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Surface Fluxes and Tropical Intraseasonal Variability: a Reassessment

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The authors argue that interactive feedbacks involving surface moist enthalpy fluxes, both turbulent and radiative, are important to the dynamics of tropical intraseasonal variability. Evidence in favor of this hypothesis includes the observed spatial distribution of intraseasonal variance in precipitation and outgoing longwave radiation, the observed relationship between intraseasonal latent heat flux and precipitation anomalies in regions where intraseasonal variability is strong, and sensitivity experiments performed with a small number of general circulation and idealized models. The authors argue that it would be useful to assess the importance of surface fluxes to intraseasonal variability in a larger number of comprehensive numerical models.

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1. Introduction

Theoretical understanding of the mechanisms responsible for tropical intraseasonal variability is limited. There is a large and increasing number of interesting and plausible ideas in the literature, but no agreement on which of them, if any, is correct. The Madden-Julian oscillation (MJO) in particular is arguably the most significant mode of atmospheric variability at any sub-decadal time scale whose essential features — its existence, energetics, spatial and temporal scales — remain so unsatisfactorily explained.

In this study, we use the phrase ''MJO'' to refer to the eastward-propagating 30-60 day mode which is dominant in southern hemisphere summer. This mode remains present in northern hemisphere summer, but northern summer also features a northward-propagating mode, manifest in northward-propagating rain bands over the Indian subcontinent and adjacent oceans. We refer to these two modes collectively as ''tropical intraseasonal variability'' and treat them to some degree as one phenomenon. We recognize that the eastward- and northward-propagating modes have some significant differences, but argue here that there may be fundamental similarities in their energetics.

General circulation models (GCMs) simulate tropical intraseasonal variability with varying degrees of fidelity. A couple of recent intercomparison studies (Lin et al. 2006, Zhang et al. 2006) show that even the best models still have significant flaws in their MJO simulations. At the same time, some members of the current generation of models show considerable improvement over previous generations. While improving GCM simulations is sometimes cited as a motivation for theoretical research into the MJO, no clear relationship exists between the fidelity of GCM simulations and the state of theoretical understanding, perhaps because of the very different levels of complexity of GCMs as compared to the idealized models used by theorists. It is not clear that recent improvements in MJO simulation owe anything to theoretical understanding of the mechanism of the MJO. GCM improvements in MJO simulation often seem to be accidental by-products of broader model development efforts, or results of trial-and-error tuning, or perhaps tuning guided by broader principles not specific to the MJO. For example, any model change which tends to inhibit deep convection tends to increase variability on a range of timescales, including the intraseasonal timescale (Tokioka et al. 1988; Wang and Schlesinger 1999; Zhang and Mu 2005; Lin et al. 2008). This constitutes improvement for models in which intraseasonal variability is too weak, as is the case with many.

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Any steps we can take to narrow the range of mechanisms which are considered possible explanations for tropical intraseasonal variability would be valuable, particularly if some mechanisms can be eliminated convincingly enough to focus the attention of the community on evaluating the remainder. We argue that models of the MJO in which variations in net surface fluxes of moist enthalpy[1] are important (which need not be coupled ocean-atmosphere models, as surface fluxes can vary strongly even if sea surface temperature is fixed) are more likely to be correct than those in which such variations are unimportant, while recognizing that the pioneering studies which first proposed this idea (Emanuel 1987; Neelin et al. 1987) have proven incorrect in the details of the specific linear theory they used to express it. Those details are all inessential to the hypothesis that variations in surface enthalpy fluxes are important to tropical intraseasonal variability.

Three primary pieces of evidence support our argument:

1. Intraseasonal variances in precipitation and outgoing longwave radiation are observed to be larger over ocean than land. This is true in both hemispheres, even in regions where the climatological mean precipitation is larger over land than ocean. Since variations in net surface enthalpy flux must be small over land on intraseasonal time scales, the land-sea contrast in intraseasonal variance is consistent with a role for that flux in generating the variance.
2. Over the oceanic regions of largest intraseasonal variability, intraseasonal variations in net surface enthalpy flux and precipitation are correlated in space and time.
3. In several general circulation models (as well as idealized models) surface enthalpy fluxes are demonstrably important to the simulated intraseasonal variability. Experiments with these models suggest that the role of surface fluxes is larger in those models whose MJO simulation is better.

