An Idealized Semi-Empirical Framework for Modeling the Madden–Julian Oscillation

ADAM SOBEL

Department of Applied Physics and Applied Mathematics, Department of Earth and Environmental Sciences, and Lamont-Doherty Earth Observatory, Columbia University, New York, New York

ERIC MALONEY

Department of Atmospheric Science, Colorado State University, Fort Collins, Colorado

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ABSTRACT

The authors present a simple semi-empirical model to explore the hypothesis that the Madden–Julian oscillation can be represented as a moisture mode destabilized by surface flux and cloud–radiative feedbacks. The model is one-dimensional in longitude; the vertical and meridional structure is entirely implicit. The only prognostic variable is column water vapor W. The zonal wind field is an instantaneous diagnostic function of the precipitation field.

The linearized version of the model has only westward-propagating (relative to the mean flow) unstable modes because wind-induced surface latent heat flux anomalies occur to the west of precipitation anomalies. The maximum growth rate occurs at the wavelength at which the correlation between precipitation and surface latent heat flux is maximized. This wavelength lies in the synoptic- to planetary-scale range and is proportional to the horizontal scale associated with the assumed diagnostic wind response to precipitation anomalies.

The nonlinear version of the model has behavior that can be qualitatively different from the linear modes and is strongly influenced by horizontal advection of moisture. The nonlinear solutions are very sensitive to small shifts in the phasing of wind and precipitation. Under some circumstances nonlinear eastward-propagating disturbances emerge on a state of mean background westerlies. These disturbances have a shocklike discontinuous jump in humidity and rainfall at the leading edge; humidity decreases linearly and precipitation decreases exponentially to the west.

1. Introduction

We propose a highly idealized model of intraseasonal disturbances. The model is motivated by a desire to understand the Madden–Julian oscillation (MJO; Madden and Julian 1971). The degree to which the model in its current form is relevant to the observed MJO is unclear. We view it as a straightforward extension of certain even simpler models, in a direction of increasing fidelity to the realities of the MJO. The model makes a number of strong simplifying assumptions, motivated by observations, theory, and particularly recent numerical modeling (e.g., Maloney et al. 2010), that embody a set of hypotheses about the dynamics of the MJO. Constructing a simple model based on these hypotheses is one way of testing them; if the model cannot produce an MJO-like disturbance within a reasonable parameter range, it is an indication that one or more of the hypotheses needs revision. The hypotheses are as follows:

1) The MJO is a ''moisture mode'', meaning that it depends essentially on a prognostic humidity equation and is not analogous to any dynamical mode that occurs in a dry atmosphere. Moisture modes (sometimes called by other names) have been studied previously in many other models with varying degrees of complexity (Neelin and Yu 1994; Sobel et al. 2001; Fuchs and Raymond 2002, 2005, 2007; Sobel and Bretherton 2003; Raymond and Fuchs 2009; Sugiyama 2009a,b; Majda and Stechmann 2009; Maloney et al. 2010; Kuang 2011; Andersen and Kuang 2012). Growth of such disturbances is governed by feedbacks that increase moisture anomalies, and their propagation

Corresponding author address: Adam Sobel, Columbia University, Dept. of Applied Physics and Applied Mathematics, 500 W. 120th St., Rm. 217, New York, NY 10027.
E-mail: latex@ahs129@columbia.edu

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is governed by processes that make moisture anomalies move horizontally. Horizontal moisture advection in particular may be important (e.g., Maloney et al. 2010). According to this hypothesis, the MJO is not a Kelvin wave; Kelvin waves may play a role in its dynamics, but the MJO does not propagate by interactions between buoyancy and pressure gradients as a Kelvin wave does. It may be that the phenomenon known as the MJO consists of two dynamically different disturbances, a slower-moving one in the Indian and western Pacific basins that is distinct from convectively coupled Kelvin waves (Wheeler and Kiladis 1999; Kiladis et al. 2009) and a faster-moving one in the central and eastern Pacific and Atlantic basins that is essentially Kelvin-wave-like. We are interested in the former, non-Kelvin-wave-like one.

2) Thermodynamic feedbacks are important energy sources for the MJO. We refer specifically to feedbacks between MJO disturbances and the sources and sinks of column-integrated moist static energy, namely surface turbulent fluxes (Emanuel 1987; Neelin et al. 1987) and radiative cooling (Raymond 2000; Fuchs and Raymond 2002). Evidence that these feedbacks are important to the MJO comes from both observations and numerical model studies [see reviews by Sobel et al. (2008, 2010)]. Negative gross moist stability has also been proposed as an important contributor to the moist static energy budget of the MJO (Raymond and Fuchs 2009; Raymond et al. 2009). This is not explored directly here but could be investigated by a straightforward extension of the model.
3) Both vertical and meridional structure can be taken as implicit. Our model has only a single prognostic PDE in longitude and time. The prognostic variable is total column water vapor. We assume that the meridional and vertical structures are known and that the processes that determine them can be taken for granted. This could be argued more formally by projection on a set of basis functions, in the meridional (Majda and Khouider 2001) or vertical (e.g., Neelin and Zeng 2000). While our single prognostic variable would be formally consistent with a single basis function in the vertical, some key effects of variable structure can nonetheless be captured implicitly by extensions of the model presented here that do not require changing its basic form. For example, the effect of variable vertical structure on gross moist stability (e.g., Haertel et al. 2008) might be represented by a parameterization of the gross moist stability as a function of humidity or zonal wind.
4) Convection in the MJO is in a state of quasi-equilibrium with its forcings. We assume that the precipitation and convective heating can at any moment be taken to be an instantaneous function of the thermodynamic state.
5) Large-scale wind anomalies associated with the MJO can be taken to be a quasi-steady response to heating. That is, the wind field can be diagnosed instantaneously from the heating field at a given moment. This amounts to an assumption that the time scale for the steady response to a fixed heat source to be established is short compared to the MJO frequency. This assumption allows a strong simplification of the model. It is suggested by observational studies showing a broadly Gill (1980)-type structure to MJO wind anomalies (e.g., Chen et al. 1996) and is shown explicitly to be the case in an idealized numerical model by Sugiyama (2009b). A plausible structure of the wind response to heating is specified in this work based on the Gill model, but we view the more precise definition of this structure as a target for future study. Processes not explicitly included, such as convective momentum transport (e.g., Houze et al. 2000; Tung and Yanai 2002a,b; Lin et al. 2004; Miyakawa et al. 2012), could be included implicitly by their influence on the projection operator that relates wind to heating. The model behavior is sensitive to details of this assumed wind structure.
6) Ocean coupling is not essential. A large number of GCM studies indicate that while ocean coupling may improve the simulation of the MJO, it is not essential to the existence of the MJO (Waliser et al. 1999; Hendon 2000; Kemball-Cook et al. 2002; Inness and Slingo 2003; Zheng et al. 2004; Maloney and Sobel 2004; Grabowski 2006; Fu et al. 2007). The essential mechanisms of MJO development, maintenance, propagation, and scale selection should operate in an uncoupled context. Ocean coupling can be straightforwardly added to the present model, but we do not do so here.

In section 2 we introduce the model. In section 3 we linearize the model and show the properties of its linear normal modes. In section 4 we present a few representative numerical solutions of the fully nonlinear model, and in section 5 we present conclusions.

