The Moist Static Energy Budget of a Composite Tropical Intraseasonal Oscillation in a Climate Model

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(Manuscript received 20 March 2008, in final form 23 July 2008)

ABSTRACT

The intraseasonal moist static energy (MSE) budget is analyzed in a climate model that produces realistic eastward-propagating tropical intraseasonal wind and precipitation variability. Consistent with the recharge–discharge paradigm for tropical intraseasonal variability, a buildup of column-integrated MSE occurs within low-level easterly anomalies in advance of intraseasonal precipitation, and a discharge of MSE occurs during and after precipitation when westerly anomalies occur. The strongest MSE anomalies peak in the lower troposphere and are, primarily, regulated by specific humidity anomalies.

The leading terms in the column-integrated intraseasonal MSE budget are horizontal advection and surface latent heat flux, where latent heat flux is dominated by the wind-driven component. Horizontal advection causes recharge (discharge) of MSE within regions of anomalous equatorial lower-tropospheric easterly (westerly) anomalies, with the meridional component of the moisture advection dominating the MSE budget near 850 hPa. Latent heat flux anomalies oppose the MSE tendency due to horizontal advection, making the recharge and discharge of column MSE more gradual than if horizontal advection were acting alone. This relationship has consequences for the time scale of intraseasonal variability in the model.

Eddies dominate intraseasonal meridional moisture advection in the model. During periods of low-level intraseasonal easterly anomalies, eddy kinetic energy (EKE) is anomalously low due to a suppression of tropical synoptic-scale disturbances and other variability on time scales shorter than 20 days. Anomalous moistening of the equatorial lower troposphere occurs during intraseasonal easterly periods through suppression of eddy moisture advection between the equator and poleward latitudes. During intraseasonal westerly periods, EKE is enhanced, leading to anomalous drying of the equatorial lower troposphere through meridional advection. Given the importance of meridional moisture advection and wind-induced latent heat flux to the intraseasonal MSE budget, these findings suggest that to simulate realistic intraseasonal variability, climate models must have realistic basic-state distributions of lower-tropospheric zonal wind and specific humidity.

1. Introduction

The Madden–Julian oscillation (MJO) is the leading mode of intraseasonal variability in tropical winds and convection, characterized by complex interactions between convection and the large-scale atmospheric circulation (e.g., Madden and Julian 2005). While observations have provided us with a good basic description of the spatial and temporal characteristics of the MJO (e.g., Sperber 2003; Kiladis et al. 2005; Tian et al. 2006; and many others), a fundamental understanding of the physical mechanism underlying the MJO remains elusive (e.g., see reviews by Zhang 2005; Waliser 2006). Further, while there have been strides made with climate models in simulating the eastward-propagating, zonal wavenumber 1–3 equatorial circulation features that are characteristic of the MJO (e.g., Zhang et al. 2006), climate models remain unable to simulate many of its other salient features (e.g., Slingo et al. 1996; Sperber et al. 1997; Lin et al. 2006), including the selection of time scale, the coherence and phase between winds and precipitation, the structure and amplitude of precipitation anomalies, and its seasonality.

One school of thought maintains that MJO dynamics are regulated by a recharge–discharge cycle, in which a buildup of column moist static energy (MSE) occurs before MJO deep convection, and MSE is discharged...
during and after MJO convection (e.g., Hendon and Liebmann 1990; Bladé and Hartmann 1993; Hu and Randall 1994; Maloney and Hartmann 1998; Kembel-Cook and Weare 2001; Myers and Waliser 2003; Sobel and Gildor 2003; Kiladis et al. 2005; Agudelo et al. 2006; Tian et al. 2006; Benedict and Randall 2007, and others). Based on recent studies, a buildup in free-tropospheric MSE via moistening would appear necessary before the onset of strong intraseasonal convective events (e.g., Raymond 2000; Bretherton et al. 2004; Sobel et al. 2004; Peters and Neelin 2006). Intraseasonal variability has been demonstrated to increase in general circulation models having convection parameterizations that exhibit strong sensitivity to free-tropospheric humidity (e.g., Wang and Schlesinger 1999; Woolnough et al. 2001), and the time scale of free-tropospheric moistening may help regulate the amplitude of the MJO precipitation anomalies. Some studies extend the recharge–discharge paradigm to include the ocean, in which a buildup of ocean heat content occurs before the onset of MJO deep convection, and this anomalous heat content is discharged to the atmosphere during the MJO convective phase (e.g., Sobel and Gildor 2003; Stephens et al. 2004; Agudelo et al. 2006).

The regulation of the atmospheric MSE budget during MJO events is not well understood. Shallow convection and associated shallow vertical circulations may help moisten the atmospheric column and contribute to moist static energy buildup before MJO deep convective commencences (e.g., Johnson et al. 1999; Kiladis and Takayuba 2004; Kiladis et al. 2005; Benedict and Randall 2007). Column-integrated MSE may then be discharged during periods of strong MJO convective and stratiform heating. Observational evidence suggests that large-scale circulations in the western Pacific and Indian Oceans, where the mean vertical velocity profile is strongly influenced by deep convection, export MSE in the mean (e.g., Neelin and Held 1987; Back and Bretherton 2006). MJO amplitude maximizes in these regions, and MSE discharge may be expected to be enhanced during the deep convective and stratiform phases of the MJO [e.g., Johnson et al. (1999); Kiladis et al. (2005); see also the schematic diagram derived for mesoscale precipitation events and Kelvin waves in Peters and Bretherton (2006)]. MSE discharge by vertical motions during MJO convective events may be modified by cloud–radiative feedbacks and wind–evaporation feedbacks that could slow, or even reverse, the MSE discharge (e.g., Raymond 2001; Lin and Mapes 2004; Peters and Bretherton 2006; Sugiyama 2009a,b, hereafter SJAS). Intraseasonal wind speed and latent heat flux anomalies are observed to have a positive covariance with intraseasonal precipitation (e.g., Zhang 1996; Raymond et al. 2003; Masunaga et al. 2006; Maloney and Esbensen 2007; Araligidad and Maloney 2008), and wind–evaporation feedbacks have been found to be important for supporting intraseasonal convection in recent modeling studies (e.g., Raymond 2001; Maloney and Sobel 2004; Fuchs and Raymond 2005; SJAS). These intraseasonal latent heat flux–precipitation relationships are consistent with the more general role of enhanced wind speed in supporting tropical precipitation; particularly in regions of sufficiently high column relative humidity (e.g., Back and Bretherton 2005). Anomalous longwave radiative heating has also been observed to augment column-integrated MSE during enhanced precipitation phases of the MJO (e.g., Lin and Mapes 2004).

