COS) was used. This SST profile is representative of regions with a broad, nearly flat SST profile around the equator, such as the western Pacific warm pool. Near the equator, this SST profile is an equivalent functional form to the \((1 - \sin^2\phi)\) “control” profile used by Neale and Hoskins (2001a), because \(1 - \sin^2\phi = \cos^2\phi \approx \cos 2\phi\) for small \(\phi\).

We also developed an idealized cross-equatorial profile that is representative of the eastern Pacific, with cool temperatures south of the equator, an equatorial cold tongue, a very warm SST peak near 10°N, and SST then decreasing northward. The profile was made by first generating a piecewise linear function of latitude,

\[
\text{SST}(\phi_j < \phi \leq \phi_{j+1}) = (T_j - T_{j-1}) \frac{(\phi - \phi_j)}{(\phi_{j+1} - \phi_j)} + T_{j-1},
\]

using the parameters \(T_1, \ldots, T_6 = 20^\circ, 22^\circ, 19.5^\circ, 26.5^\circ, 29.0^\circ, 23.33^\circ\) and \(\phi_1, \ldots, \phi_6 = -30^\circ, -7.5^\circ, -2^\circ, 3^\circ, 15^\circ, 32^\circ\) latitude. This profile, defined on a regularly spaced grid with \(\Delta y = 20.87\) km, was then smoothed 20 times with a 1–2–1 filter. The resulting “idealized eastern Pacific” profile (IEP) is shown in Fig. 2b. The parameters and smoothing were engineered so that the profile matched closely to the SST profile at longitude 95°W during the EPIC field campaign as shown by, for example, McGauley et al. (2004, their Fig. 3) and Pyatt et al. (2005, their Fig. 3). To better handle the Hadley circulation on the north side of the ITCZ, the domain size for these simulations was increased to 320 grid points in the meridional direction, with the southern boundary remaining at 28°S and the northern boundary extending to 32.0°N.

3. Cross-equatorial simulations with symmetric SST profiles

a. Comparisons to prior results

For every simulation, the model is first integrated for a period of 90 days to allow moist convection and the overturning circulations to reach equilibrium. After 90 days, model output is saved every 3 h of simulation time. Most of the temporally and zonally averaged results (referred to as time-zonal means) are averaged over all grid points every 3 h for 60 days; see Table 1 for details.

We first consider whether, in the updated model configuration, the idealized ITCZ and its multilevel flows,
defined by the BLI, SRF, MLI, and ULO are similar to those produced in the half-channel simulations of NZC07. Figure 3 shows the time-zonal mean circulations for simulation PKD. Figures 3a,b show the zonal wind $u$ and potential temperature $\theta$ across the entire meridional extent of the domain. The idealized ITCZ and Hadley circulations generate a subtropical jet that is balanced by a large meridional temperature gradient in the middle troposphere. The jet is unstable and supports repeated events of baroclinic instability and mid-latitude cyclones, but these do not appear in the mean. The multilevel flows are clearly present, although they do have some structural differences from NZC07 (Fig. 1a). The SRF is at a slightly lower altitude and is slightly thinner in the vertical. The MLI is at a slightly higher altitude and is also thinner. Although the SRF and MLI are slightly weaker than in NZC07, so are the BLI and ULO, indicating that the entire overturning becomes slightly weaker when cross-equatorial dynamics are permitted. An artifact of the half-channel domain of NZC07 is that the peak vertical velocities occur only within one or two grid points of the equatorial boundary and are 3 times larger than in PKD. This localization of convection to such a small region may enhance the precipitation efficiency, leading to a stronger overturning circulation for the same SST profile.

Simulation HYP2.5 generated very similar results (Fig. 4). The full-domain means for $u$ and $\theta$ are nearly identical to PKD and are not shown. The ITCZ is about twice as wide and mean vertical velocities are about half as strong. The peak strengths of the primary meridional flows, the BLI and the ULO, are about 10% weaker, whereas the SRF and the MLI are weaker by about a third.

It is evident for both PKD and HYP2.5 that the time-zonal mean fields are not perfectly symmetric about the equator. These asymmetries are most pronounced for the MLI for HYP2.5, as shown in Fig. 4a, where the MLI on the north side is 25% stronger than on the south side. Some asymmetries are inevitable when averaging over a finite time period, even as long as 60 days. However, they are also caused by another process, which is a zonally symmetric oscillation of the meridional flow across the equator. This will be shown in more detail at the end of section 3c.

A plan view of an instantaneous precipitation field for HYP2.5 is shown in Fig. 5a. The plot shows the base 10 logarithm of the total (resolved and parameterized) rain rate in units of millimeters per hour, after adding 1 mm h$^{-1}$ to the rain rate everywhere to make the minimum value of the logarithm equal to 0. The peak logarithmic values around 1.3 indicate rain rates around 20 mm h$^{-1}$. The simulated ITCZ lies very close along the equator. The time of this snapshot was chosen because it shows other typical features of the simulated ITCZ for HYP2.5: first, that the ITCZ meanders a few degrees north and south of the equator typically with a wavelength equal to the domain length, and second, that the ITCZ shows signs of splitting into two bands of precipitation in some places. This splitting does not occur for the peaked SST profile (not shown); for an only slightly broader SST profile, the ITCZ splits completely. This will be discussed further in the next section.