2. Model framework

a. Basic equations

The only prognostic equation for the atmosphere in our model is one for column-integrated water vapor $W(x, t)$:

$$\frac{dW}{dt} - M_q\Delta = E - P + k_W\frac{\partial^2 W}{\partial x^2}, \tag{1}$$

where d/dt is a material derivative following the zonal flow (discussed further below); E is surface evaporation and P precipitation; and M_q is a gross moisture stratification (Neelin 1997), which is the proportionality coefficient relating the column-integrated moisture convergence to the upper-level mass divergence Δ associated with the baroclinic flow. The last term is zonal diffusion with constant diffusivity k_W. The dry static energy equation is made diagnostic by the weak temperature gradient (WTG) approximation:

$$M_s\Delta = P - R, \tag{2}$$

where M_s is the gross dry stability and R is vertically integrated radiative cooling. We have neglected the surface sensible heat flux—a good approximation over tropical oceans—and, consistent with the assumption that horizontal temperature gradients are small, have also neglected horizontal diffusion of dry static energy. The WTG approximation need not rule out all effects of temperature variations; those that are correlated with moisture variations can be included implicitly. The dynamics of gravity and Kelvin wave propagation, on the other hand, are excluded. Adding (1) and (2) gives the moist static energy equation:

$$\frac{dW}{dt} = -\Delta M + E - R + k_W\frac{\partial^2 W}{\partial x^2}, \tag{3}$$

where $M = M_s - M_q$ is the gross moist stability (Neelin and Held 1987; Neelin 1997). Eliminating the divergence using (2) in (3) and expanding the total derivative on the left-hand side gives

$$\frac{dW}{dt} = \frac{\partial W}{\partial t} + u\frac{\partial W}{\partial x} = -\tilde{M}P + E - (1 - \tilde{M})R + k_W\frac{\partial^2 W}{\partial x^2}, \tag{4}$$

where $u(x, t)$ is an advecting zonal wind at a nominal steering level and $\tilde{M} = M/M_s$ is the "normalized gross moist stability" [defined slightly differently than in Raymond et al. (2009), in that those authors normalize by moisture convergence rather than dry static energy divergence].

b. Time and zonal mean budgets

Our model domain represents a longitudinal section through a domain with implicit latitudinal structure; we solve (4) on $0 < x < L_{\max}$, with periodic boundary conditions and $L_{\max} = 40\,000$ km. We do not assume, however, that the flow lies fully in the zonal plane with zero meridional component, and accordingly we do not require that $\Delta = -\partial u/\partial x$. Thus the mass, energy, and moisture budgets do not close in the domain integral. They are in weak temperature gradient balance with an implicit mean meridional (i.e., Hadley) circulation. In the special case in which time dependence, horizontal advection, and diffusion are all negligible, the precipitation at any x is given by the single-column local expression

$$P = \tilde{M}^{-1}[E - (1 - \tilde{M})R]. \tag{5}$$

While advection in particular is generally not negligible, (5) is nonetheless useful in understanding some basic properties of the model, as discussed further below. In the general case, integrating (4) over the domain in x gives

$$\frac{\partial}{\partial t}\int W\,dx = \int\left[-\tilde{M}P + E - (1 - \tilde{M})R - u\frac{\partial W}{\partial x} + k_W\frac{\partial^2 W}{\partial x^2}\right]dx. \tag{6}$$

In steady state, the precipitation satisfies

$$\int \tilde{M}P\,dx = \int\left[E - (1 - \tilde{M})R - u\frac{\partial W}{\partial x} + k_W\frac{\partial^2 W}{\partial x^2}\right]dx. \tag{7}$$

This equation does not give a closed relationship between domain-averaged quantities; for example, even if E and R are specified and $\tilde{M}$ is taken constant, computation of the advection term requires knowledge of the longitudinal structure of u and W. Because Δ and u are not uniquely related, the explicit horizontal advective transport can have a nonzero domain average and it is not helpful to phrase the model in flux form. Equation (7) simply shows the nature of the zonal mean WTG balance. While there are implicit latitudinal transports, at this stage we do not explicitly model latitudinal transports associated with meridional gradients—there are no advective or diffusive terms involving derivatives of W with respect to latitude. Such terms may under some circumstances be quantitatively nonnegligible, and could be added in parameterized form.

Taking the zonal wind to advect the entire column water vapor, as in (4), may overestimate the effect of horizontal advection. Horizontal advection of moisture has been shown to be greatest in the lower free troposphere in observations (Benedict and Randall 2007) and simulations (e.g., Maloney 2009; Maloney et al. 2010), but the zonal wind tends to change sign with height in MJO events while the moisture gradient generally does

not. In the quasi-equilibrium tropical circulation model (QTCM), for example (Neelin and Zeng 2000; Zeng et al. 2000), the horizontal advection term is multiplied by a coefficient of order 0.3, computed from explicit projection on the assumed vertical structures, to capture this. On the other hand, the lower-tropospheric water vapor is presumably the more important for controlling convection while the upper-tropospheric water vapor is expected to vary more as a passive response to convection (e.g., Sherwood 1999; Sobel et al. 2004), so one could argue that for the model to capture that feedback it should weight lower-level advection more heavily.

c. Model physics

Our convective closure models P as a function of W, $P = P(W)$. The relationship between precipitation and column water vapor is the subject of both theoretical and observational work (Raymond 2000; Bretherton et al. 2004; Peters and Neelin 2006; Neelin et al. 2009; Muller et al. 2009), and it has been suggested that the simulation of the MJO in global models is sensitive to it (Benedict and Randall 2009; Zhu et al. 2009). Here, we choose $P(W)$ according to the observational study of Bretherton et al. (2004):

$$P = P_R \exp(a_d F), \tag{8}$$

where $a_d = 15.6$, $P_R = 8.22 \times 10^{-5}$ mm day^{-1}, and F is the saturation fraction

$$F = \frac{W}{W_{\max}}, \tag{9}$$

with $W_{\max}$ being the saturation column water vapor. [The expression (8) is equivalent to that used by Bretherton et al. (2004) with the equivalence $P_R = \exp(-a_d r_d)$, where $r_d = 0.603$.] Here $W_{\max}$ is chosen to be 70 mm, consistent with typical warm pool values.

We parameterize atmospheric radiative cooling by a clear-sky term, taken constant, plus a cloud–radiative feedback term taken proportional to precipitation, with the additional requirement that the net effect of radiation must be to cool, rather than heat, the atmosphere:

$$R = \max(R_0 - rP, 0). \tag{10}$$

We expect that this radiative feedback will be destabilizing and assist in the development and maintenance of intraseasonal disturbances (e.g., Raymond 2001; Sobel and Gildor 2003; Bony and Emanuel 2005; Zurovac-Jevtic et al. 2006; Sobel et al. 2008, 2010; Andersen and Kuang 2012; Landu and Maloney 2011). Bretherton and Sobel (2002) estimated $r \approx 0.15$–0.2 from observations, depending on the dataset used. Lin and Mapes (2004) performed a more thorough study with a broader range of datasets and estimated $r \approx 0.1$–0.15. Our control value is $r = 0.1$, at the low end of this range; sensitivity of the model to r is discussed below.