Recent observational evidence suggests that horizontal moisture and temperature advection cannot be neglected as an important regulatory mechanism for the atmospheric MSE budget and deep convection (e.g., Back and Bretherton 2006). As a prime example, Mapes and Zuidema (1996) documented the frequent occurrence of dry tongues in tropical Pacific soundings in which low moist static energy air is advected into the equatorial region from the subtropics. Further, advection of dry air from the subtropics has been cited as a possible means of drying and stabilizing the atmosphere subsequent to the onset of deep MJO convection in both observations (e.g., Myers and Waliser 2003), and in models (e.g., SJAS). Benedict and Randall (2007) note that drying by horizontal advection might be most important to the MJO in the Indian Ocean due to its proximity to dry Asian land areas.

This study explores the MSE budget in an atmospheric general circulation model that produces eastward-propagating tropical wind and precipitation variability that has similar amplitude and propagation speed to the observed MJO. A modified version of the National Center for Atmospheric Research (NCAR) Community Atmosphere Model version 3.1 (CAM3) is used in which a relaxed Arakawa–Schubert (RAS) convection parameterization (Moorthi and Suarez 1992) is implemented to replace the standard Zhang and McFarlane (1995) convection scheme. A predecessor version of this model configuration [CAM2.0 with RAS convection; see Maloney and Sobel (2004)] produced a realistic intraseasonal oscillation in which equatorial precipitation variability was strongly regulated by wind–evaporation feedback. As will be shown below, the climate model used here exhibits a pattern of behavior that is consistent with the recharge–discharge paradigm, in which column MSE builds before deep intraseasonal convection commences, and then is discharged after the onset of deep convection. Analysis of the model moist static energy budget will show a strong role
for horizontal advection in regulating this recharge–discharge cycle.

This analysis will also suggest an important role for tropical synoptic-scale eddies in the recharge–discharge process. Previous work has shown that tropical synoptic-scale eddy activity is enhanced during the convectively active phase of the MJO in the western Pacific and Indian Oceans, and eddy activity is suppressed during the MJO convectively suppressed phase (e.g., Nakazawa 1986; Hendon and Liebmann 1990; Straub and Kiladis 2003; Maloney and Dickinson 2003; Batstone et al. 2005). The influence of extratropical high-frequency transients on the equatorial region also varies with the phase of the MJO (e.g., Matthews and Kiladis 1999).

More generally, previous work has suggested important mutual feedbacks between the MJO and faster timescale–shorter space-scale disturbances (e.g., Moncrieff 2004). For example, modeling work has suggested that upscale transfers of temperature and momentum from the synoptic scale to planetary scale may help maintain large-scale MJO circulations (e.g., Biello and Majda 2005; Biello et al. 2007). Further, variations in synoptic-scale activity during an MJO life cycle can generate a significant upscale modulation of the MJO time-scale wind speed. This eddy influence can either augment or cancel wind speed fluctuations forced by intraseasonal vector wind components, depending on location (e.g., Batstone et al. 2005; Araligidad and Maloney 2008). High-frequency/high-wavenumber features of the flow can also regulate the moisture budget above the boundary layer. For example, Houze et al. (2000) showed that midlevel convective inflow within larger western Pacific convective envelopes can be an important drying agent, in addition to being an important agent of mass and momentum exchange.

The model used in this study is described in section 2. Section 3 will present some first-order diagnostics of the model’s ability to produce realistic eastward-propagating tropical intraseasonal variability during November–April. This section will also comment upon the sensitivity of the model intraseasonal variability to the convective rain reevaporation fraction. Section 4 will examine the MSE budget of model intraseasonal disturbances. Section 5 will examine the importance of higher-frequency transient eddies to the intraseasonal MSE budget. Section 6 presents our conclusions and a discussion.

2. Model description

The National Center for Atmospheric Research (NCAR) Community Atmosphere Model Version 3.1 (CAM3; Collins et al. 2006) is used in this study. The CAM3 is the atmosphere component of the NCAR Community Climate System Model version 3. Tropical intraseasonal variability in the standard version of the CAM3 with the Zhang and McFarlane (1995) convection parameterization is significantly weaker than observed (e.g., Mu and Zhang 2006). As in previous modeling studies such as Maloney and Sobel (2004), the Zhang and McFarlane (1995) deep convection parameterization is replaced with the relaxed Arakawa–Schubert convection parameterization of Moorthi and Suarez (1992) to improve model intraseasonal variability. In the version of RAS used here, convection is initiated from the lowest model level. The Hack (1994) parameterization is retained to simulate shallow convection.

Convective adjustment with RAS is governed by a cloud work function, which is very strongly regulated by the MSE of the boundary layer. Yang et al. (1999) did an adjoint sensitivity analysis to show that RAS precipitation responds most strongly to humidity variations in the boundary layer, with comparatively little sensitivity to free-tropospheric humidity. However, the version of RAS used here includes a minimum entrainment rate for cloud ensemble members that may affect the sensitivity of RAS to humidity above the level of convective initiation. Cloud ensemble members that entrain less than a prescribed minimum entrainment parameter are suppressed. Tokioka et al. (1988) developed a version of the Arakawa–Schubert convection scheme that directly tied the minimum entrainment rate ($\mu_{\text{min}}$) to the sounding-dependent depth in meters of the boundary layer ($D$), such that $\mu_{\text{min}} = \alpha/D$ (where $\alpha$ is a constant). A realistic MJO simulation was obtained in Tokioka et al. (1988) by setting $\alpha = 0.1$. The scheme used here fixes the minimum entrainment at a rate of 0.1/(1000 m). While the minimum entrainment rate and thus triggering of deep convection in the model is not allowed to vary with variations in boundary layer depth, this minimum entrainment rate is of the same order as that prescribed in Tokioka et al. (1988) and may, thus, help explain the ability of the model to produce a reasonable MJO. Given how entrainment is calculated within RAS for a given cloud ensemble member (e.g., Moorthi and Suarez 1992), and in the presence of weak tropical temperature gradients (e.g., Sobel and Bretherton 2000) and relatively small temporal variability in the boundary layer moist static energy, clouds with very small entrainment rates (typically deep clouds) are suppressed until the free troposphere is sufficiently moistened. Sufficient moistening can increase entrainment rates in RAS and allow for the existence of the deepest clouds. The details of how this minimum entrainment parameter regulates the intraseasonal
variability in the model used here are a topic of ongoing work.