NZC07 claimed that the strength of the SRF is modulated by the activity of deep convection in the nearby ITCZ. This was shown to be true in the simulations by computing composite fields for a limited region (one quarter of the zonal length of the domain) for times

<table>
<thead>
<tr>
<th>Simulation name</th>
<th>SST profile type</th>
<th>Lat of SST max</th>
<th>Width parameter</th>
<th>Modifications to physics</th>
<th>Length of dataset for analysis (days)</th>
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<tr>
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<td>Linear</td>
<td>Equator</td>
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<td>—</td>
<td>60</td>
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<tr>
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<td>Hyperbolic</td>
<td>Equator</td>
<td>$d = 2.5^\circ$</td>
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<td>Hyperbolic</td>
<td>Equator</td>
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<td>Latent heat of freezing 1% of normal value</td>
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<tr>
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<td>Warm-rain microphysics and no cumulus parameterization</td>
<td>15</td>
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<td>30</td>
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<td>No cumulus parameter</td>
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<tr>
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<td>8$^\circ$N</td>
<td>$d = 2.5^\circ$</td>
<td>—</td>
<td>60</td>
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<tr>
<td>COS</td>
<td>Cosine</td>
<td>Equator</td>
<td>—</td>
<td>—</td>
<td>30</td>
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<tr>
<td>COS-NC</td>
<td>Cosine</td>
<td>Equator</td>
<td>—</td>
<td>No cumulus parameter</td>
<td>30</td>
</tr>
<tr>
<td>IEP</td>
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<td>—</td>
<td>—</td>
<td>45</td>
</tr>
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<td>10$^\circ$N</td>
<td>—</td>
<td>Latent heat of freezing 1% of normal value</td>
<td>30</td>
</tr>
</tbody>
</table>
when the mean SRF was either one standard deviation above or below its mean value in that region. (As the convection was organized by the passing CCKW, compositing on a limited subdomain allowed the sampling of convectively active or inactive regions.) The strong SRF composite had substantially weaker heating and vertical motion at the equator, whereas the weak SRF composite had relatively intense convection at the equator.

Here, we repeat the composite analysis for HYP2.5. Instead of compositing on the strength of the SRF, we composite on the strength of the moist heating (defined as the sum of the diabatic heating tendencies from both the microphysical and cumulus parameterizations) along a 520-km-long segment (1/8th) of the domain, choosing times when the mean moist heating between 8.1 and 9.2 km (model levels 27–29) and within 200 km of the equator was one standard deviation above or below its...
As shown in Fig. 6b, the SRF is indeed weaker when the deep moist heating (convection) is stronger (Fig. 6a) and vice versa (Figs. 6c,d). Composites of the mean flow based on total (resolved and parameterized) rain rate, rather than upper-level moist heating, showed nearly identical changes in the SRF. NZC07 also found that the strength of the MLI followed the same pattern as the SRF, being substantially stronger when ITCZ convection is weaker. Such a relationship is not apparent in Fig. 6. For the strong convection composite, the MLI is slightly stronger and penetrates closer to the equator. For the weak convection composite, it is quite asymmetric and weaker, and it does not penetrate as close to the equator as for the full mean. This would be opposite to the pattern found by NZC07.

b. Broader SST profiles and double ITCZs

Time-zonal means for simulation HYP5.0 are shown in Figs. 7a,b. Despite what appears to be a very small difference between HYP2.5 and HYP5.0 (see Fig. 2a), HYP5.0 clearly generates a “double ITCZ” circulation (Zhang 2001). Note that the shading and contour values in Figs. 7a,b have been reduced by a factor of 2 for $v$ and by a factor of 4 for $w$, so the circulation is much weaker. The double ITCZ is also evident in a plan view snapshot of the total rain rate (Fig. 5b). Each of the north and south ITCZs periodically “break down” into weak tropical cyclones (TCs), as is illustrated in this case. More profound examples of ITCZ breakdown into TCs will be presented in the next section.

Unfortunately, the transition from a single to double ITCZ configuration for broader SST profiles is parameterization dependent. Figure 7c shows the mean $v$ field for a simulation with the HYP5.0 profile but with the cumulus parameterization deactivated such that only explicitly resolved convection occurs (HYP5.0-NC). There is a single ITCZ again, with robust multilevel circulations such as the SRF and the MLI (contours and shading values are returned to their previous values). Many studies, mostly using global models with very coarse resolution, have found similar sensitivities of double ITCZ development to the shape of SST profiles around the equator and choice (or lack) of cumulus parameterization (Hayashi and Sumi 1986; Swinbank et al. 1988; Lau et al. 1988; Hess et al. 1993; Neale and Hoskins 1993).
The present results are summarized in Fig. 7d. For HYP5.0, the double ITCZ is evident, whereas, for HYP5.0-NC, the rain rate profile collapses back to a nearly perfect match to the HYP2.5 rain rate. The rain rate for simulation COS also shows a double maximum with cumulus parameterization and a single maximum without (COS-NC).

Worth noting, however, are the changes in the SRF and MLI that occur for the broader SST profile (HYP5.0) with and without cumulus parameterization. The SRF is very weak in HYP5.0. Although it is more robust in HYP5.0-NC, it is still weaker than for HYP2.5, perhaps because of the weaker SST gradient. The MLI is robust in both simulations. Although the MLI is weaker in HYP5.0 in the absolute sense, it is approximately the same strength relative to the ULO as it is for HYP2.5, whereas the SRF is substantially (relatively) weaker in HYP5.0. This suggests the MLI is not dynamically related to the SRF but instead is closely related to the intensity and distribution of the deep convection in the ITCZ.

c. Tropical waves and large-scale oscillations for symmetric SSTs

Convectively coupled tropical waves are active in these simulations. A complete analysis of these waves is beyond the scope of the present study, so we shall only document the dominant features. We first return to simulation PKD. Time–longitude Hovmöller diagrams are shown in Figs. 8a,b for \( u \) and \( v \) anomalies (i.e., deviations from their mean values over the 60-day interval) on model level 7, near 850 hPa, over the region from \( 2.5^\circ \text{S} \) to \( 2.5^\circ \text{N} \). A clear, eastward-propagating signal of anomalous zonal winds can be seen with zonal wavenumber \( k = 1 \) (note that here and below this wavenumber refers to a single wave within the 4174-km domain, not a planetary wavenumber). The period of the wave is about 2.1 days, corresponding to an eastward propagation speed of 23 m s\(^{-1}\), or 27 m s\(^{-1}\) relative to the mean flow of \( \pm 3.8 \text{ m s}^{-1} \) from the surface to the tropopause and within 2.5° of the equator. The Earth-relative 23 m s\(^{-1}\) speed is a bit faster than the 20 m s\(^{-1}\)
speed of the Kelvin waves in the global idealized simulations of Frierson (2007); it is also consistent with the 20–25 m s\(^{-1}\) speeds predicted by a theoretical model of convectively coupled Kelvin waves presented by Dias and Pauluis (2009), for ITCZ widths of 200–400 km. The comparison cannot be more precise, because their model defines the ITCZ as a discrete, moist region of specified width, whereas the width of the ITCZ cannot be so precisely defined here.