Surface evaporation is parameterized as a function of steering level wind speed:

$$E = E_0 + C_u|u|. \tag{11}$$

The dependence on wind speed is motivated by results from the simulation of Maloney et al. (2010). Figure 1 shows daily mean values of surface latent heat flux and 850-hPa zonal wind from the simulation described in that study; values shown in Fig. 1 are taken only from the warm pool region where the simulated MJO disturbances are most active. While the latent heat flux in these simulations does depend on an air–sea humidity difference according to a standard bulk formula, to first order the wind speed at 850 hPa appears to contain sufficient information to compute the flux. Parameterizing E as a function of $|u|$ but not W allows us to avoid representing the surface air humidity as a function of column water vapor, something that is difficult to do well without an explicit boundary layer. Plots analogous to Fig. 1 made from observed data or reanalyses show greater scatter than does Fig. 1 but have similar regression slopes (not shown). The linear regression line in Fig. 1 implies parameter choices of $E_0 = 99.9$ W m^{-2} and $C_u = 7.9$ W m^{-3} s; we use $E_0 = 100$ W m^{-2} and $C_u = 7.5$ W m^{-3} s as our control values.

The normalized gross moist stability is taken as constant and positive in this study. Variable $\tilde{M}$ can be included and most obviously could be parameterized as a function of W (such dependence would be required to write a closed moisture budget in flux form in the zonal plane, but as discussed above our system is open, with implicit meridional transport).

With the radiative parameterization (10), if P remains smaller than $R_0 r^{-1}$, the precipitation in steady state for the special case of negligible horizontal advection and diffusion, (5), can be written

$$\tilde{M}_{\text{eff}} P = E - (1 - \tilde{M}) R_0, \tag{12}$$

where $\tilde{M}_{\text{eff}} = \tilde{M}(1 + r) - r$ is a normalized "effective gross moist stability" including radiative feedbacks (e.g., Bretherton and Sobel 2002; Su and Neelin 2002). For our control parameters, $M_{\text{eff}} = 0.01$, while E and $(1 - \tilde{M})R_0$ are typically close in value. Thus the steady-state precipitation in this idealized case results from a delicate balance in the moist static energy budget in which the

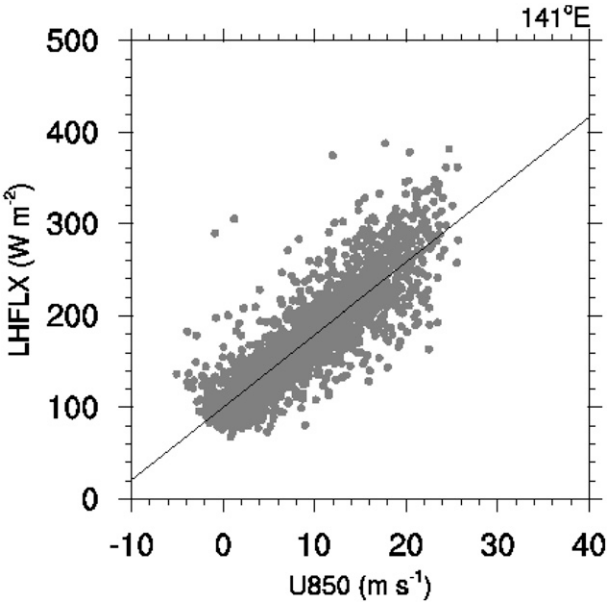

FIG. 1. Daily surface latent heat flux (y axis) vs 850-hPa zonal wind (x axis), with both quantities averaged from 0° to 20°S at the longitude 141°E, from the aquaplanet simulation of Maloney et al. (2010). The linear regression line is also shown.

forcing and effective gross moist stability are both small. Our actual solutions (in the nonlinear regime) are in general strongly influenced by horizontal advection, so that in statistically steady state they obey (7) in the domain average rather than the simpler (12), but M_{eff} and $E - (1 - \bar{M})R_0$ remain important quantities for the control of the mean state. Because E is strongly controlled by the wind and thus the amplitude of propagating disturbances, the disturbances and mean state are in general coupled whether horizontal advection is important or not. Tuning is required to keep the mean precipitation close to values observed in the earth's tropics. While this is not in principle a desirable feature, it appears to be broadly consistent with observations showing that the gross moist stability (and by implication the effective gross moist stability as well) is not clearly distinguishable from zero in the rainiest parts of the tropics (Back and Bretherton 2006).

d. *Zonal wind as a diagnostic function of precipitation*

We impose a constant background wind U such that $u(x,t) = U + \hat{u}(x,t)$. We do this even in the nonlinear model, although the hat is used in section 3 to denote linear perturbations in other quantities as well as u. The constant background wind reflects the influence of implied Hadley and Walker circulations unresolved by our model. The perturbation $\hat{u}(x,t)$ need not have zero mean in general. If Δ were proportional to $\partial u/\partial x$, as in idealized "mock Walker" models (e.g., Bretherton and Sobel 2002), we could simply integrate Δ to find $\hat{u}$. Instead we assume that $\hat{u}$ is both divergent and rotational, but that it can be computed instantaneously from the total atmospheric heating, $P - R$, via a projection operator, similar to a Green's function:

$$\hat{u}(x,t) = \int G(x\,|\,x')[P(x',t) - R(x',t)]\,dx'. \quad (13)$$

We can determine G empirically or theoretically. Here, we derive G from the solutions of Gill (1980) to the linear shallow water system on an equatorial beta plane subject to a localized mass source forcing and Rayleigh damping on the mass and momentum fields, but we also allow an ad hoc zonal shift of the wind response δ relative to the heating:

$$G(x\,|\,x') = -Ae^{-[x-(x'+\delta)]/L}, \quad x > x' + \delta, \quad (14)$$

$$G(x\,|\,x') = 3Ae^{3[x-(x'+\delta)]/L}, \quad x < x' + \delta, \quad (15)$$

with A and L being constants. For $\delta = 0$, this can be derived from Gill's model in the equatorially symmetric case if the heating has the meridional structure assumed by Gill but is a delta function in longitude. The length scale L can be interpreted as the group velocity of free Kelvin waves divided by the Rayleigh damping rate (e.g., Sarachik and Cane 2010, 157–162). The factor of 3 in (15) expresses the fact that the group velocity of Kelvin waves is 3 times that of long Rossby waves. Use of a Gill solution for an off-equatorial forcing strengthens the westerly component relative to the easterly (if the near-equatorial wind is still taken to be the relevant one) but does not significantly shift the relative longitudinal position of the peak westerlies. Sensitivity of the model behavior to this change, and other plausible variations in G, will be addressed in future work.

In this study we choose L to be 1500 km. This is consistent with an equivalent depth of 40 m (Kelvin wave speed = 20 m s^{-1}) if the dissipation time scale for the wave response to heating is 1 day, characteristic of boundary layer drag (arguably appropriate for surface wind, though too small for free-tropospheric wind). The value of 40 m is slightly higher than that found by Wheeler and Kiladis (1999) for convectively coupled waves but considerably smaller than would be appropriate for dry equatorial waves. While there is a temperature and wind response that propagates at around 40 m s^{-1} (Bantzer and Wallace 1996), the convective signal does not propagate this quickly. Our value of 20 m s^{-1} is

consistent with the speed found in observations by Maloney and Shaman (2008) for the observed propagation of the MJO into the Atlantic, and that found in GCM simulations for the Kelvin wave response to the switching on of a localized SST anomaly (Maloney and Sobel 2007). Sensitivity of our model to L is clear in the linear calculations below; sensitivity to it in the nonlinear system is deferred to future work.