While most of this evidence is not new, the GCM results in particular have started to become more convincing, partly because the state of the art in GCM simulations of intraseasonal variability has improved (e.g., Zhang et al. 2006). This, in conjunction with continuing observational and theoretical work, has led us to the position that the case for the importance of surface enthalpy fluxes to observed intraseasonal variability has become stronger than it was a decade ago, and deserves to be systematically re-examined.

[1] The total surface moist enthalpy flux is the sum of the surface sensible and latent heat fluxes and the net surface radiative flux. Moist enthalpy is defined, e.g., by Emanuel (1994). The term ''total surface heat flux'', as commonly used in climate science, would be equivalent, but we mildly prefer ''enthalpy'' to ''heat'' following arguments in Bohren and Albrecht (1998); ''surface energy flux'' is also approximately correct in practice but strictly includes kinetic energy as well. For brevity we sometimes simply say ''surface fluxes''.

In the following section, we provide a highly selective review of some results and ideas, mostly from the theoretical and modeling literature, which are relevant to the hypothesis that surface enthalpy fluxes are important to the dynamics of tropical intraseasonal variability. In section 3, we review some observational results which are consistent with this hypothesis. This is followed in section 4 by a discussion of GCM results, including a presentation of new results from the NOAA GFDL AM2 model which show that the hypothesis appears to have merit in that model, consistent with the results of Maloney and Sobel (2004) who used a version of the NCAR model. In section 5, we discuss the implications of these results and propose avenues for further research.

2. Theory and modeling: An unbalanced review

2.1. Southern summer intraseasonal variability (MJO): Models of Emanuel (1987) and Neelin et al. (1987)

We do not attempt to provide a comprehensive or balanced review of MJO theory. This is beyond our intended scope, and several recent reviews cover the ground well (Wang 2005; Zhang 2005; Waliser 2006). Instead we focus on theories in which variations in surface enthalpy fluxes figure prominently.

Approximately simultaneously, Emanuel (1987) and Neelin et al. (1987) proposed that air-sea interaction could destabilize a moist Kelvin wave, leading to intraseasonal variability in the tropics. The arguments involved linear analysis of idealized moist models in which the temperature structure is assumed to be represented by a first baroclinic mode, and convection is controlled by quasi-equilibrium principles. Essentially, convection acts in these systems to eliminate some local measure of stability of the column to deep convection. The atmosphere is adjusted by convection towards a reference value of the stability measure [e.g., convective available potential energy (CAPE)], which for present purposes may be assumed zero. The dynamics of such models is discussed in more detail in a number of reviews (e.g., Emanuel et al. 1994; Neelin 1997, Stevens et al. 1997; Smith 1997; Arakawa 2004; Emanuel 2007). Neelin et al. (1987) also performed numerical simulations with a general circulation model (GCM).

We refer to models of the type discussed by Emanuel et al. (1994) as '''quasi-equilibrium' (QE) models''. In its most general sense, the term QE refers to a broader category, including all models in which the convection is assumed close to statistical equilibrium with its forcings. Traditional QE models incorporate additional assumptions, in particular that of a pure first baroclinic mode vertical structure, which can be relaxed without relaxing the assumption of QE per se. In models assuming a first baroclinic mode structure as well as QE, the interaction of deep convection with large-scale dynamics alone does not generate unstable large-scale modes. In the simplest such models convection is assumed to respond instantaneously to large-scale forcing (which can come from large-scale dynamics, radiation, or surface fluxes) so as to remove all local instability completely; Emanuel et al. (1994) called this ''strict quasi-equilibrium''. In this case, the interaction of convection and large-scale dynamics reduces the effective stratification and thus the phase speed of convectively coupled gravity and Kelvin waves, but does not stabilize or destabilize them (in the sense of linear instability of a dynamical mode, as opposed to the local instability which causes convection). If instead the convection is assumed to relax the CAPE (or other measure of column instability) towards its reference value with a finite timescale, then the interaction damps disturbances, a phenomenon known as ''moist convective damping''. Disturbances in these models cannot become linearly unstable through the interaction of convection with large-scale dynamics alone, but only through feedbacks involving processes which can act as sources of moist static energy (or moist entropy) to the column. The two most important such processes are surface turbulent fluxes and radiative cooling.