The version of RAS used here also allows a certain fraction \( \alpha \) of the convective precipitation generated to be exposed to environmental air and allowed to evaporate, subject to environmental relative humidity, temperature, and pressure, and also microphysical assumptions such as the droplet size distribution and fall velocities (e.g., Sud and Molod 1988). One can interpret the evaporation fraction \( \alpha \) as the fraction of convective precipitation falling outside of the saturated downdraft cores. As will be shown below, intraseasonal variability in the model is extremely sensitive to \( \alpha \). The control simulation analyzed throughout this paper has \( \alpha \) set to 0.6 to maximize the intraseasonal variability, although a sensitivity test is presented in section 3 where \( \alpha \) is varied from 0.05 to 0.6 to demonstrate the impact on model intraseasonal variability. In the parameterization used here, convective rain reevaporation is not allowed to explicitly generate subgrid-scale downdraft mass fluxes, although the rain reevaporation parameterization likely has significant impacts on vertical velocities on the resolved scale through modifications to the grid-scale heating and thus has impacts on resolved temperatures and humidity (e.g., Johnson 1976).

Sixteen-year simulations of CAM3 with RAS convection are conducted for \( \alpha = 0.6 \) and \( \alpha = 0.05 \), with \( \alpha \) set to 0.6 in the primary simulation analyzed. Simulations are integrated using a spectral dynamical core at T42 resolution (approximately \( 2.8^\circ \times 2.8^\circ \) grid resolution) with climatological seasonal cycle sea surface temperatures and insolation. Twenty-six levels are resolved in the vertical, and the model time step is 20 min.

3. Diagnostics of eastward-propagating variability

\textit{a. General diagnostics and sensitivity to rain reevaporation}

The control simulation used here simulates robust eastward-propagating intraseasonal variability in a comparable manner to the related model configuration used in Maloney and Sobel (2004). This ability is briefly documented here. The analysis is confined to November–April, when the observed equatorial convective variability associated with the MJO is at its maximum in amplitude (e.g., Madden and Julian 2005).

In Fig. 1 and during the rest of the paper, 30–90-day bandpass-filtered fields are constructed using two 60-weight linear nonrecursive filters with half-power points at 30 and 90 days. Observed winds are taken from National Centers for Environmental Prediction–National Center for Atmospheric Research (NCEP–NCAR)
reanalysis data (Kalnay et al. 1996), and daily averaged precipitation data are taken from the Tropical Rainfall Measuring Mission (TRMM) level 3B-42 version 6 product (e.g., Huffman et al. 2001). Data during 1998–2006 are used. Figure 1 shows observed and modeled equatorial (10°N–10°S averaged) 30–90-day bandpass-filtered 850-hPa zonal wind (contours) and precipitation (color) anomalies regressed onto a filtered 850-hPa zonal wind reference time series at 155°E as a function of lag.

Both observed and model intraseasonal wind anomalies exhibit coherent eastward propagation across the western Pacific Ocean, with average propagation speeds of 7.8 m s$^{-1}$ in the observations and 7.0 m s$^{-1}$ in the model (as calculated between 80°E and 180°). The amplitude of the model western Pacific intraseasonal variability is slightly higher than that found in the observations. At least in the western Pacific, the amplitude of the model precipitation anomalies and the phase relationship relative to the zonal wind anomalies are similar to observed. The magnitudes of the model western Pacific wind and precipitation anomalies will be better quantified later (see Fig. 3). Consistent with the results of Maloney and Sobel (2004), intraseasonal wind and precipitation variabilities in this model are weaker than those observed in the Indian Ocean. Hence, our results will concentrate on eastward-propagating intraseasonal variability in the western Pacific. However, this model captures slow eastward-propagating zonal wind and precipitation variabilities across the Pacific with similar propagation speeds and amplitudes to those of the observed MJO, an essential measure of success in any model–observation comparison of intraseasonal variability.

Wavenumber–frequency spectra of equatorial (10°N–10°S averaged) 850-hPa zonal winds for the NCEP–NCAR reanalysis, the simulation with $\alpha = 0.05$, and the primary simulation with $\alpha = 0.6$ are shown in Fig. 2. The observed spectrum is characterized by enhanced eastward power centered at zonal wavenumber 1 and 30–90-day periods characteristic of the MJO. The simulation with the reduced reevaporation fraction ($\alpha = 0.05$) is characterized by significantly diminished power at low
wavenumbers and eastward intraseasonal periods as compared to the observations. The strong reevaporation run ($\alpha = 0.6$) is characterized by eastward spectral power at low wavenumbers and intraseasonal frequencies that are comparable to the observations, with a similar preference for eastward propagation versus westward propagation. The $\alpha = 0.6$ simulation has a spectrum that is more red than the observed spectrum, a common problem in many general circulation models that exhibit strong intraseasonal variability (e.g., Zhang et al. 2006).

The strong dependence on the reevaporation fraction is consistent with previous MJO studies that have cited the relevance of moisture–convection feedbacks, whereby tropospheric moistening by convection makes the troposphere more favorable for subsequent deep convection, promoting local convective organization. As explained in Grabowski and Moncrieff (2004) and Bony and Emanuel (2005), higher reevaporation fractions should increase the strength of the moisture–convection feedback. In fact, Grabowski and Moncrieff (2004) conducted an experiment similar to that described here in a model with the Emanuel convective parameterization, and found that a strengthened model MJO resulted from an increased reevaporation fraction.

An alternate hypothesis to the moisture–convection feedback in explaining the model sensitivity to rain reevaporation is suggested by Maloney and Hartmann (2001). In their model experiments using a version of RAS, Maloney and Hartmann (2001) showed that rain reevaporation and convective downdrafts improved intraseasonal variability through their impact on the mean climate of the model. Downdrafts and rain reevaporation moistened the lower and middle equatorial troposphere in the mean, which appeared to be the crucial basic-state change for producing realistic intraseasonal variability. As shown below (Fig. 6), increasing the reevaporation fraction similarly moistens the equatorial troposphere in the model. As a first attempt at explaining this sensitivity, it was examined whether a moister equatorial troposphere with an increased reevaporation fraction could augment column-integrated moisture convergence per unit vertical mass flux, thus decreasing the mean gross moist stability (e.g., Neelin and Held 1987). An analysis of the normalized gross moist stability calculated as the mean column-integrated MSE export due to vertical advection per unit dry static energy export due to vertical advection did not suggest any significant changes in equatorial Pacific convective regions as a function of reevaporation fraction (not shown here). However, the MSE budget analysis described in section 4 below does suggest another interesting hypothesis for how moistening the equatorial troposphere improves the model intraseasonal variability, involving horizontal moisture advection.

b. Composites

The simulation with $\alpha = 0.6$ will be examined throughout the rest of this paper. Because the strongest intraseasonal variability in this model occurs in the western Pacific, a composite model intraseasonal event is generated based on equatorial (10$^\circ$S–10$^\circ$N averaged) 30–90-day bandpass-filtered 850-hPa zonal wind events at 155$^\circ$E. Intraseasonal events are defined as positive maxima in the bandpass-filtered zonal wind time series at 155$^\circ$E when they exceed one standard deviation from zero. Using this criterion, 34 events were selected during November–April of the 16-yr simulation. A composite intraseasonal event is derived by averaging fields across all 34 events, with evolution examined as a function of time relative to the event maximum.