The parameter δ, introduced in (14) and (15), is a distance by which the response G is shifted relative to where it would be relative to that obtained from the Gill solution. Such departures could result from a number of factors not present in the highly idealized linear shallow water system including more complex vertical structure, convective momentum transport (Houze et al. 2000; Tung and Yanai 2002a,b; Lin et al. 2004; Miyakawa et al. 2012), meridional momentum transport by synoptic-scale disturbances (Biello et al. 2007; Showman and Polvani 2010), or nonlinearity in the direct flow response to MJO-scale heating (Gill and Phlips 1986). These factors might well change the structure of G as well as its phase; such structural changes may be of interest in future studies but are not considered here. We do not argue that any particular value of δ is appropriate to represent reality, but we find the sensitivity of the model to this parameter interesting. We do require δ to be relatively small compared to L; the largest value used in this study is $4L/15$, or $0.27L$. The amplitude A is chosen so that precipitation anomalies of planetary spatial scale have wind anomalies whose magnitude in meters per second is comparable to that of the precipitation anomalies in millimeters per day, as occurs in the simulations of Maloney et al. (2010).

e. *Summary remarks on model construction*

In essence, the present model can be viewed as an extension of a simple WTG single-column model (e.g., Sobel and Gildor 2003; see also Maloney and Sobel 2004) to incorporate one horizontal dimension (longitude). As in such models, the present model's essential dynamics include those relating moisture and convection under the weak temperature gradient approximation (2), the convective closure (8), and the simple cloud-radiative feedback (10). These processes (in less heavily parameterized form) have been found to cause spontaneous self-aggregation of convection in large-domain cloud-resolving simulations (Bretherton et al. 2005), behavior that is captured by the existence of multiple equilibria (convecting and nonconvecting) in WTG single-column (Sobel et al. 2007) or small-domain cloud-resolving models (Sessions et al. 2010). The key new additions here result from the addition of horizontal structure, with horizontal wind modeled as a quasi-steady response to precipitation. This allows horizontal advection and large-scale wind–evaporation [wind-induced surface heat exchange (WISHE)] feedbacks to be explicitly included.

The lack of explicit, self-consistent representation of meridional and vertical structure (basis functions, vertical layers, etc.) results from a conscious choice. We hypothesize that a model of the form above with suitable physics may focus our attention usefully on understanding the roles of the different processes as captured by key bulk parameters—for example, the gross moist stability, response of wind to heating via L and δ, etc.—and the magnitudes of the terms necessary to generate MJO-like disturbances at some desired level of realism. We postpone any attempt at determining how the full three-dimensional structure of the primitive equations should best be truncated to achieve a self-consistent representation of those processes in terms of explicit but simple vertical and horizontal structures. It is in this sense that we refer to the model as "semi-empirical." At the same time there is nothing particularly new in the essential thermodynamic equation of the model; essentially the same terms and tunable parameters result, in one form or another, from a variety of vertical truncation schemes used in theoretical tropical meteorology, including a two-level or two-layer model (e.g., Neelin et al. 1987; Wang 1988) or projection on vertical basis functions derived from either a dry vertical structure equation (e.g., Stevens and Lindzen 1978) or quasi-equilibrium convective constraints on the temperature profile (e.g., Emanuel 1987; Neelin 1997). Any novel aspects here result from the use of a prognostic moisture equation but diagnostic (WTG) temperature equation—the opposite choice being more typical historically—and the diagnostic computation of the wind field.

3. Linear analysis

We linearize the model about a background state W_0,

$$W = W_0 + \hat{W}(x,t), \tag{16}$$

where the hat indicates a small perturbation; the more standard prime is reserved here for the dummy spatial variable x' used in the spatial projection to obtain $\hat{u}$. The background state is also assumed to have a uniform zonal wind, so that (as in the nonlinear model) the total wind is $U + \hat{u}(x,t)$. The linearized model is

$$\frac{\partial \hat{W}}{\partial t} + U\frac{\partial \hat{W}}{\partial x} = -\tilde{M}\hat{P} + \hat{E} - (1 - \tilde{M})\hat{R} + k_w\frac{\partial^2 \hat{W}}{\partial x^2}. \tag{17}$$

We linearize our convective parameterization about the state $W = W_0$, $P = P_0 = P_R \exp[a_d(W_0/W_{\max})]$, thus obtaining

$$\hat{P} = \frac{\hat{W}}{\tau_c}, \tag{18}$$

where

$$\tau_c = \frac{W_{\max}}{a_d P_0} = (a_d P_R)^{-1} W_{\max} \exp[-a_d(W_0/W_{\max})]. \tag{19}$$

Our linearized radiative perturbations are then

$$\hat{R} = -r\tau_c^{-1}\hat{W}, \tag{20}$$

while, if we assume westerly mean winds, our perturbation latent heat flux is

$$\hat{E} = C_u \hat{u}. \tag{21}$$

Now assuming sinusoidal perturbations,

$$\hat{W} = \mathcal{W} e^{i(kx-ct)}, \tag{22}$$

with $\mathcal{W}$ being a complex amplitude and c a (potentially) complex phase speed, and substituting we obtain

$$ik(U - c)\hat{W} = -(\tau_c^{-1}\tilde{M}_{\text{eff}} + k_w k^2)\hat{W} + C_u\hat{u}. \tag{23}$$

With a projection function of the form (14) and (15) and using (18) and (20), (13) can be written

$$\begin{aligned}\hat{u}(x,t) &= \int G(x\,|\,x')\hat{P}(x',t)(1 + r)\,dx' \\ &= (1 + r)\tau_c^{-1}\int G(x\,|\,x')\hat{W}(x',t)\,dx'.\end{aligned} \tag{24}$$

For sinusoidal disturbances as described by (22), $\hat{u}$ can be found analytically:

$$\hat{u} = \Gamma(k)\hat{P} = \Gamma(k)\frac{\hat{W}}{\tau_c}, \tag{25}$$

where

$$\Gamma(k) = (1 + r)\left\{\frac{4Ae^{-ik\delta}\left[\dfrac{2k^2}{L} + i\left(\dfrac{3k}{L^2} + k^3\right)\right]}{(9/L^2 + k^2)(1/L^2 + k^2)}\right\}. \tag{26}$$

From this it follows that the phase angle α by which $\hat{u}$ lags $\hat{P}$ is

$$\alpha = \tan^{-1}\left(\frac{3}{2kL} + \frac{kL}{2}\right) - k\delta. \tag{27}$$

For $\delta = 0$, the zonal wind and precipitation fields are thus in quadrature ($\alpha = \pi/2$) when the dimensionless ratio of length scales kL is either zero (long wavelength $2\pi/k$ compared to wind decay scale L) or infinite (short wavelength). For $\delta = 0$ the wind is most nearly in phase with precipitation at the value $kL = \sqrt{3}$ at which $\alpha = \tan^{-1}\sqrt{3} = \pi/3$; when δ is small (compared to $\pi/3k$) but nonzero, the minimum phase lag is $\alpha = \pi/3 - k\delta$. The fact that α differs at all from $\pi/2$ for any finite L is a consequence of the asymmetry of G, resulting from the different decay scales of the Rossby and Kelvin components. For an antisymmetric projection function in which the decay scales are the same to the east and west of the forcing—as would result (for example) in an analog to the Gill model for planetary vorticity gradient $\beta = 0$, in which only inertia–gravity waves could carry the linear response to forcing, and would have the same wave speeds to the east or west—u and P will be in quadrature for any value of that decay scale.