Westward-propagating disturbances can be seen in the Hovmöller diagram of anomalous \(v\) (Fig. 8b), with short zonal scales and speeds ranging from 2 to 4 m s\(^{-1}\), suggesting their motion is due to westward advection by the mean flow. These anomalies in \(v\) represent the anomalous wind fields associated with long-lived, convective clusters that are not centered on the equator.

Figure 8d shows a Hovmöller plot of the anomalous value of the base 10 logarithm of the rain rate between 2.5\(^\circ\)S and 2.5\(^\circ\)N. The \(k = 1\) Kelvin wave signal is also evident. Indication of a different wave type, however, can be seen in the Hovmöller diagram of \(v\) at 10 m in Fig. 8c. A large-scale, \(k = 1\) feature is evident, propagating westward. Its phase speed is quite slow, 3.2 m s\(^{-1}\) to the west, indicating that it may be stationary or even moving eastward relative to the mean flow.

Hovmöller diagrams of the same fields for HYP2.5 are shown in Fig. 9. Remarkably, changing from the PKD profile to the only slightly smoother HYP2.5 profile causes a noticeable weakening of the Kelvin wave signal: it is much less visible in either Fig. 9a or Fig. 9d. The phase speed is nearly identical to PKD. The \(k = 1\),
westward-propagating feature in the 10-m $v$ fields (Fig. 9c) is much more prominent for HYP2.5.

Wavenumber–frequency spectra of the anomalous logarithm of the rain rates for PKD and HYP2.5 are shown in Fig. 10. The spectra were made from 60 days of three-hourly model output: that is, 480 points in time. The ends of the time series were windowed to zero with Gaussian functions with an exponential decay width of 40 points. Finally, both spectra were smoothed in frequency–wavenumber space with a 1–2–1 filter just twice in each direction. Note that the base 10 logarithm of the full spectra are shown, not the deviation of the spectra from a background “red” spectrum as presented in Wheeler and Kiladis (1999). Overlaid on the plots are solid lines indicating phase speeds of 23 m s$^{-1}$ to the east for both diagrams, and 3.8 and 4.3 m s$^{-1}$ to the west for PKD and HYP2.5, respectively; these are the mean easterly flow speeds between the surface and the tropopause from 2.5°S to 2.5°N.

Both spectra show a highly localized maximum near frequency $v = 1$ cycle per day (cpd) and $k = 0$. This represents the diurnal cycle. PKD shows a distinct maximum at $k = 1$, $v = 0.55$ cpd, with an additional peak along the same Kelvin wave speed for $k = 2$, corresponding to the CCKW. HYP2.5 has similar peaks at $k = 1$ and 2. The figures show the base 10 logarithm of the spectral power; thus, the peaks for the diurnal cycle and the Kelvin wave, having a log$_{10}$ difference of 1.0 above the surrounding values, are 10 times stronger than the surrounding noise. For as little as 30 or even 15 degrees of freedom represented by the 60-day dataset for signals at 1 or 0.5 cpd, these peaks are statistically significant.

The PKD and HYP2.5 results also show broad spectra of westward-propagating disturbances that are moving at or more slowly than the mean flow speed, with maximum power at the lowest wavenumbers as suggested by the westward-propagating feature in the 10-m $v$.
FIG. 8. Hovmöller diagrams for anomalies from mean values between 2.5°S and 2.5°N of various fields for simulation PKD: (a) $u$ at model level 7 (mean pressure 855 hPa); (b) $v$ at model level 7; (c) surface (10 m) $v$; and (d) logarithmic rain rate.
FIG. 9. As in Fig. 8, but for HYP2.5.
FIG. 10. Wavenumber–frequency diagrams of spectral power of the anomalous logarithmic rain rate for (a) simulation PKD and (b) HYP2.5. The dark lines in each plot correspond to 23 m s$^{-1}$ (to the east) and the mean zonal wind speed in the averaging region (to the west), which are $-3.8$ and $-4.1$ m s$^{-1}$, respectively. The contour interval is 0.2 in both panels.
for both cases. For HYP2.5, there are distinct peaks at $k = 1$ and $k = 2$ along the mean flow speed. The meridional flow across the equator indicates this oscillation is, or is similar to, a mixed Rossby–gravity wave (Matsuno 1966; Wheeler and Kiladis 1999). Although the $n = 2$ westward inertia–gravity wave also has cross-equatorial flow, its much faster propagation speed would not be compatible with this wave. For the higher wavenumber features, representative of the individual convective clusters, the westward speeds are less than the mean flow, more so for PKD.

In section 3a it was noted that asymmetries across the equator of some of the time-zonal mean fields, such as $v$ in Fig. 4, were caused by a zonally symmetric oscillation of the flow. This oscillation is best illustrated with a time–height Hovmöller diagram of mean $v$ along the length of the domain and within 100 km of the equator (Fig. 11). Pronounced oscillations in $v$ are evident at $z = 9$ and 16 km, which appear to be the peaks of growing, vertically propagating waves. As of yet, we have no understanding of the physics of this oscillation, but it is very likely an artifact of some aspect of the model configuration. For both PKD and HYP2.5, it disappears completely with cumulus parameterization turned off. Nor does it appear in any of the equatorially asymmetric simulations presented below.