Composite equatorial (10$^\circ$S–10$^\circ$N averaged) 850-hPa zonal wind and precipitation anomalies at 155$^\circ$E are displayed in Fig. 3. Peak intraseasonal precipitation, having an amplitude of about 1.5–2.0 mm day$^{-1}$, leads the maximum zonal winds by about 5–10 days. The amplitude of the zonal wind anomalies is stronger than is observed, which is consistent with Fig. 1. However, the amplitude of the equatorial averaged precipitation anomalies compares very favorably to that associated with the observed MJO in the western Pacific, as does the phase relationship between precipitation and the 850-hPa zonal wind anomalies [as inferred by comparison to Fig. 1 in Maloney and Sobel (2004)]. The spatial structures of the model’s intraseasonal 850-hPa wind anomalies, precipitation anomalies, and latent heat flux anomalies at the time of strongest
equatorial averaged 155°E precipitation are shown in Fig. 4. As in Maloney and Sobel (2004), boreal winter precipitation anomalies in the model exhibit two off-equatorial precipitation maxima symmetric about the equator, unlike observed precipitation anomalies that are characterized by a dominant precipitation maximum in the Southern Hemisphere centered near 7.5°–10°S (see Fig. 12 of Maloney and Sobel 2004). As in observations (e.g., Lau and Sui 1997; Zhang and McPhaden 2000), enhanced latent heat fluxes occur near and to the west of anomalous precipitation within regions of anomalous westerly flow.

In summary, the $\alpha = 0.6$ simulation analyzed here captures the eastward-propagating equatorial zonal wind variability in the western Pacific warm pool with similar propagation speed to the observed MJO and an amplitude somewhat larger than that observed. The eastward-propagating wind anomalies are associated with equatorial precipitation anomalies of similar amplitude to the observed MJO, and these precipitation anomalies have a similar phase relationship to the anomalous zonal wind as in the observations. Intra-seasonal variability in the model exhibits some features that significantly differ from the observed MJO in the following ways. For one, the spatial structures of the intra-seasonal precipitation anomalies exhibit significant biases relative to the observations. Second, Indian Ocean intra-seasonal precipitation anomalies and wind variabilities are weaker than observed. Last, the frequency spectrum of the model winds and precipitation is redder than the observed spectrum, which is characterized by more concentrated power at 30–90-day periods. However, the realistic eastward propagation speed, amplitude, and phase relationships that characterize the equatorial intra-seasonal variability suggest that this model captures some of the essential characteristics of the observed MJO. Thus, examining the model intra-seasonal MSE budget should be a useful exercise.

4. Moist static energy budget

Figure 5a shows the composite vertical distribution of the intra-seasonal MSE anomalies at 155°E as a function of time relative to the peak westerly anomalies. MSE is defined as $m = c_p T + g z + L q$, where $T$ is temperature, $c_p$ is the specific heat at constant pressure, $z$ is height, $g$ is the gravitational acceleration, $L$ is the latent heat of vaporization at 0°C, and $q$ is the specific humidity. As a
reference, recall that precipitation peaks near day –5 (Fig. 3).

As in the observations (e.g., Kemball-Cook and Weare 2001), lower-tropospheric model m anomalies grow before the onset of intraseasonal convection, with 850-hPa m anomalies peaking around day –15. Anomalies in m peak in the middle to upper troposphere near day –5, in association with the maximum in precipitation (e.g., Fig. 3). Strong lower-tropospheric drying commences around day –5 in association with enhanced intraseasonal precipitation, and negative 850-hPa m anomalies reach their highest magnitude near day +10. As is shown in Fig. 5b, the m anomalies in the lower troposphere are almost entirely regulated by water vapor. Upper-tropospheric m anomalies are primarily caused by temperature anomalies.

Interestingly, peak lower-tropospheric m anomalies in the model reach only 550 J kg\(^{-1}\), about one-quarter to one-third of the observed amplitude (e.g., Kemball-Cook and Weare 2001; Kiladis et al. 2005), although the magnitude of the precipitation anomalies is close to the observed. These results suggest that the convective parameterization in the model may not be fostering the same buildup of instability as in the observations, possibly because of the insufficient sensitivity of the convection parameterization to humidity above the boundary layer. This trait is exhibited by many modern convection parameterizations (e.g., Derbyshire et al. 2004). The results may also suggest that the discharge of m in the model is associated with a larger moisture convergence per unit mass flux than in the observations, since model precipitation anomalies are of the same amplitude as the observed for a smaller m buildup. Figure 6 shows that oceanic column-integrated precipitable water during November–April is higher near the equator in the simulation with evaporation fraction equal to 0.05 (black, dashed), the primary simulation (black, solid), and SSM/I (gray, solid). The SSM/I data during 1992–2007 were used in this analysis. The SSM/I data were averaged to 2.5° × 2.5° cells, and then interpolated to the model grid. SSM/I data are missing over land, and hence model data over land are not included in the zonal average.
anomalies between the model and the observations should be kept in mind when comparing the simulated variability to reality.

The vertically integrated \( m \) budget of the model intraseasonal wind and precipitation events is now examined. As in Neelin and Held (1987), the vertically integrated \( m \) budget can be written as

\[
\left\langle \frac{\partial m}{\partial t} \right\rangle + \left\langle \frac{\partial m}{\partial p} \right\rangle + \langle \mathbf{v} \cdot \nabla m \rangle - \text{LH} - \text{SH} - \langle \text{LW} \rangle - \langle \text{SW} \rangle = 0,
\]

where the brackets represent a mass-weighted vertical integral from the surface to the top of the troposphere, here taken to be 100 hPa, although a vertical integral to the top of the atmosphere produces similar results. The pressure is \( p \), \( \mathbf{v} \) is the horizontal wind vector, and \( \omega \) is the vertical pressure velocity. Gradient operators apply along surfaces of constant pressure. The first term in (1) represents the vertically integrated \( m \) tendency, the second term represents the column-integrated export of \( m \) due to vertical advection, and the third term represents the horizontal advection of \( m \). The surface latent heat flux is LH, SH is the surface sensible heat flux, \( \langle \text{LW} \rangle \) represents the vertically integrated longwave heating rate, and \( \langle \text{SW} \rangle \) represents the vertically integrated shortwave heating rate.