The requirement for moist static energy sources to be involved in any linear instability is an interesting feature of first baroclinic mode quasi-equilibrium models. In extratropical atmospheric dynamics it has proved extremely useful to separate dry adiabatic dynamical mechanisms (e.g., Hoskins et al. 1985) from those in which ''diabatic'' processes (defined for a dry working fluid with phase changes of water considered external), which break the conservation of potential temperature and potential vorticity, are fundamentally involved. It is clear that dry adiabatic dynamics are inadequate to describe many important aspects of the tropical atmospheric circulation and its variability, but the relative importance of moist adiabatic dynamics — as opposed to dynamics in which moist diabatic processes (those which break the conservation of moist static energy and moist entropy) are critical — remains unresolved. The analogy to extratropical dynamics, and the overall centrality of quasi-conserved variables in all of physics, suggests that it is fruitful to ascertain the relative importance of moist adiabatic and diabatic processes to intraseasonal variability. This is a separate and more fundamental question than that regarding the validity of first baroclinic mode QE models, since the latter make a number of additional restrictive assumptions. Nonetheless those models provide a useful starting point for discussion since they make a clear prediction on the relevance of diabatic processes, as well as being both relatively tractable and based on principles that are at root physically reasonable (convection acts to eliminate instability) even though some of their simplifying assumptions may be too strong for some purposes.

In the models of Emanuel (1987) and Neelin et al. (1987), Kelvin waves are destabilized by the interaction of a convectively coupled wave with surface flux perturbations induced by the wave's surface wind perturbations. This interaction was called ''wind-evaporation feedback'' by Neelin et al. (1987) and ''wind-induced surface heat exchange (WISHE)'' by Emanuel (1987). For an eastward moving Kelvin wave in a westward mean flow, a positive surface wind speed anomaly occurs a quarter wavelength ahead of the location where the positive precipitation and vertical velocity anomalies would be in the absence of surface flux anomalies, but in phase with the temperature anomaly. Under the strict quasi-equilibrium assumption, the convection responds immediately to surface flux anomalies, so the surface flux anomaly causes the heating anomaly to shift eastward, putting it partly in phase with the temperature anomaly and destabilizing the wave.

Key features of this linear theory for intraseasonal variability are that the waves must occur in an easterly mean surface flow, that the winds under the convective phase of the disturbance are easterly, and that the intraseasonal disturbances are Kelvin waves. All of these features have been shown to be inconsistent with observations. It was immediately recognized that the strongest MJO events occur in regions of mean westerlies (Wang 1988; Emanuel 1988; Neelin 1988). It was then shown that the active phases, featuring enhanced precipitation, occur in surface westerlies (e.g., Kiladis et al. 1994; Zhang and McPhaden 2000). Wheeler and Kiladis (1999) then showed that ''convectively coupled'' Kelvin waves do exist, but that their spectral signatures are quite distinct from that of the MJO, indicating that the two are different phenomena. These observations showed that the linear models of Emanuel (1987) and Neelin et al. (1987) are, in their specifics, incorrect as explanations of the MJO.