4. Off-equatorial ITCZs

a. Hyperbolic SST profile with SST maximum off the equator

As the simplest example of an off-equatorial ITCZ, we shift the maximum of the HYP2.5 profile to 8°N (HYP2.5–8N). The time-zonal mean fields for this simulation are shown in Fig. 12. The ITCZ has moved to 4°N. Significant convection and upward motion are also spread several degrees poleward of their maxima; note that the contour intervals and shading ranges for $w$ and moist heating are half as large as for the symmetric cases shown in Figs. 3 and 4.

Consistent with observations, reanalyses, and theoretical ITCZ dynamics (e.g., Lindzen and Hou 1988), the meridional circulations become highly asymmetric about the equator, with significant cross-equatorial flows on the equatorward side of the simulated ITCZ and very weak meridional flow on the poleward side, except for the ULO. In addition, new inflows and outflows have appeared in between the SRF and MLI. These flows seem to feed into and emanate from a local maximum in moist heating in the ITCZ near $z = 4.5$ km. Indications of these additional multilevel flows are present in some of the previous figures for the equatorially symmetric simulations, as in Figs. 6b,d.

A plan view of the logarithmic rain rate for this simulation is shown in Fig. 5c. At the time shown, a large TC is present and a 1500-km-long stretch of the convecting region is drawn away from the ITCZ latitude and wraps into the cyclone. Animations of this simulation show the ITCZ frequently breaking down into a TC that then migrates poleward until it separates from the ITCZ itself, which then reforms, intensifies, and breaks down again. In their survey of synoptic variability of the Pacific ITCZ, Wang and Magnusdottir (2006) found ITCZ breakdown to be a frequent occurrence during the hurricane season in the eastern and central Pacific. The primary mechanism for breakdown is the instability of the zonal flow across the ITCZ, leading to the so-called vortex-rollup process (Nieto Ferreira and Schubert 1997; Wang and Magnusdottir 2005). In idealized simulations of the ITCZ using the quasi-equilibrium model of Neelin and Zeng (2000), Wang et al. (2010) generated similar structures as shown in Fig. 5c. They referred to the long band of convection west of the TC as the “tail” and showed examples of the tail in observations.

To evaluate the effect of these evolving and migrating TCs on the ITCZ and its circulations, time-zonal mean fields for this simulation were computed again, but only for zonal locations that had a TC in some proximity north of the ITCZ. The proximity of a TC was determined based on the minimum value of the sea level pressure (SLP) north of the ITCZ along each zonal grid point; if SLP was above 1010 hPa somewhere between the equator and 19°N, then the data from that zonal location was removed from the computation of the time-zonal mean; the TC-influenced points comprised 15% of the total dataset. The results of this calculation are
shown Figs. 12e,f. The time-mean $\nu$ shows that the regions influenced by a TC have stronger BLI and MLI reaching further poleward to a strong heating maximum around 9ºN. The complementary time-zonal mean $\nu$ for the non-TC-influenced points (not shown) was nearly identical to the time-zonal mean for all points (Fig. 12a), whereas the time-zonal moist heating was nearly identical to the total time-mean shown in Fig. 12d, with the secondary heating maxima from 7º to 10ºN completely absent.

This analysis suggests that, for an ITCZ relatively close to the equator, the periodic breakdown of the ITCZ into TCs results in regions of enhanced cross-equatorial flow and a secondary maximum in heating farther poleward. However, the strong similarity between the non-TC-influenced circulation (not shown) and the complete time-zonal mean indicates that the TCs do not significantly enhance the mean cross-equatorial circulations.
b. Idealized eastern Pacific SST profile

As described in section 2b, the IEP profile was developed with similar features to the mean SST profile in the eastern Pacific during the boreal summer and fall (see Fig. 2b). Time-zonal means for simulation IEP are shown in Fig. 13. The results are mostly similar to HYP2.5–8N, but there are some differences. In contrast with HYP2.5–8N, where the ITCZ only reaches to 4°N, here the ITCZ is a few degrees poleward of the SST maximum at 10°N. However, in HYP2.5–8N, the profile of SST drops sharply with latitude on both sides of the maximum. In development of the IEP profile, it was found that the location of the ITCZ was quite sensitive to the SST gradient north of the SST maximum. In preliminary tests using SST profiles that decreased even more slowly north of the maximum (which in fact would be more realistic for late summer eastern Pacific SST profiles), the simulated ITCZs settled even farther poleward.

Time-zonal means for regions influenced and not influenced by TCs were computed for IEP in the same manner as above. Remarkably, both exhibited very little difference from the full mean in Fig. 13d. This may be understood from viewing the precipitation at one time for the IEP case (Fig. 5d). This time is representative of most times of the 2-month-long dataset in that there are multiple TC circulations lying along the ITCZ. In fact, the ITCZ for this case usually consists of series of “linked” TC-like circulations.

All of the previous simulations showed SRFs that were quite low in altitude, 1–3 km above the surface. However, in Fig. 13a the low-level SRF appears to have been replaced by a much stronger, deeper, and more elevated SRF ranging from 3 to 5 km in altitude. That the SRF in this case emanates out of the ITCZ from above the low-level maxima in moist heating and $w$ suggests that it is in fact more like the shallow circulations proposed by Wu (2003), which are essentially detrainment from shallow (though precipitating) convection. However, it may still be at least in part due to the sea-breeze type of response to the very large temperature gradient in this case, with SST decreasing from 28° to 19.5°C over a distance of only 1200 km (Fig. 2b). If these SST values and the simulated surface pressure at their respective locations (1012 and 1015 hPa) are applied to the simple hydrostatic model used in NZC07 (see section 2 of that paper), the depth of the diagnosed SRF
increases substantially with the top of the predicted SRF layer reaching to over 8 km in height.