Of interest are 30–90-day time-scale variations in the vertically integrated \( m \) budget. Thus, a 30–90–day bandpass filter can be applied to (1) to produce

\[
\left\langle \frac{\partial m}{\partial t} \right\rangle_{\text{ISO}} = -\left\langle \frac{\partial m}{\partial p} \right\rangle_{\text{ISO}} - \langle \mathbf{v} \cdot \nabla m \rangle_{\text{ISO}} + \text{LH}_{\text{ISO}} + \text{SH}_{\text{ISO}} + \langle \text{LW} \rangle_{\text{ISO}},
\]

where quantities with a subscript ISO represent a 30–90–day bandpass-filtered field. Note that \( \langle \text{SW} \rangle_{\text{ISO}} \) is negligible in the model and is, thus, omitted from (2).

Equatorial-averaged composites at 155°E are generated for budget terms in (2) and are displayed in Fig. 7. In Fig. 7, LHISO and SHISO are combined, since LHISO is an order of magnitude larger than SHISO. Composite 30–90–day precipitation anomalies in energy units (LPISO) are also shown as a reference.

The term \( \langle \partial m / \partial t \rangle_{\text{ISO}} \) is approximately in quadrature with the precipitation. The recharge of \( m \) occurs before peak precipitation, with the discharge of \( m \) occurring during and after the precipitation event, which is consistent with Fig. 5. The vertically integrated \( m_{\text{ISO}} \) budget is dominated by LHISO and \( -\langle \mathbf{v} \cdot \nabla m \rangle_{\text{ISO}} \). The term \( -\langle \mathbf{v} \cdot \nabla m \rangle_{\text{ISO}} \) is large and positive before peak precipitation, and \( -\langle \mathbf{v} \cdot \nabla m \rangle_{\text{ISO}} \) is strong and negative during and after peak intraseasonal precipitation, coinciding with the discharge of column \( m \). The dominance of the horizontal \( m \) advection over the vertical \( m \) advection in the \( m \) budget is surprising, given previous results that suggest shallow cumulus clouds and associated vertical motions are important for moistening the lower troposphere before the onset of deep MJO convection (e.g., Kiladis et al. 2005). Shallow convection and associated shallow heating and vertical motion profiles would be accompanied by convergence near cloud base and divergence near cloud-top level. Given that \( m \) decreases with height from the boundary layer to the middle troposphere, the vertically integrated effect of shallow convection would be to increase column \( m \) (e.g., Peters and Bretherton 2006). The term \( -\langle \mathbf{v} \cdot \nabla m \rangle_{\text{ISO}} \), which by mass continuity and integration by parts over the depth of the troposphere is equal to \( -\langle \mathbf{m} \cdot \nabla \mathbf{v} \rangle_{\text{ISO}} \), is generally out of phase with the precipitation and of relatively modest amplitude. An examination of the vertical distributions of the anomalous vertical velocity, diabatic heating, and divergence fields supports the relatively modest role for shallow convection in the model \( m \) budget (not shown here).

The term LHISO is generally out of phase with \( -\langle \mathbf{v} \cdot \nabla m \rangle_{\text{ISO}} \), although it is of slightly smaller amplitude, such that the sign of the sum of these terms is the same as that of \( -\langle \mathbf{v} \cdot \nabla m \rangle_{\text{ISO}} \). It appears that LHISO slows the recharge of \( m \) by \( -\langle \mathbf{v} \cdot \nabla m \rangle_{\text{ISO}} \) before the peak precipitation, and retards the discharge of \( m \) by \( -\langle \mathbf{v} \cdot \nabla m \rangle_{\text{ISO}} \) during and after the intraseasonal precipitation event. Comparing Figs. 3 and 7, it is seen that...
\(- (v \cdot \nabla m)_{ISO}\) recharges column \(m\) within the 850-hPa easterly anomalies and discharges \(m\) within the 850-hPa westerly anomalies. The interplay of \(LH_{ISO}\) and \(- (v \cdot \nabla m)_{ISO}\) may regulate the time scale of the \(m\) recharge–discharge process during intraseasonal precipitation events in the model. Previous evidence also suggests an important role for \(LH_{ISO}\) in regulating intraseasonal precipitation. In a study that used a predecessor version of the model employed here, Maloney and Sobel (2004) found that \(LH_{ISO}\) was crucially important for producing realistic intraseasonal precipitation variability. In the model, \(LH_{ISO}\) is about 20% of \(LP_{ISO}\) at the time of peak precipitation, a magnitude consistent with the western Pacific observational estimates of Araligidad and Maloney (2008). Model \((LW)_{ISO}\) is about 10% of \(LP_{ISO}\) at the time of peak precipitation, a magnitude that is also consistent with the observations (e.g., Lin and Mapes 2004).

Before proceeding, it is noted that the relationships shown in Fig. 7 are similar across the western Pacific warm pool. As a demonstration, \(- (v \cdot \nabla m)_{ISO}\), the sum of \(- (v \cdot \nabla m)_{ISO}\) and \(LH_{ISO}\), and \((\frac{\partial m}{\partial t})_{ISO}\) are regressed onto equatorial \(U_{850ISO}\) at 155°E as a function of lag in days during November–April (using the method in Fig. 1b). The term \((\frac{\partial m}{\partial t})_{ISO}\) is consistently positive within easterly 850-hPa zonal wind anomalies and negative within westerly 850-hPa zonal wind anomalies (Fig. 8c). Note the smaller contour interval for Fig. 8c than in the rest of the plots. Figure 8a shows that \(- (v \cdot \nabla m)_{ISO}\) consistently recharges column \(m\) within regions of easterly 850-hPa zonal wind anomalies and discharges column \(m\) within regions of anomalous westerlies, although with notable disruptions in the Maritime Continent region near the longitudes of Borneo and New Guinea (where vertical advection also increases in importance due to orographic effects; not shown). The relative behavior of \(- (v \cdot \nabla m)_{ISO}\) and \(LH_{ISO}\) also remains consistent across the warm pool (Fig. 8b), with \(LH_{ISO}\) consistently being of opposite sign to, but slightly smaller amplitude than, \(- (v \cdot \nabla m)_{ISO}\). The recharge and discharge of column \(m\) is thus effectively slowed by \(LH_{ISO}\) versus if \(- (v \cdot \nabla m)_{ISO}\) were acting alone.