The observations are not, however, inconsistent with the general notion that surface flux anomalies may be important to the dynamics of the MJO, but only with the specific linear models proposed by Emanuel (1987) and Neelin et al. (1987). If the disturbance is something other than a linear Kelvin wave, the requirements for mean easterly flow and net easterly flow in regions of active convection no longer apply. Some studies with nonlinear models have identified such ''nonlinear WISHE'' as being important in simulated MJO-like disturbances (Raymond 2001; Maloney and Sobel 2004). In general, there is not a simple or straightforward relationship between the formulation of a given model (whether a full-physics GCM or an idealized model) and the importance of surface flux feedbacks. For example, while WISHE was first proposed in first baroclinic mode quasi-equilibrium linear models, there is no a priori reason it cannot occur in models with more complex vertical structure, different physical parameterizations, and full nonlinearity. Whether WISHE or cloud-radiative feedbacks are active in any particular model is often most easily determined by disabling these mechanisms and examining the resulting changes in the model solution at intraseasonal timescales.

A misperception that sometimes arises is that the enhancement of precipitation by surface fluxes in a model in which WISHE is active operates directly through the moisture budget — that is, that a perturbation in surface latent heat flux results in an equal perturbation in rainfall, simply because the extra water vapor put into the atmosphere by the extra flux precipitates out locally. If this were the case, the observation that precipitation anomalies in the MJO (as in many other tropical circulations) tend to greatly exceed surface evaporation anomalies would seem to contradict WISHE. This is incorrect. A positive surface flux anomaly in QE models enhances precipitation through its effect on buoyancy, not through simple local moisture recycling, and in those models (as in observations) it is generally true that the lion's share of anomalous precipitation results from moisture convergence. The moisture budget by itself is not particularly useful in discriminating between different theories for the dynamics of the MJO.

In the two decades since the publication of Emanuel (1987) and Neelin et al. (1987), much more work has been done with idealized moist models which aim to explain either the MJO, other aspects of tropical intraseasonal variability (such as the northward-propagating mode found in northern hemisphere summer, discussed further below), or other parts of the convectively coupled wave spectrum. These studies have broadened and deepened our understanding of the spectrum of possible mechanisms that can occur in large-scale geophysical flows with embedded deep convection. In the case of the MJO, however, none of them has been broadly accepted as providing a satisfactory explanation of the essential mechanisms (Zhang 2005; Wang 2005; Waliser 2006).

A number of recent studies have developed idealized models which incorporate effects not included in earlier studies. Many of these include a second baroclinic mode in addition to the first (Mapes 2000; Majda and Shefter 2001a; Khouider and Majda 2006a,b; Kuang 2008), and some include non-equilibrium effects in their convective closures. These additional effects can under some circumstances render disturbances linearly unstable purely through the interaction of convection with large-scale dynamics, without feedbacks via surface enthalpy fluxes. Other recent studies focus on the role of upscale energy transfer from synoptic- or mesoscales to the planetary-scale intraseasonal disturbances (Biello and Majda 2005, 2006; Biello et al. 2007). As with previous studies, these continue to broaden and deepen our understanding of potentially relevant mechanisms, but have not yet led to a coalescence of agreement on their centrality to observed intraseasonal variability.

The lack of broad agreement on the mechanism of the MJO does not necessarily indicate that previous studies are not correct to some degree. It may instead reflect the fact that a number of very different models predict the occurrence of intraseasonal oscillations which are comparable in the degree to which they resemble the observed ones, so that distinguishing between them is difficult; or that simulated intraseasonal oscillations (in models at all levels of complexity) are very sensitive to the representation of deep convection, a difficult and controversial problem from many perspectives. Regardless, this lack of agreement is a fact, and its existence at this point should lead us to consider whether there is something we can do beyond what we have been doing to resolve it. Further development of idealized models is surely warranted, and may yet lead to a breakthrough which will be broadly recognized as one. We argue that it is also worth treating the assessment of available theoretical ideas against available evidence as an important problem in its own right, and spending more effort on it than we, the community of researchers interested in this problem, have done. The motivation for doing this increases further as new modeling and observational resources become available. We emphasize this problem here preferentially over more detailed discussion of recent theoretical developments, interesting and promising as those may be.