Another significant change in the SRF for this case involves its apparent degree of moisture transport. In the prior simulations of NZC07 and in the simulations shown above, the SRF is relatively short in meridional distance and moves across locally strong gradients in mean relative humidity (RH) while not appearing to advect RH significantly. However, for IEP, the SRF is clearly linked to a long, meridional deformation in the RH contours, indicating significant transport of humidity. These differences will be quantified in section 6.

c. Convectively coupled waves in the off-equatorial ITCZs

For both HYP2.5–8N and IEP, power spectra were computed for the anomalous logarithmic rain rates in 5° bands around the equator and the mean latitude of each ITCZ; these are shown in Fig. 14. Since there is far less precipitation along the equator for these simulations, the total spectral power in anomalous precipitation is much less. However, for HYP2.5–8N, clear signals in convection can still be seen moving westward with the mean flow and eastward with CCKWs (Fig. 14a). Along the ITCZ (Fig. 14b), the power spectrum is similar to HYP2.5 (Fig. 10b), except that the mean flow to the west is slower and there is considerably more power in slow, eastward motions. Some of this slow eastward propagation is associated with TCs, which meander both eastward and westward in the doldrums between the ITCZ and the strong baroclinic westerly flow to the north. Nonetheless, there is still a clear signal along the Kelvin wave eastward phase speed.

For IEP, the power in anomalous rain rate along the equator is nearly gone (Fig. 14c), though the diurnal cycle is still evident. Because of the large eastward deflection of the cross-equatorial BLI, the mean zonal flow at the ITCZ is actually eastward, and almost all the spectral power in the ITCZ convection (Fig. 14d) lies along and above (meaning faster than) the mean tropospheric flow speed. This is because the tropical cyclones propagate even faster toward the east in this case.
5. Causes and implications of the multilevel flows

a. Understanding the midlevel inflow

Because of the absence of strong horizontal temperature gradients in the tropics, scaling analysis shows that significant vertical motions must be closely correlated with diabatic heating (Holton 2004). Above the boundary layer, divergent horizontal motions are connected by mass continuity to vertical gradients in vertical motion, which must be correlated with vertical gradients in diabatic heating.

To illustrate the relationship between diabatic heating, vertical velocity, and the MLI in HYP2.5, we show in Fig. 15a vertical profiles of moist heating, radiative heating, eddy heating, and \( w \) within \( 2.5^\circ \) of the equator and rescaled to fit on the plot; \( w \) was multiplied by 100 and each heating term was multiplied by \( 2 \times 10^3 \); (b) dry lapse rates inside the ITCZ (2.5°S–2.5°N) and outside the ITCZ (5°–10°) on either side; (c) as in (a), but for the IEP case, using a profile of \( v \) at 5°N with the sign reversed to be consistent with (a), and \( w \) and heating profiles from 10° to 12°N; (d) as in (b), but for IEP, with the lapse rate inside the ITCZ from 10° to 12°N and outside the ITCZ from 5°S to the equator.

![Fig. 15. Vertical profiles of various fields in and around the ITCZ for simulations HYP2.5 and IEP: (a) vertical profile of the symmetric part of \( v \) at 4° from the equator, along with \( w \), moist, radiative, and eddy heating, each averaged within 2.5° of the equator and rescaled to fit on the plot; \( w \) was multiplied by 100 and each heating term was multiplied by \( 2 \times 10^3 \); (b) dry lapse rates inside the ITCZ (2.5°S–2.5°N) and outside the ITCZ (5°–10°) on either side; (c) as in (a), but for the IEP case, using a profile of \( v \) at 5°N with the sign reversed to be consistent with (a), and \( w \) and heating profiles from 10° to 12°N; (d) as in (b), but for IEP, with the lapse rate inside the ITCZ from 10° to 12°N and outside the ITCZ from 5°S to the equator.](image)

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To illustrate the relationship between diabatic heating, vertical velocity, and the MLI in HYP2.5, we show in Fig. 15a vertical profiles of moist heating, radiative heating, eddy heating, and \( w \) within \( 2.5^\circ \) of the equator and the symmetric part of \( v \); that is, \( v_{\text{sym}} = (v^+ - v^-)/2 \), where \( v^+ \) and \( v^- \) are the meridional wind profiles at latitudes 4°N and 4°S, respectively (Fig. 15a). The eddy heating is the tendency due to the divergence of vertical eddy fluxes,

\[
\frac{\partial \theta}{\partial t} \sim -\frac{1}{|p|} \frac{\partial}{\partial z} \left[ \rho w \theta^* \right],
\]

where, following the notation of Peixoto and Oort (1992), bars and square brackets refer to time and zonal means and primes and stars refer to deviations from those means, respectively. Although a significant contribution from this eddy flux tendency is often seen in cloud-resolving simulations (e.g., Sui et al. 1994), we find this heating to be small in simulations with cumulus parameterization.

As it appears in Fig. 15a, the eddy heating is fairly well anticorrelated with, though slightly weaker in amplitude than, the radiative heating; this anticorrelation is evident...
at $z = 2.0$, 8.0, 10.5, and 14 km in Fig. 15a. This anticorrelation is caused by the destabilization of the atmosphere by radiative cooling at the tops of persistent cloud decks, leading to convective overturnings that enhance the vertical eddy flux. Similarly, the capture of upwelling longwave radiation on the lower sides of clouds stabilizes the atmosphere below it and causes eddy motions to drive a reverse temperature flux. A similar anticorrelation is evident in Fig. 15c for the IEP case.

More to the point regarding the MLI in Fig. 15a, there are large peaks in moist heating and $w$ around $z = 9$ km and a large negative peak in $v_{\text{sym}}$ at $z = 8$ km. Although the profiles of $v_{\text{sym}}$ and $w$ are quite smooth, moist heating shows large jumps on the scale of the vertical discretization.