**Fig. 8.** Lag regression of the model equatorial (10°S–10°N averaged) intraseasonal 850-hPa zonal wind anomalies (colors) along with (a) vertically integrated horizontal \(m\) advection anomalies (contours), (b) the sum of the vertically integrated horizontal \(m\) advection anomalies plus the surface latent heat flux anomalies (contours), and (c) the \(m\) tendency onto a reference zonal wind time series at 155°E. Regression coefficients are scaled by a 1σ value of the reference time series. Only data during November–April are used. Contours in (a) and (b) are plotted every 5 W m\(^{-2}\), starting at 2.5 W m\(^{-2}\). Contours in (c) are plotted every 2 W m\(^{-2}\), starting at 1 W m\(^{-2}\). Negative contours are dashed.
Given the importance of horizontal $m$ advection to the intraseasonal $m$ budget in the model, the processes responsible for the regulation of $-(v \cdot \nabla m)_{ISO}$ are now diagnosed in further detail. Figure 9 shows the composite vertical distribution of $-(v \cdot \nabla m)_{ISO}$ at 155°E as a function of time relative to the peak westerly winds. Consistent with the relatively large cycle in lower-tropospheric $m_{ISO}$ shown in Fig. 5, $-(v \cdot \nabla m)_{ISO}$ peaks in amplitude between 700 and 850 hPa. Recharge of $m_{ISO}$ by $-(v \cdot \nabla m)_{ISO}$ occurs before precipitation peaks, with subsequent $m_{ISO}$ discharge by $-(v \cdot \nabla m)_{ISO}$. Secondary amplitude maxima of $-(v \cdot \nabla m)_{ISO}$ occur in the upper troposphere, although the contribution of these upper-tropospheric anomalies to the mass-weighted vertical integral of (2) are relatively modest.

The term $-(v \cdot \nabla m)_{ISO}$ can be further partitioned into zonal and meridional components as follows:

$$-(v \cdot \nabla m)_{ISO} = - \left( u \frac{\partial m}{\partial x} \right)_{ISO} - \left( v \frac{\partial m}{\partial y} \right)_{ISO}, \quad (3)$$

where the $m_{ISO}$ variations near 700 hPa have their largest contributions from $-(u \frac{\partial m}{\partial x})_{ISO}$ (Fig. 10), whereas those near 850 hPa have their largest contributions from $-(u \frac{\partial m}{\partial y})_{ISO}$. Here, $-(u \frac{\partial m}{\partial x})_{ISO}$ tends to be more vertically oriented than $-(u \frac{\partial m}{\partial y})_{ISO}$, which tilts in time toward the west. The tilt of $-(u \frac{\partial m}{\partial x})_{ISO}$ with time can also be interpreted as a westward tilt with height, given the eastward propagation of the intraseasonal anomalies. This is confirmed by examining the composite height–longitude structure (not shown). The tilt of $-(u \frac{\partial m}{\partial x})_{ISO}$ with height is consistent with the tilted structure of $m_{ISO}$ in Fig. 5. Composite zonal wind anomalies exhibit little vertical tilt (not shown), and so the $-(u \frac{\partial m}{\partial y})_{ISO}$ anomalies at 500 hPa in Fig. 10 and their tilt relative to 700 hPa appear to be due to the advection of $m_{ISO}$ by the mean zonal flow.

The processes responsible for the regulation of $-(u \frac{\partial m}{\partial x})_{ISO}$ and $-(v \frac{\partial m}{\partial y})_{ISO}$ during a composite intraseasonal wind event are diagnosed by separating the variables into their time mean component and deviation from the time mean. For example, $u$ is written as $u = \bar{u} + u'$, where $\bar{u}$ represents a 51-day running mean and $u'$ represents the deviation from the 51-day running mean. The analysis presented here is not sensitive to the exact means by which the time average is defined. Thus,
to good approximation, \( (\overline{u^m}_{x/y})_{ISO} \) and \( (\overline{u'^m}_{x/y})_{ISO} \) can be written as

\[
\left( \frac{\partial \overline{m}}{\partial x} \right)_{ISO} \approx \left( \frac{\partial \overline{m}}{\partial x} \right)_{ISO} + \left( \frac{\partial \overline{m'}}{\partial x} \right)_{ISO}
\]

and

\[
\left( \frac{\partial \overline{m}}{\partial y} \right)_{ISO} \approx \left( \frac{\partial \overline{m}}{\partial y} \right)_{ISO} + \left( \frac{\partial \overline{m'}}{\partial y} \right)_{ISO}
\]

The extent to which the sum of the terms on the right of (4) and (5) reproduces \( (\overline{u^m}_{x/y})_{ISO} \) and \( (\overline{u'^m}_{x/y})_{ISO} \) was verified by examination of the vertical distribution of these quantities, and the agreement was found to be excellent (not shown).

Figure 11 shows the partitioning of \(- (\overline{u^m}_{x/y})_{ISO} \) as prescribed by (4) and also includes the sum of the terms on the right-hand side (total). Quantities were vertically integrated from 500 hPa to the surface. Advection by the perturbation zonal flow across the mean \( m \) gradient \(- (\overline{u^m}_{x/y})_{ISO} \) is the most important lower-tropospheric contributor to the flow. The contributions from \(- (\overline{u^m}_{x/y})_{ISO} \) and \(- (\overline{u'^m}_{x/y})_{ISO} \) are nonnegligible but are of lesser amplitude.

It is doubtful that the term \(- (\overline{u^m}_{x/y})_{ISO} \) is responsible for the continuous propagation of intraseasonal anomalies across the western Pacific seen in the model. The model lower-tropospheric \( \overline{u^m}_{x/y} \) changes sign across the equatorial western Pacific (not shown). The observed mean zonal lower-tropospheric humidity gradient is also not of consistent sign across the Indian and western Pacific Oceans, making its contribution to the anomalous moistening and eastward propagation inconsistent across the region where the MJO is observed to be strong. Further, a comparison of the mean zonal specific humidity gradients among the model and the NCEP–NCAR reanalysis and the 40-yr European Centre for Medium-Range Weather Forecasts (ECMWF) Re-Analysis (Uppala et al. 2005) indicates that the direction of the mean model western Pacific zonal humidity gradient near 155°E is opposite to that in the observations. The mean humidity gradient controls the \( m \) gradient in the model and observations due to the relative homogeneity of the tropical tropospheric temperatures (e.g., Sobel and Bretherton 2000). Mean lower-tropospheric specific humidity in the model increases from the Maritime Continent eastward (not shown here), whereas the mean humidity gradient in the observations appears flat (NCEP–NCAR) or directed westward (ERA-40). Thus, whereas an easterly lower-tropospheric wind anomaly in the model would tend to moisten in advance of precipitation at 155°E, comparable wind anomalies in the observations would generate anomalous dry advection. As a final point, the mean westerly flow in the lower troposphere is stronger in the western Pacific in the model than in either observational reanalysis product, suggesting that \(- (\overline{u^m}_{x/y})_{ISO} \) may be unrealistically strong in the model.

Figure 12 shows the partitioning of the composite \(- (\overline{u^m}_{x/y})_{ISO} \) at 155°E, as described by (5), and integrated from 500 hPa to the surface. Recall that composite \(- (\overline{u^m}_{x/y})_{ISO} \) peaks near 850 hPa. Interestingly, \(- (\overline{u^m}_{x/y})_{ISO} \) is dominated by \(- (\overline{m'}_{x/y})_{ISO} \). The regulatory mechanisms for \(- (\overline{m'}_{x/y})_{ISO} \) will be examined in section 5. Notably, this analysis will suggest a role for tropical synoptic eddies in regulating the intraseasonal \( m \) budget.