The importance of the phase lag between wind and precipitation is apparent if we take the linear atmospheric equation (17), multiply by $\hat{W}$, substitute $\hat{P} = \tau_c^{-1}\hat{W}$ and $\hat{R} = -r\hat{P}$, and integrate over the domain, assumed periodic, to obtain an equation for the variance of W':

$$\begin{aligned}\frac{\partial}{\partial t}\int \frac{1}{2}\hat{W}^2\,dx &= -\tau_c^{-1}\tilde{M}_{\text{eff}}\int \hat{W}^2\,dx \\ &\quad + \int \hat{E}\hat{W}\,dx - k_w\int\left(\frac{\partial \hat{W}}{\partial x}\right)^2 dx.\end{aligned} \tag{28}$$

While (28) assumes a periodic domain, it does not assume sinusoidal perturbations. The last term on the right-hand side is negative definite, and the first is as well if M_{eff} is positive. In that case amplitude growth can result only from the second term, which is the covariance of surface latent heat flux and column water vapor. Under our assumptions this wind–evaporation feedback term will be positive in general for a westerly mean state and perturbations whose maximum winds are westerly and lag the precipitation with respect to longitude. These conditions are fundamentally different from those under which WISHE is destabilizing for a convectively coupled Kelvin wave destabilized by wind–evaporation feedback in an easterly flow (Emanuel 1987; Neelin et al. 1987).

In this model, as discussed above, the existence of a spatial correlation between u and P results from the east–west asymmetry in G. Since E is assumed proportional to

u while P is proportional to W, (28) this implies that the destabilizing effect of WISHE in this model also results from that asymmetry in G. That destabilization is thus a consequence of the symmetry-breaking effect of the planetary vorticity gradient β.

For the linear system, using (25), (23) becomes

$$ik(U - c)\hat{W} = \tau_c^{-1}[r - \tilde{M}(1 + r) + C_u\Gamma - \tau_c k_w k^2]\hat{W}$$

or (29)

$$c = U + \frac{\tilde{M}_{\text{eff}} - C_u\Gamma + \tau_c k_w k^2}{ik\tau_c}. \quad (30)$$

The phase speed c and wind–precipitation proportionality factor Γ are both complex in general; all other coefficients in (29) are real. The disturbances propagate at a real phase speed $\text{Re}(c)$, while $\text{Im}(c)k$ is the linear growth rate. These two quantities are shown in Fig. 2 for the parameters shown in Table 1, except that k_w is set to zero; the value used in the nonlinear calculations is sufficiently small that including it does not significantly change the results (not shown). In these calculations the mean state column water vapor is $W_0 = 45$ mm, resulting in a convective time scale $\tau_c = 2.4$ days. The growth rate and phase speed are both inversely proportional to τ_c, as can be seen from (30). The growth rate has its maximum at zonal wavenumber 7.3, or a wavelength of 5440 km, consistent with the value $kL = \sqrt{3}$ as discussed above. A larger value of the parameter L would of course lead to a larger wavelength at the maximum growth rate.

Figure 3 shows the same calculations for a range of values in the key parameters r and δ. We see that increasing either parameter increases the growth rate, while the shapes of the curves remain qualitatively similar as these two parameters are varied.

Figure 4 shows the sensitivity to the inverse convective time scale τ_c^{-1}, which grows exponentially with W_0 according to (19). The growth rate increases rapidly with τ_c^{-1}, as does the (westward) phase speed. Logarithmic scales are used on the y axis of both plots. For large τ_c^{-1} (small τ_c; the smallest value shown is 0.03 days) the growth rates are so large that nonlinearity would rapidly become important, and the phase speeds are also so large as to render key assumptions of the model (e.g., quasi-stationarity of wind response to heating) invalid.

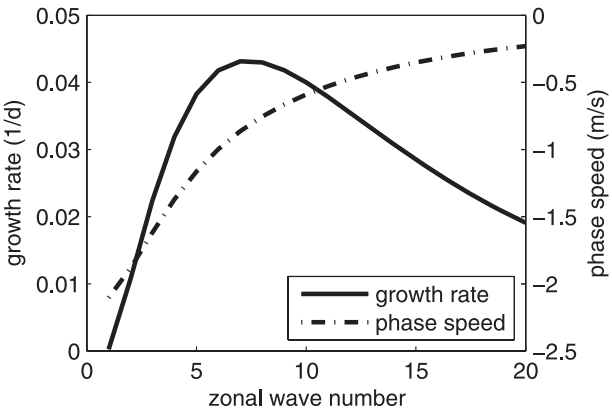

FIG. 2. Growth rate (solid, day^{-1}) and phase speed (dot-dashed, m s^{-1}). For the linear model with a background value $W_0 = 45$ mm and a convective time scale $\tau_c = 2.4$ days. Here the wind shift $\delta = 0$ km and the background wind is assumed westerly, but the phase speed shown is that relative to the mean wind, rather than relative to the surface.

In these linear solutions it is clear why the wind–evaporation feedback (or WISHE) is destabilizing, and how this mechanism controls scale selection in a configuration similar to that of the real MJO where the mean low-level wind is westerly and the largest low-level winds are westerlies occurring to the west of precipitation maxima. At the same time these solutions do not appear explain the propagation dynamics of the MJO at all. They predict only westward phase propagation relative to the basic flow, with phase speeds comparable to or greater than that of typical warm pool low-level mean westerlies so that ground-relative westward propagation, or at best stationarity, is implied. It is possible that these solutions

TABLE 1. Control values of model parameters.

Parameter	Value	Definition
R_0	4.8 mm day^{-1}	Clear-sky radiative cooling
$W_{\max}$	70 mm	Saturation column water vapor
P_R, a_d	8.22×10^{-5} mm day^{-1}, 15.6	Constants in convective scheme
L	1500 km	Length scale for wind response to precipitation
A	$0.8/L$ (m s^{-1})(mm day^{-1})$^{-1}$ m^{-1}	Magnitude of wind response to precipitation
k	2604 m^2 s^{-1}	Diffusivity for moisture
U	5 m s^{-1}	Background low-level zonal wind
$\tilde{M}$	0.1	Normalized gross moist stability (when fixed)
r	0.1	Cloud-radiative feedback parameter
E_0, C_u	100 W m^{-2}, 7.5 W m^{-3} s	Surface Latent heat flux for $u = 0$, latent heat flux change per $\lvert u \rvert$
W_0, τ_c	45 mm, 2.4 day	Background W and convective time scale for linear model