Considering that the real ITCZ rarely sits on the equator, the same profiles are shown for the more realistic IEP simulation (Fig. 15c). The upper-level heating maximum is also present (though not as sharp) with the same relationship to the MLI. Thus, in both HYP2.5 and IEP, a 50%–100% increase in the total heating between $z = 7.5$- and 8.5-km altitude appears to be responsible for the upper-level peak in $w$ and the forcing of the MLI. This cannot be the entire story, however, because large peaks in heating at $z = 3$ and 4.5 km in HYP2.5 and IEP, respectively, are not matched with similar variations in $w$ and $v$. For IEP, this peak at $z = 4.5$ km corresponds to the freezing level. Such a peak appears in our simulations wherever the heating is dominated by “resolved” convection; this occurs when the convection is strongly forced by boundary layer convergence, as is the case for IEP, or for simulations without cumulus parameterization (not shown).

Large-scale vertical motion forced by diabatic heating $Q$ is modulated by stability, which may be written in terms the lapse rate $\Gamma$: that is,

$$w \approx \frac{Q/c_p}{g(c_p + \Gamma)} \quad (5.2)$$

where $Q$ is a diabatic heat source, $c_p$ is the heat capacity of dry air at constant pressure, and $g$ is the gravitational constant. As shown in Figs. 15b,d, there are large vertical variations in $\Gamma$ both inside and outside the ITCZ. Above the boundary layer, $\Gamma$ varies around $-4 \text{ K km}^{-1}$ up to about 3 km, above which it decreases and then increases again. This variation in $\Gamma$ above and below 5 km is caused by melting of ice below the freezing level and is frequently observed (e.g., Johnson et al. 1999; Folkins et al. 2008; note our simulated lapse rates around the freezing level are quite consistent with these). However, within both the HYP2.5 and IEP ITCZs the simulated lapse rates show another large oscillation around $z = 9$ km. The local minima of $\Gamma = -6.6$ and $-6.9 \text{ K km}^{-1}$ near $z = 8$ km for HYP2.5 and IEP, respectively, correspond to large increases in both moist heating and $w$ and thus further enhance the MLI.

Unfortunately, this feature is not seen in any observational composites that we have been able to find (see, e.g., Mapes 2001, Fig. 7; Folkins et al. 2008, Fig. 12), so it too may be a model artifact. The $\Gamma$ profile outside of the ITCZ, also shown in Fig. 15b, shows a much smoother decrease in $\Gamma$ with height between 7 and 9 km, although there is some signal of the upper-level variation near 10 km. Despite the nearly complete absence of convection in this region, there is a robust signal of the melting level in the $\Gamma$ profile outside of the ITCZ. This is a robust illustration of an effect observed and discussed by Johnson et al. (1999) that the stable layer in convecting regions at and above the freezing level is communicated outward to the surrounding atmosphere.

The enhancement of midlevel inflow by decreased stability below the upper-level heating maximum height was identified by Mapes (2001) and is also evident in calculations of the atmospheric response to observed mean heating profiles in tropical convection by Zhang and Hagos (2009). When their observed mean heating is applied to a linear model to diagnose the atmospheric response, a midlevel inflow appears that is more pronounced than what would be expected solely from the local vertical gradient of the heating.

With these previous and the current results in mind, it would be tempting to simply state that the upper-level heating maximum associated with deep convection is the “cause” of the midlevel inflow. However, such a statement would be incomplete. Vertical profiles of moist heating, $w$, $\Gamma$, and $v$ are all highly interrelated, and it would be incorrect to state that any one of these fields completely controls the others. Nonetheless, it seems likely that the “first” cause of the MLI is the maximum in upper-level heating and that the circulations it generates provide some feedback, particularly through modification of the environmental sounding below the heating maximum by forced ascent.

With this in mind, we considered changes to the model that might eliminate the upper-level heating maximum. A partially successful change was to simply change the latent heat of freezing $L_f$ to 1% of its actual value. As shown in Fig. 16, this produces a tropical circulation quite similar to the unmodified HYP2.5. The MLI has not been eliminated but has been reduced in strength by about half. Accordingly, there is still a small upper-level maximum in moist heating. Nearly identical results were obtained for an IEP simulation with $L_f$ changed 1% of its value (not shown).
In further attempts to eliminate or reduce the MLI, we considered other changes to the model. One was to eliminate the cumulus parameterization and change the microphysics to the Kessler (1969) “warm rain” scheme. Unfortunately, this produced a radically different time-zonal mean flow. Without ice processes, the atmosphere becomes very inefficient at removing water from the atmosphere so that the equilibrated simulation has large amounts of water vapor and condensate throughout the upper troposphere and a very weak ITCZ rising to only 10 km height (not shown). We also considered that the pronounced upper-level structures in moist heating and lapse rate could be related to radiation feedbacks. To eliminate such feedbacks, we modified the radiation parameterizations so that the combined vertical profile of heating and cooling from radiation was set to be an idealized profile with constant heating above 200 hPa, constant cooling below 400 hPa, and a linear transition in between. The total tropospheric heating from the idealized profile at each point in time and space was equal to the total heating from the direct output of the interactive longwave and shortwave schemes. Simulations with these “noninteractive” radiative heating profiles had similar or even stronger MLIs (not shown).

What can be gleaned from these results? Simply that the MLI is closely related to the widely observed and simulated strong upper-level heating maximum associated with ice processes in deep convection. The MLI is very robust: it is difficult to modify the atmosphere in such a way as to eliminate it without drastically changing the mean circulation in unrealistic ways.

b. Implications for budgets of mass, moisture, and moist static energy

If the SRF and MLI are robust features of the ITCZ, what impact do they have on budgets of mass, water, and MSE? The issue of water transport was first addressed in NZC07: they found that about 10% of the water transported into the ITCZ region in the moist BLI, and from surface fluxes, was transported out of the ITCZ in the SRF.