5. A role for eddies in regulating the MSE budget

A comparison of Figs. 3 and 12 indicates that \(- (\overline{m'}_{x/y})_{ISO} \) recharges \( m_{ISO} \) during periods of 850-hPa easterly anomalies, and discharges \( m_{ISO} \) during periods of 850-hPa westerly anomalies. Note that primed terms not only include anomalies on the 30–90-day time scale, but also include higher-frequency (<30 day) variability. To provide a clue as to what is driving the \(- (\overline{m'}_{x/y})_{ISO} \)
variations, composite intraseasonal anomalies in eddy kinetic energy (EKE) at 155\degree E are plotted for 850 and 700 hPa (Fig. 13). Eddy kinetic energy is defined as

$$EKE = \frac{u'^2 + v'^2}{2},$$

(6)

where double-primed quantities represent deviations from the 11-day running mean. Model EKE is anomalously low during periods of intraseasonal easterly anomalies and anomalously high during periods of intraseasonal westerly anomalies. The enhancement (suppression) of EKE during periods of intraseasonal westerly (easterly) flow anomalies is consistent with that observed. For example, Maloney and Dickinson (2003) and Batstone et al. (2005) demonstrated a significant modulation of EKE during MJO events, with suppression of synoptic eddies during MJO easterly phases, and enhancement during westerly phases. The amplitudes of the EKE variations in those studies compare favorably to that found here.

Given this modulation of EKE during intraseasonal wind events, it is plausible that higher-frequency transient disturbances have a significant influence on the dominant horizontal advection term $- \left( \left[ u' \frac{\partial m'}{\partial y} \right]_{ISO} \right)$. To test this influence, a vertically integrated (500 hPa to the surface) composite $- \left( \left[ u' \frac{\partial m'}{\partial y} \right]_{ISO} \right)$ at 155\degree E is plotted, along with the same term after the first high-pass filtering $u'$ and $\frac{\partial m'}{\partial y}$ to retain periods of less than 30 days (Fig. 14). As shown in Fig. 14, high-frequency transients generate most of the composite $- \left( \left[ u' \frac{\partial m'}{\partial y} \right]_{ISO} \right)$ signal in the model, suggesting that the suppression of eddies during periods of intraseasonal easterly flow enables anomalous moistening via meridional advection, and enhancement of eddies during intraseasonal westerly flow causes anomalous drying via meridional advection. About 70\% of the eddy contribution to $- \left( \left[ u' \frac{\partial m'}{\partial y} \right]_{ISO} \right)$ comes from time scales retained in the EKE calculation of (6) (effectively less than the 22-day period).

Maps of the 850-hPa EKE anomalies ($EKE_{ISO}$) during the times of peak anomalous horizontal advection moistening (day –20) and drying (day 0) are shown in Fig. 15. Also shown are 30–90-day bandpass-filtered 850-hPa wind vectors. Model EKE is suppressed on the poleward flanks of the 850-hPa easterly anomalies in regions of anomalous anticyclonic zonal wind shear.
(day –20), consistent with the observational findings of Maloney and Dickinson (2003). The model EKE is similarly enhanced in association with low-level anomalous westerlies (day 0). Variations in EKE between the westerly and easterly phases have similar magnitude to those observed (e.g., see Fig. 9 in Maloney and Dickinson 2003). Maloney and Dickinson (2003) suggested that variations in the background horizontal shear through its impact on barotropic conversion could help regulate synoptic eddies during MJO events. Anomalous moistening occurs just equatorward of the tilted bands of suppressed EKE on day –20 (not shown), and anomalous drying occurs poleward of these bands. Similar results, but of opposite sign, are found on day 0. These results suggest that eddies are mixing humidity across the mean humidity gradient from the tropics to the subtropics (e.g., Fig. 6). Suppression of eddies thus generates anomalous moistening on the equatorial side of the EKE minima, leading to equatorial m recharge, and anomalous drying on the equatorward side of EKE maxima, leading to m discharge.

What do the dominant eddies in the model look like? Eddy wind components and eddy vorticity \( \left( \nabla \times \vec{v}_e - \nabla \times \vec{V}_0 \right) \) are regressed against eddy vorticity at Northern Hemisphere (15°N, 150°E) and Southern Hemisphere (15°S, 160°E) reference points during November–April (Fig. 16). These reference points are near the locations of the strongest EKE anomalies at days –20 and 0 in the composites in Fig. 15. The dominant eddies have southwest–northeast (northwest–southeast) tilted structures in the Northern (Southern) Hemisphere that resemble the tropical depression (TD) type disturbances frequently observed (e.g., Lau and Lau 1992; Takayabu and Nitta 1993). The orientation of these disturbances relative to the mean shear (not shown) indicates that barotropic conversion may be an important energy source for these disturbances, as in the observations (e.g., Lau and Lau 1992), although convective diabatic heating likely also strongly contributes to the EKE budget. The Northern (Southern) Hemisphere disturbances propagate northwestward (southwestward) in time (not shown) and have comparable 2500–3000-km wavelengths to the observed TD-type disturbances (e.g., Lau and Lau 1992). Hence, tropical synoptic-scale disturbances such as TD-type
disturbances appear to contribute importantly to the recharge–discharge cycle of moist static energy during model intraseasonal oscillation events. Burpee (1974) and Reed et al. (1977) suggested that horizontal advection could explain lower-tropospheric humidity variations associated with tropical synoptic-scale waves, supporting the results shown here.

6. Conclusions and discussion

The intraseasonal moist static energy (MSE) budget is analyzed in a climate model that produces realistic eastward-propagating tropical intraseasonal wind and precipitation variability in the western Pacific Ocean. Consistent with the recharge–discharge paradigm for tropical intraseasonal variability, a buildup of column-integrated MSE occurs within low-level easterly anomalies in advance of intraseasonal precipitation, and a discharge of MSE occurs during and after precipitation when westerly anomalies occur. The strongest MSE anomalies peak in the lower troposphere and are primarily regulated by specific humidity anomalies. MSE anomalies in the model are only one-quarter to one-third of those associated with the observed MJO, although precipitation anomalies are approximately the same strength. The results suggest that the discharge of model MSE is associated with a larger moisture convergence per unit mass flux than is observed, supported by the moister mean equatorial troposphere in the model than in the observations.