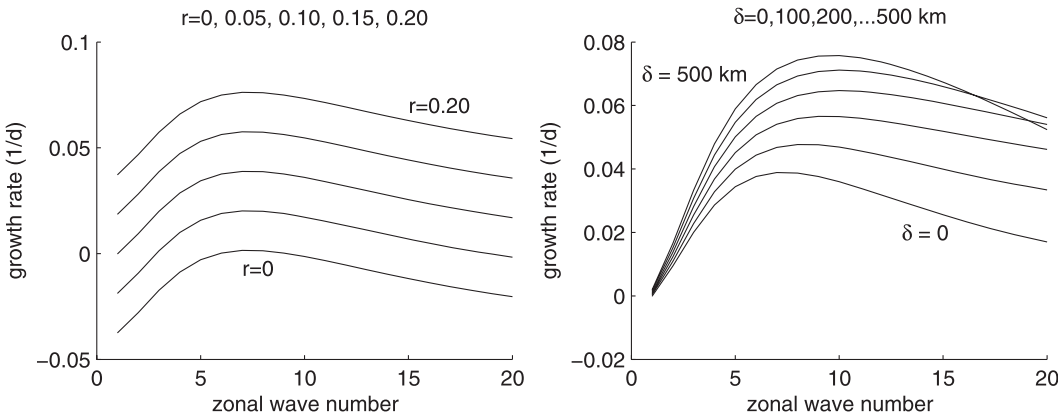

FIG. 3. Growth rate (day^{-1}) for the linear model with the same parameters as in Fig. 2, except that (left) the cloud–radiative feedback parameter r is varied from 0 to 0.2 in increments of 0.05, and (right) the wind shift parameter δ is varied from 0 to 500 km in increments of 100 km. Greater growth rate at fixed wavenumber (for $k < 15$) occurs at greater r and greater δ; the curves with the maximum and minimum values of r are labeled on the plots.

may have some relevance to the dynamics of westward-propagating intraseasonal variability (e.g., Murakami 1980; Kemball-Cook and Wang 2001). Some of this is associated with convectively coupled equatorial Rossby waves (e.g., Kiladis and Wheeler 1995; Wheeler and Kiladis 1999), which may at times interact with the MJO (Roundy and Frank 2004).

4. Nonlinear numerical solutions

a. Numerical model configuration and parameters

We solve the nonlinear system numerically on a periodic domain of length 40 000 km, with 1000 grid points so the horizontal grid spacing is 40 km. We use a first-order upwind scheme for horizontal advection and a leapfrog time stepping procedure with a Robert–Asselin filter and a time step of 0.001 days. Simulations are initialized somewhat arbitrarily with the initial condition

$$W = W_0 + \Delta W \sin\left(\frac{\pi x}{L_m}\right),$$

where here $W_0 = 50$ mm, $\Delta W = 2$ mm, and $L_m = 40\,000$ km is the domain size. It is found through experimentation that the qualitative results of interest are not sensitive to the initial conditions, although our exploration of the initial conditions is not at all exhaustive.

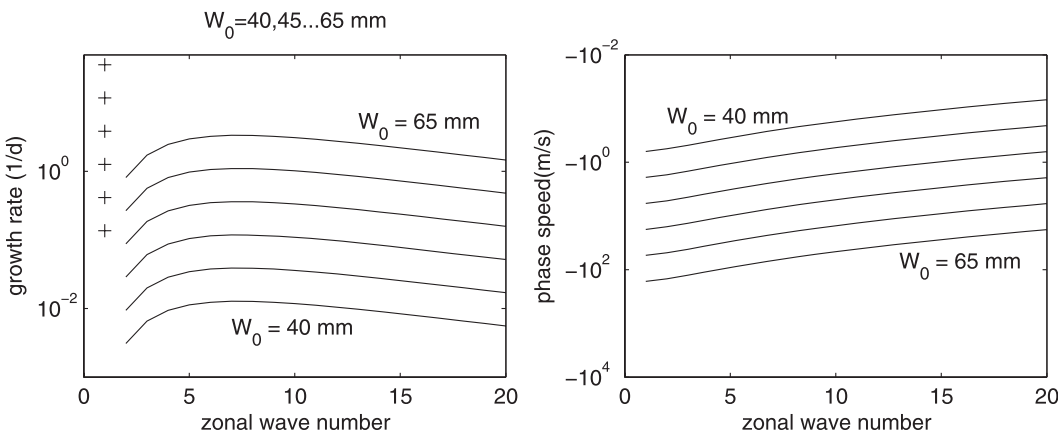

FIG. 4. (left) Growth rate (day^{-1}) and (right) phase speed (m s^{-1}) for the uncoupled linear model with the same parameters as in Fig. 2, except that the background column water vapor W_0 is varied from 40 to 65 mm in increments of 5 mm; the saturation value is 70 mm. The inverse of the resulting linearized convective time scale τ_c^{-1} is shown by the plus signs on the left; τ_c itself varies from 7.3 to 0.03 days. Smaller τ_c (larger W_0) corresponds to larger growth rate and larger westward (more negative) phase speed; the curves with the maximum and minimum values of W_0 are labeled on the plots.

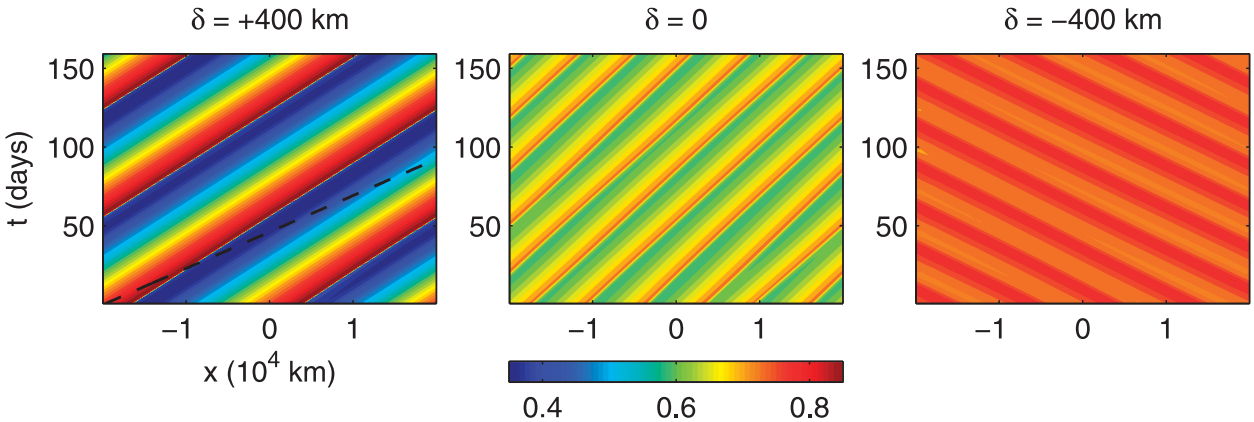

FIG. 5. Hovmöller plots of saturation fraction W/W_{max} in integration with δ = (left) 400, (middle) 0, and (right) −400 km; other parameters are as in Table 1. The x axes show longitudinal distance (10^4 km) while the y axes show time (days). The first day shown is day 321 of each integration, with the time values shown on the y axes corresponding to time since then. The dashed line in the left panel shows the basic flow speed, 5 m s^{-1}.

We impose a background westerly wind of 5 m s^{-1}. This influences both horizontal advection and surface fluxes. We imagine that our domain, despite having an extent approximately equal to the circumference of the earth, consists solely of a ''warm pool'' or region of relatively high sea surface temperature, while elsewhere (at both other latitudes and longitudes, though neither is explicitly included) the SST is lower. Thus precipitation is focused on the warm pool, and the quasi-steady response includes low-level westerly winds, as is the case in the tropical Indian and western Pacific Oceans on earth.

b. Results

Figure 5 shows Hovmöller plots of saturation fraction from calculations with parameters shown in Table 1 and δ = +400, 0, and −400 km. Results are shown for periods of 160 days; in each case the period shown starts on day 321 of each integration (labeled as day 1 in the figure). By this time, initial transient features have decayed, leaving nearly periodic disturbances that are close to steady in their respective comoving reference frames. The first two show eastward propagation, while the third shows westward propagation. In the δ = 400 km solution in the left panel, the eastward propagation speed is close to but less than (i.e., easterly relative to) that of the background wind, 5 m s^{-1}; the latter value is indicated by the dashed line in the figure. The dominant spatial structures have wavenumber 2 for δ = 400 km, and wavenumbers 6 and 4 in the latter two calculations respectively (as is made more apparent in the figures below).