We consider the same question, extending our analyses to transports of total mass; total water; and MSE $h$,

$$h = c_p T + gz + L_v q_{vapor},$$

where $c_p$ is the heat capacity of dry air at constant pressure, $T$ is the temperature, $L_v$ is the latent heat of vaporization, and $q_{vapor}$ is the mixing ratio of water vapor to air. (Results of calculations using the “frozen moist static energy,” which accounts for changes in $h$ due to ice condensate, were nearly identical.) For each simulation, a domain representing the ITCZ region was subjectively selected. For example, the ITCZ region for HYP2.5 ranges from 4°S to 4°N; in each case, these boundaries were chosen to collectively maximize the multilevel fluxes. Fluxes of mass, water, and $h$ in and out of the ITCZ were computed for each of the BLI, SRF, MLI, and ULO flows on both the north and south boundaries. Dividing levels between the flows were determined by examining the altitudes of the flow reversals for $v$ along each vertical boundary. For the off-equatorial simulations, the locations and depths of the circulations were different on each side of the ITCZ. In some cases there were additional reversals between the SRF and MLI, so the dividing line between them was chosen arbitrarily. The fluxes are computed by vertically integrating the mean values of the actual fluxes,

$$F = \sum_{j=L_2}^{L_1} (m_j v_j \Delta p_j)/g.$$
where \( L_1 \) and \( L_2 \) refer to the bottom and top model levels contained by the flow and \( m \) varies according to the substance in question. For water, \( m = q_{\text{vapor}} + q_{\text{condensate}} \). For MSE, \( m = h \). For mass, \( m = 1 \) (the mass contribution of water is neglected). The salient point is that these fluxes are computed from the mean values of the products, including the varying values of mass between the model levels \( \Delta p/g \), rather than the products of the means.

The budgets for HYP2.5 are summarized in Fig. 17. Looking first at the mass budget, there is one surprise: the mass flux contribution of the MLI to the ITCZ circulation is quite large, being nearly half as large as the mass flux of the BLI. In other words, in addition to the “in–up–out” circulation universally associated with the forcing of large-scale motions by convection in the tropics, about \( \frac{2}{3} \) of the mass transported out of the ITCZ in the ULO comes from the convergence of mid-level air.

The water budget is quite similar to that found by NZC07, with the budget dominated by BLI, surface fluxes, and precipitation. The SRF transports about 10% of the incoming water out of the ITCZ, and the upper-level circulations make very small contributions.

The MSE budget also shows significant contributions from the MLI and SRF transports, with their MSE fluxes being 31% and 9%, respectively, of the outward flux of MSE by the ULO. The combined positive contribution of surface fluxes of heat and moisture, as well as the negative contribution of longwave and shortwave radiation, is quite small compared to the other terms, but this is in part because the ITCZ region has been defined to be quite narrow.

As the ITCZ is rarely symmetric about the equator, HYP2.5–8N and IEP are better analogs for the real atmosphere. Budget diagrams for these two cases are shown in Figs. 18 and 19. Accounting for the asymmetries in the flow on either side of the ITCZ regions, the results noted above still apply. About 30% of the air transported out of the region in the ULO originates in the MLI; the SRF water flux is about 10% of the total budget; and the SRF and MLI fluxes make significant contributions to the MSE budget. For IEP, it was noted above that there appears to be an even larger transport

![Fig. 17. Budget diagrams for dry mass, total water, and MSE for HYP2.5, separated into BLI, SRF, MLI, and ULO on either side of the equator, and relevant fluxes for each case. For mass, units are \( 10^3 \) kg s\(^{-1}\) m\(^{-1}\); for water, units are kg s\(^{-1}\) m\(^{-1}\); for MSE, units are \( 10^9 \) W m\(^{-2}\). The budget residual is shown in each case. For water, the fluxes are evaporation (up) and precipitation (down); for MSE, they are combined fluxes of heat and moisture from the surface and total radiation out the top (this includes incoming solar radiation). Also shown in the lower right are the vertically integrated total and eddy fluxes of MSE, with units \( 10^7 \) W m\(^{-1}\).]
of water vapor out of the ITCZ region by the cross-equatorial SRF (see Fig. 13c). This observation is confirmed by the budget diagram, which shows the cross-equatorial MSE flux by the SRF to be about \(1/3\) of the BLI and ULO values, where it is only about \(1/10\) of those values for two previous cases.

Several recent studies have proposed that eddy transports of water and MSE in and out of the ITCZ may be significant. Back and Bretherton (2006) found that a simplified calculation of MSE export in the eastern Pacific ITCZ, using the products of mean values and neglecting eddy transports, resulted in a large net transport of MSE into the ITCZ region. Peters et al. (2008) computed horizontal eddy transports for column-integrated MSE using 22 yr of data from the ERA-40. They found that outward (divergent) horizontal eddy transports of MSE balanced net inward transports of MSE by the mean overturning circulations. In their construction of a simplified, axisymmetric ITCZ model using two vertical modes of horizontal motion, Sobel and Neelin (2006) found that a large horizontal diffusion was required to produce realistic solutions.

Are there large eddy transports of MSE in these simulations? We have computed the mean total and eddy MSE fluxes, defined here as

\[
F_{h\text{ total}} = [\rho \overline{\overline{u}} h],
\]

where \(\rho\) is the total density and the overbar and square brackets refer to the time and zonal means, respectively, and

\[
F_{h\text{ eddy}} = [\rho \overline{\overline{u}} h] = [\rho \overline{\overline{t}} h] + [(\rho u)^{\overline{\overline{t}}} h] + [(\rho v)^{\overline{\overline{t}}} h^{\overline{\overline{t}}}].
\]

The middle expression in (5.6) has been used to compute \(F_{h\text{ eddy}}\), so the relative contributions of the three eddy terms have not been evaluated. The first eddy term should be very small because it consists of zonal variations of the time means, which should be nearly zero because of the zonally symmetric configuration. The second eddy term consists of zonally symmetric perturbations, which are also small, except perhaps because of the zonally symmetric oscillation for HYP2.5 as discussed in section 3c.