The leading terms in the column-integrated intraseasonal MSE budget are horizontal advection and surface latent heat flux, where the latent heat flux is dominated by the wind-driven component. Horizontal advection causes recharge (discharge) of MSE within regions of anomalous equatorial lower-tropospheric easterly (westerly) anomalies, with the meridional component of the moisture advection dominating the MSE budget near 850 hPa. Latent heat flux anomalies oppose the MSE tendency due to horizontal advection, making the recharge and discharge of the column MSE more gradual than if horizontal advection were acting alone. This relationship has consequences for the time scale of intraseasonal variability in the model. Intra-seasonal zonal moisture advection is the largest term in the MSE budget at 700 hPa, although it appears to be less important than lower-tropospheric meridional advection for regulating the model intraseasonal variability.

Eddies dominate the intraseasonal meridional moisture advection in the model. During periods of low-level intraseasonal easterly anomalies, the eddy kinetic energy (EKE) is anomalously low due to a suppression of tropical synoptic-scale disturbances and other variability on time scales shorter than 20 days. Anomalous moistening of the equatorial lower troposphere occurs during intraseasonal easterly periods through suppression of the eddy moisture advection between the equator and poleward latitudes. During intraseasonal westerly periods, EKE is enhanced, leading to anomalous drying of the equatorial lower troposphere through meridional advection.

This analysis has been primarily focused on the western Pacific warm pool, since intraseasonal variability in the model is most realistic there. In the western Pacific, observed MJO events are in their mature stage, westerly wind bursts are most prominent, and the phase relationships among MJO-related surface fluxes, wind anomalies, and other dynamic and thermodynamics variables may differ from those in the Indian Ocean (e.g., Zhang and McPhaden 2000). Thus, it should be noted that the balance of terms in the MSE budget during the initial development of an MJO in the Indian Ocean may differ from that found in the western Pacific.

The results of this paper and others suggest a couple of hypotheses for the basic state a model must achieve for it to have success in simulating the tropical intraseasonal oscillation. First, a model must have a realistic basic-state meridional specific humidity gradient from the equator toward the poles to allow the recharge–discharge mechanism based on meridional moisture advection discussed here to operate, consistent with the results of Maloney and Hartmann (2001), who found that the basic-state tropical humidity distribution in their model strongly regulated the strength of intraseasonal variability.

As shown in Fig. 6, the equatorial troposphere is moister in the control simulation than in the weak reevaporation simulation, thus producing a stronger average equator-to-midlatitude humidity gradient. Variations in EKE on intraseasonal time scales, and thus variations in horizontal mixing across this humidity gradient, will have different impacts on the equatorial MSE budget between the two simulations. As compared to the observations, the control simulation does not have a clearly superior humidity gradient to the weak reevaporation simulation. The equator-to-midlatitude humidity gradient appears to be somewhat stronger in the observations than in the weak reevaporation simulation (Fig. 6), although the control simulation has a slightly stronger gradient than is observed. Regardless, these results suggest that aquaplanet simulations with no equator-to-pole SST gradient and weak corresponding meridional humidity gradients are unlikely to produce realistic tropical intraseasonal variability.
A second hypothesis derived from these results is that surface latent heat flux anomalies must be of sufficiently large magnitude, and of the correct phase relative to equatorial zonal wind anomalies, to moderate the effects of horizontal advection on the lower-tropospheric moisture budget. Otherwise, the MSE recharge and discharge by horizontal advection is too rapid, which would increase the eastward propagation speed and generate a shorter time scale for model intraseasonal variability. Collocation of intraseasonal zonal wind anomalies with a westerly basic-state flow dominates the equatorial latent heat flux anomalies in the model and in the observations. Thus, a climate model must have realistic equatorial lower-tropospheric westerly basic-state flow to produce a realistic simulation of the surface fluxes and the intraseasonal MSE budget. Inness and Slingo (2003) and Inness et al. (2003) found a similar basic-state dependence in a coupled general circulation model, citing the importance of basic-state westerlies for producing realistic propagation of intraseasonal anomalies into the western Pacific Ocean. Maloney and Sobel (2004) showed that wind-induced surface latent heat flux anomalies were crucial for producing realistic intraseasonal variability in the atmospheric GCM they used, with maximum latent heat flux anomalies occurring in anomalous westerly flow.

Sensitivity experiments with idealized modeling frameworks could be used to test the hypotheses presented above. Aquaplanet simulations in which the strength of the equator-to-pole SST gradient is varied could be used to test the sensitivity of intraseasonal variability to the meridional humidity gradient. Relaxation toward reference specific humidity and temperature profiles could also be used. To test the dependence of intraseasonal variability on the basic-state wind, the strength of equatorial westerlies can be varied by changing the zonal symmetry of the equatorial SST distribution.

As one last note, horizontal advection dominates the recharge and discharge of MSE in the model intraseasonal oscillation diagnosed here. However, observations indicate substantial lower-tropospheric moistening due to midlevel and shallow convection prior to MJO deep convection events (e.g., Johnson et al. 1999; Tung et al. 1999; Kiladis et al. 2005; Agudelo et al. 2006), which would contribute to the vertical advection term in the vertically integrated MSE budget (1). Although it does not appear that shallow convection is significantly contributing to moistening in the model, it is likely that shallow cumulus convection is not well simulated, and it is possible that horizontal moisture advection is playing a compensating role in the MSE budget. The vast majority of the time, the highest convective cloud tops in the western Pacific model grid cells occur in the upper troposphere (Fig. 17), also a characteristic of other general circulation models (e.g., Inness et al. 2001). A minor secondary frequency maximum occurs in the lower troposphere. Observations suggest that the population of convective clouds in the western Pacific warm pool is trimodal, characterized by shallow, congestus, and deep clouds (e.g., Johnson et al. 1999), with shallow cumulus and cumulus congestus clouds being far more prevalent than deep convection. While the comparison to the observations is not direct, since the RAS convection scheme used here is designed to simulate an ensemble of clouds detraining at different heights, Fig. 17 does suggest that deep convection is far too prevalent in the model relative to shallower clouds.

Budget analyses using modeling approaches such as the Colorado State University Multiscale Modeling Framework (e.g., Khairoutdinov and Randall 2001), in which the NCAR CAM3 convection parameterization is replaced by cloud-resolving models embedded within individual grid cells, may help to elucidate the role of shallow convection to MSE recharge in the MJO. This multiscale modeling approach appears to generate a very robust MJO (e.g., Khairoutdinov et al. 2007). Further investigations are also required.

![Fig. 17. Percent of the time that the highest convective cloud top within a grid cell occurs at a given pressure.](image-url)
using large-scale observational analysis products to elucidate the role of horizontal advection in the intra-seasonal MSE budget, including the role of transient eddies.

Acknowledgments. The author would like to thank Larissa Back and two anonymous reviewers for their insightful comments. This work was supported by the Climate and Large-Scale Dynamics Program of the National Science Foundation under Grants ATM-0832868 and ATM-0828531. The statements, findings, conclusions, and recommendations do not necessarily reflect the views of NSF.

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