Figures 6–8 show snapshots of precipitation and perturbation zonal wind (left panels) and water vapor path

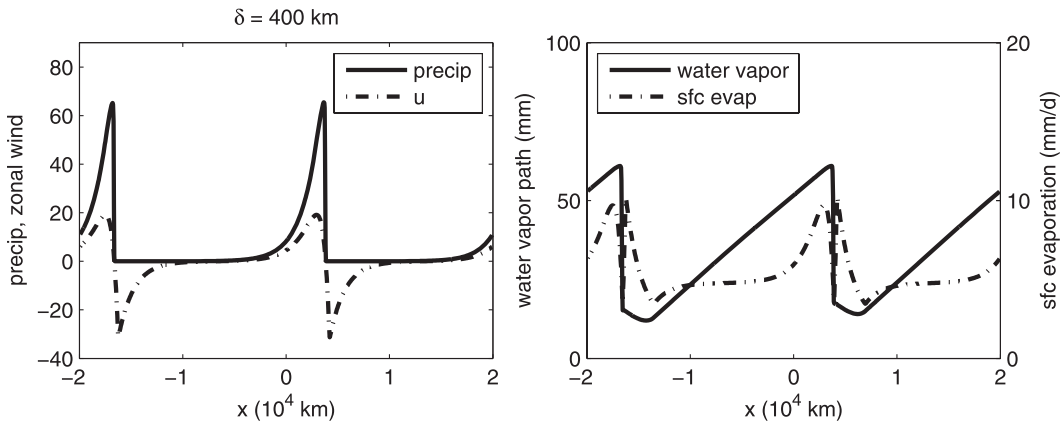

FIG. 6. Results from integration with δ = 400 km; other parameters are as in Table 1. Panels show snapshots at day 321 of (left) perturbation zonal wind (m s^{-1}) and precipitation (mm day^{-1}) and (right) water vapor path (mm) and surface latent heat flux (mm day^{-1}).

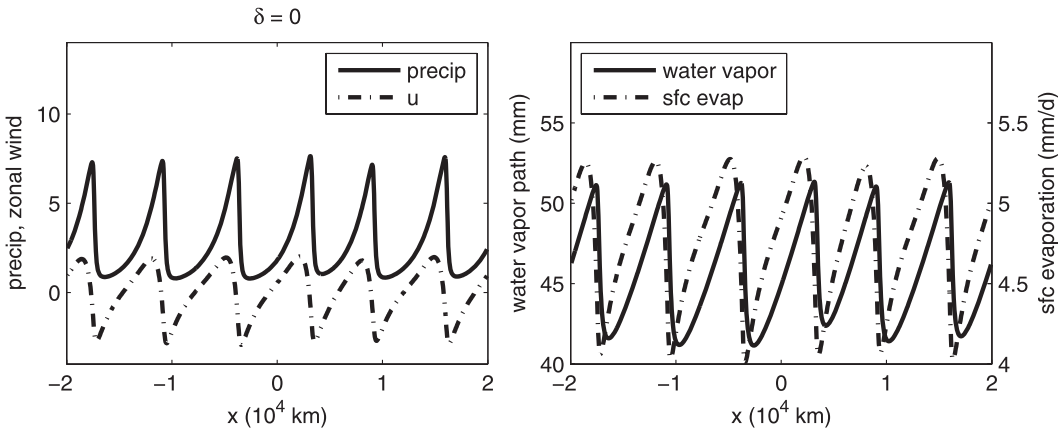

FIG. 7. As in Fig. 6, but for $\delta = 0$ km.

and surface evaporation (right panels) on day 321 for the same three calculations shown in Fig. 5, in the same order. Note that the scales on the vertical axes are not the same for corresponding fields in the different figures. In Fig. 6 we see two isolated peaks in precipitation, with roughly exponential increases on the western sides and then steplike decreases back down to a zero background on the eastern sides. This structure [including the number of peaks (two)] appears at least qualitatively independent of initial conditions, for a range of initial perturbations we have tried (not shown). The zonal wind has westerlies roughly in phase with precipitation with strong easterlies ahead in the dry regions. The water vapor path in this solution has a nearly sawtooth wave pattern, with a linear increase followed by a step decrease. The latent heat flux has two peaks separated by a sharp minimum at the point where the wind perturbation switches from easterly to westerly.

Figure 7 shows the same fields as in Fig. 6 but for $\delta = 0$. Recall from Fig. 5 that there is still some eastward propagation in this solution (also recall that there is a background eastward wind of 5 m s^{-1}) but the perturbations are smaller in both amplitude and spatial scale than in Fig. 6. The precipitation features resemble the sharp structures in Fig. 6, but the maxima are closer together so that the large regions of zero precipitation in Fig. 6 are absent. Close inspection reveals that E lags W by more than in the previous figure, as expected. These differences are still more pronounced in Fig. 8, with $\delta = -400$ km. Recall from Fig. 5 that the disturbances in this solution move rapidly westward, despite the eastward basic flow. The structures in all fields are smoother and closer to sinusoidal than those in Figs. 6 and 7. The precipitation variations, from maximum to minimum, are roughly a factor of 3 smaller than those in Fig. 6, while the variations in W are smaller by a factor of 5. The lack of proportionality is due to the exponential convective closure; the solution with $\delta = -400$ km is considerably moister in the mean, resulting in a stronger response to small W variations (smaller τ_c in the linearized model).

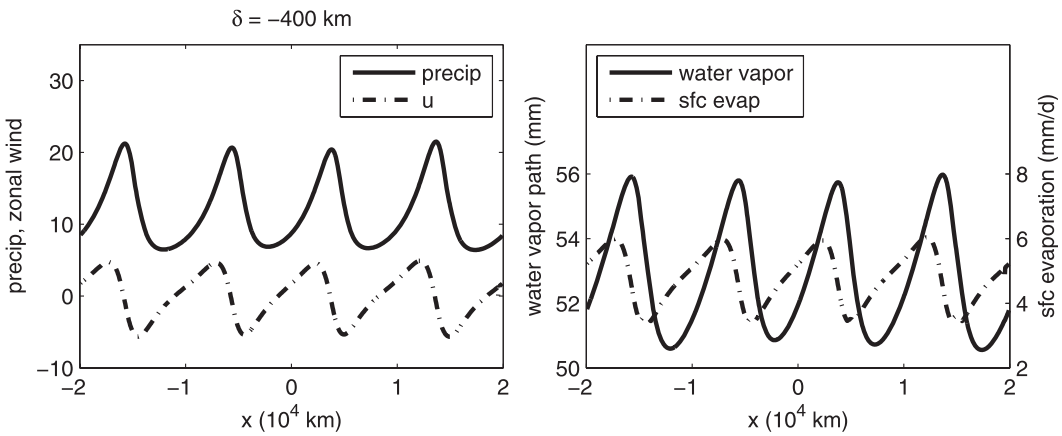

FIG. 8. As in Fig. 6, but for $\delta = -400$ km.