The values of \(F_{h\text{ total}}\) and \(F_{h\text{ eddy}}\) are shown for HYP2.5–8N in Fig. 20. The relatively important contributions of the SRF and MLI to the MSE budget are evident. A first look at the figures indicates the eddy MSE fluxes play only a tiny role in the overall budget. The peak total flux values of about \(2.9 \times 10^6\) W m\(^{-2}\) are over two orders of magnitude larger than the largest eddy flux values around \(1.2 \times 10^5\) W m\(^{-2}\).
However, in theories of the role of MSE in controlling precipitation and in the papers cited above, it is the vertically integrated transports that are considered. At the equator for this case, the vertically integrated MSE flux (southward) out of the ITCZ region is only 4.32 ± 10^7 W m^-2, whereas the northward flux across 13°N is 6.29 ± 10^7 W m^-2. The vertically integrated eddy fluxes are 0.76 ± 10^7 and 3.53 ± 10^7 W m^-2, respectively. Thus, although the contributions of the symmetric overturnings are large in each layer, their total contributions nearly cancel such that on the poleward side the total MSE export is actually larger than on the equatorward side. Furthermore, over half of the poleward MSE export is due to eddy fluxes.

The eddy flux of total water, which has only a very small contribution from condensate, is also shown in Fig. 20. The structure of this field is nearly identical to the eddy MSE flux; given that horizontal temperature gradients in the ITCZ region are small (see Fig. 3b), it is certain that the eddy transport of MSE is almost entirely due to eddy transport of water vapor. Furthermore, most of this transport occurs in the lower troposphere. Vertically integrated total and eddy export of MSE were also computed for HYP2.5 and IEP, and they are illustrated for all three cases in the flux diagrams in Figs. 17–19. For all three cases, the eddy MSE transport is 1/3–2/3 of the total transport on the poleward side of the ITCZ. For HYP2.5–8N and IEP, the eddy fluxes are mostly associated with the TCs; for HYP2.5, the eddy fluxes are maximized only a few degrees on either side of the equator and are associated with the meridional circulations generated by the CCKWs. Although pure Kelvin waves have zero meridional flow, CCKWs in the atmosphere (Wheeler et al. 2000), in idealized simulations (NZC07; Frierson 2007), and in the theory of Dias and Pauluis (2009) are found to have significant meridional flow into and out of their convecting regions.

Thus, our simulations of both highly idealized and more realistic (but idealized) ITCZs support the proposition that eddy transports play an important role in the MSE budget of the ITCZ region. Three additional facts should be noted: 1) almost all of the eddy transport is due to eddy transport of water vapor; 2) most of the eddy transport is in the lower troposphere; and 3) for the off-equatorial ITCZs the eddy transports on the poleward side provide the majority of the total MSE transport.

6. Summary and conclusions

We have used a mesoscale model in a tropical channel configuration to investigate the vertical and horizontal structure of the ITCZ and its associated mean meridional...
and vertical circulations. Our simulations used idealized profiles of SST in the meridional direction, with no zonal variation and no landmasses. As such, the results of these simulations are highly idealized circulations with relatively little deviation from the time-zonal mean fields. Averages over sufficiently long times and in the zonal direction showed the idealized ITCZ and the primary large-scale overturning that it generates.

Consistent with recently scrutinized observations and reanalyses of some parts of the real ITCZ and with previous idealized simulations, we find that the meridional flow in and out of a zonally symmetric ITCZ consists of four “multilevel” flows: the boundary layer inflow (BLI), the shallow return flow (SRF), the midlevel inflow (MLI), and the upper-level outflow (ULO). The SRF is primarily a sea-breeze-like response to surface gradients of temperature and pressure; as such, its existence depends on the presence of a strong meridional SST gradient near the equator. The MLI owes its existence to diabatically forced convergence around \( z = 8 \) km due to the peak in moist heating around \( z = 9 \) km, which in turn is due to enhanced heating from the latent heat of fusion and the accelerated growth and deposition of condensate due to ice processes. The MLI is enhanced by a reduction in static stability that occurs from 6 to 8 km as the moist adiabat becomes congruent with the dry adiabat, because of the large decrease in the ability of air to hold water with decreasing temperature (Mapes 2001).

For simulations with the ITCZ along the equator, convection is organized by convectively coupled Kelvin waves. A weaker wave is also present that propagates very slowly westward and is similar to a mixed Rossby–gravity wave. For simulations with the ITCZ north of the equator, there is a weak signal of CCKWs along the equator, whereas convection along the ITCZ is organized by the formation and motion of TC-like circulations.

When the SST maximum is off the equator, the multilevel circulations become strong on the equatorward side and weaker or nonexistent on the poleward side. Structurally, the cross-equatorial flows are quite similar to those seen in the ERA-40 for July in the eastern Pacific (Fig. 1b). Much like the real ITCZ in the eastern Pacific, the simulated off-equatorial ITCZs are unstable and frequently spawn TCs that deform and disrupt the ITCZ band. Remarkably, these TCs have only small impact on the time-zonal mean flow on the equatorward side of the ITCZ. They do make a large contribution, however, to the eddy fluxes of water and MSE on the poleward side. As the vertically integrated eddy fluxes

![Fig. 20. Meridional transports of MSE h for HYP2.5–8N: (a) time-zonal mean of total transport \( \rho u h \); (b) eddy fluxes of \( h \); and (c) eddy fluxes of water.](https://example.com/fig20.png)
on the poleward side account for \( \frac{1}{4} - \frac{1}{2} \) of the total vertically integrated transport, it can be said that the eddies, consisting in part of both equatorial waves and tropical cyclones, play a significant role in the MSE budget of the ITZ. This supports the proposition by Sobel and Neelin (2006) and the results of Peters et al. (2008) that horizontal eddy transports of MSE are a significant factor in determining the structure of the ITZ.

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