The highly nonlinear solution for positive δ has no analog in the linear system. While simply shifting G as we have done is ad hoc, the results demonstrate a strong sensitivity of this system to the phase relationship between zonal wind and precipitation. The existence of these nonlinear solutions depends on both horizontal advection by the perturbation winds (an inherently nonlinear effect) as well as by the mean flow, and on the shifting of E forward so that it is more nearly in phase with W and P (an effect that is present in the linearized model as well). When either of these effects is disabled, the nonlinear mode shown in Fig. 6 is strongly weakened or otherwise altered (not shown). An exploration of other parameter choices (not shown) suggests that advection of moisture by the perturbation flow is the more important of the two effects in generating this mode. This is consistent with shock dynamics as found in other systems where an active scalar generates a flow that advects itself in one dimension. While precipitation fronts can also occur in a quasi-linear system without nonlinear horizontal advection (Frierson et al. 2004; Stechmann and Majda 2006; Pauluis et al. 2008), here the role of nonlinear advection is apparent in the correspondence of the front location with that of strong zonal confluence. The effect of this confluence can be seen in the initial ~20–30 days (not shown) of the integration shown in the left panel in Fig. 5, where eastward- and westward-moving moist regions on either side of $x = 0$ move toward each other before colliding near $x = 0$ around 50 days and coalescing into the narrower structure evident in Fig. 5.

While the eastward propagation and planetary horizontal scale of the nonlinear mode might be viewed as encouraging, the discontinuity at the leading edge is not MJO-like. Observations show that the transition from suppressed to active phase is as gradual as that from active back to suppressed, if not more so (Kemball-Cook and Weare 2001; Kiladis et al. 2005; Benedict and Randall 2007). This could indicate that the processes that lead to the slow deepening of convection at the leading edge (e.g., Mapes et al. 2006) need to be represented better in this model. On the other hand, it could simply be an indication that the nonlinear mode is an artifact of excessively strong horizontal advection and is irrelevant to the real atmosphere.

The model is nonlinear, and many parameter choices affect the mean state as well as the existence or properties of time-dependent perturbations to it. As an example, Fig. 9 shows an example in which all parameters are the same as in Fig. 6—in particular, $\delta = 400$ km—but the cloud–radiative feedback parameter r has been increased from 0.1 to 0.15, rendering the effective gross moist stability slightly negative, $\tilde{M}_{\mathrm{eff}} = -0.035$. The mean state is dramatically changed, with strong rainfall

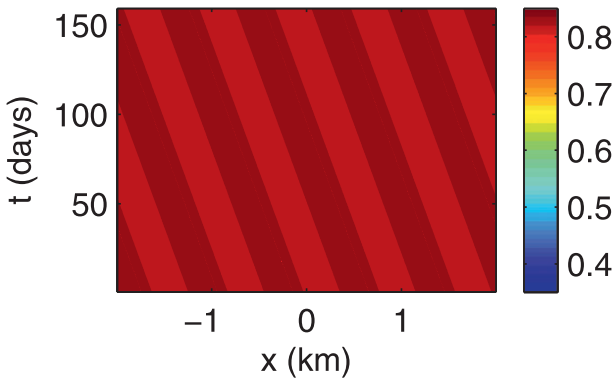

FIG. 9. As in Fig. 5, but with $\delta = 400$ km and the radiative feedback parameter $r = 0.15$.

everywhere and consequently a much greater zonal mean rain rate than in Fig. 6. There are still self-sustained oscillations, but their properties are quite different from those in Fig. 6. The disturbances move predominantly westward, and the mean state is quite humid (the same color scale is used as in Fig. 5). Of the solutions with $r = 0.1$ shown in Fig. 5 the solution looks more similar to that for $\delta = -400$ km than that for $\delta = +400$ km, although the latter is the value of δ used in Fig. 9.

The solution in Fig. 9 illustrates the general property that changes in the perturbations are generally accompanied by changes in the mean state (although the latter are usually smaller than those shown here if $\tilde{M}_{\mathrm{eff}}$ does not change sign). It is not obvious whether this property is an advantage or disadvantage for an idealized MJO model. In full-physics general circulation models, changes in simulated MJO amplitude are also accompanied by mean state changes (Kim et al. 2011) (though not ones as large as here), and it is possible that interactions between the seasonal mean circulation and MJO disturbances are important in the real climate as well. If desired, various devices can be used to constrain the mean state in the present model, but these will in general also influence the disturbance dynamics.

We have performed a wide range of sensitivity studies. Solutions tend to resemble, qualitatively, one of those in Figs. 5–9, but a number of parameters influence which type of solution emerges. Besides δ, r, and $\tilde{M}$, the saturation column water vapor and the wind length scale are also important. We do not present these studies here, as they are not yet particularly informative. The key result here is the emergence of the strongly nonlinear solution for positive δ in some parameter regimes. It is clear that this solution does not emerge for parameters that cause the background state to be very humid and rainy, as in Figs. 8 or 9; such states tend to feature westward-propagating wavelike disturbances whose amplitude in

precipitation is modest compared to the mean precipitation (although not necessarily small in absolute terms). The properties of the strongly nonlinear solution appear qualitatively similar in all cases in which it occurs, although there is some variation in the amplitude and spacing between disturbances.

5. Conclusions

We have constructed a one-dimensional semi-empirical framework that we hope may be useful for developing MJO theories. The framework incorporates a number of strong assumptions but is not based explicitly on a set of predefined vertical or meridional structures. This has the disadvantage that internal consistency is not enforced to the degree that it could be, but the advantage that the roles of key parameters or functions can be examined independently and could in principle be determined based on observations or numerical modeling studies. While the particular implementation presented here can, under some circumstances, produce planetary-scale, eastward-propagating disturbances when aided by an eastward basic flow, we do not claim at this point that these capture the essential dynamics of the MJO. We believe they capture hypotheses worth exploring, and we use them as examples to demonstrate some sensitivities and basic properties of the framework.

In the linear regime, unstable normal modes exist but are all westward-propagating relative to the mean flow. Their growth rates and phase speeds are strongly sensitive to the convective time scale, which is a strong function of the mean-state humidity. The spatial scale of the fastest-growing mode is set by the scale of the quasi-stationary response of the wind to heating, which in reality depends on the effective stratification and damping that act on the forced-dissipative waves that determine that response.

In some reasonable parameter regimes, if the projection function G that determines the zonal wind field as a function of precipitation is taken from the equatorially symmetric Gill model but is shifted a few hundred kilometers to the east relative to precipitation, a nonlinear mode emerges that propagates eastward at nearly the speed of the background eastward wind. The disturbances in this mode have a nearly sawtooth structure in humidity and precipitation maxima that decay exponentially westward and with a step function on the eastward side.

Extensions of the framework currently under exploration include: variations in the deterministic convective parameterization $P(W)$; addition of a stochastic component (either additive or multiplicative) to the convective parameterization; variable gross moist stability, parameterized as a function of column water vapor W or perhaps zonal wind u; zonal variation in the basic state; explicit (parameterized) representation of meridional advection, including eddy transports that may vary with MJO phase (Maloney 2009; Andersen and Kuang 2012); and coupling to a mixed layer ocean.

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