2.2. Northern summer intraseasonal variability

In northern summer, intraseasonal variability modulates the Asian and western Pacific monsoons. Spectra of atmospheric variability exhibit two significant peaks in the intraseasonal range: one at 10-20 days and one at 30-60 days (Goswami 2005). The 10-20-day mode is characterized by convective disturbances which propagate from the western Pacific warm pool and the maritime continent towards the northern Bay of Bengal and South Asia. These disturbances have been associated with equatorial Rossby waves deviated northward by the mean monsoon flow (Chatterjee and Goswami 2004). The 30-60-day mode is characterized by the northward propagation of approximately zonally-oriented rain bands from $5^\circ S$ to $25^\circ N$ (Wang et al. 2006). This northward propagation is sometimes accompanied by eastward propagation (Wang and Rui 1990; Lawrence and Webster 2002). Nevertheless, the northward propagating mode appears to be an independent regional mode of variability, rather than simply a local response in the South Asian region to the eastward-propagating disturbances (Jiang and Li 2005), though this is still controversial in some quarters (e.g., Sperber and Annamalai 2008). We focus here on this northward-propagating mode, assuming that the eastward-propagating mode is essentially similar to the southern summer MJO.

Given the nearly zonal orientation of the rain bands and their nearly meridional direction of propagation, a number of studies have assumed that longitudinal structure is inessential to the dynamics of this mode, and modeled it axisymmetrically (Webster and Chou 1980; Goswami and Shukla 1984; Gadgil and Srinivasan 1990; Nanjundiah et al. 1992; Srinivasan et al. 1993; Jiang et al. 2004; Drbohlav and Wang 2005; Bellon and Sobel 2008a,b). These studies have obtained linearly unstable northward propagating modes which resemble the observed one to varying degrees. In earlier studies, land-atmosphere interaction was proposed as crucial to the northward propagation (Webster and Chou 1980; Webster 1983). However, northward propagating modes were also later obtained in aquaplanet simulations

(Goswami and Shukla 1984; Nanjundiah et al. 1992). The northward propagation has been attributed in several recent studies to dynamical mechanisms that involve low-level convergence north of the propagating rain band. This convergence is caused by Ekman pumping under a maximum of barotropic vorticity which itself leads the maximum convection (Jiang et al. 2004; Goswami 2005; Bellon and Sobel 2008b; Bellon and Sobel 2008a). The mechanisms explaining the generation of this barotropic vorticity maximum are still debated (Jiang et al. 2004; Drbohlav and Wang 2005; Bellon and Srinivasan 2006; Bellon and Sobel 2008a).

The question of what destabilizes the mode is distinct from that of what causes its propagation. In the model of Bellon and Sobel (2008a,b), interactive surface fluxes were found to be important to the instability of the northward propagating mode. They used the quasi-equilibrium model developed by Sobel and Neelin (2006), which has a barotropic mode and prognostic boundary layer in addition to a first baroclinic mode in the free troposphere. Because of this more complex vertical structure, the set of possible dynamical mechanisms in this model is broader than that in the pure first baroclinic mode QE models. It is possible for linear instability to occur in this model without surface flux feedbacks. Nonetheless, Bellon and Sobel (2008a,b) found that WISHE is critical to the linear instability of the northward-propagating mode in the parameter regime which appears most justified based on observations. As usual with idealized models, one can easily challenge various details of this model (which has some similarities to earlier ones (e.g. Jiang et al. 2004) as well as some differences). The results of Bellon and Sobel (2008a,b) just show that it is possible to construct a plausible model of the northward-propagating mode of intraseasonal variability — one based on physics that is within the broad envelope of what is commonly found in idealized models of tropical atmospheric dynamics, and also broadly consistent with observations — in which surface flux feedbacks are essential.

2.3. The near-equivalence of surface fluxes and radiation in quasi-equilibrium

The primary radiative effects of the high clouds associated with deep convection are a cooling of the surface due to reflection and absorption of shortwave radiation and a warming of the atmosphere due to the greenhouse effect in the longwave and the absorption of shortwave. To the extent that these effects have similar magnitudes, so that they cancel at the top of the atmosphere, they lead to a cooling of the ocean and equal warming of the atmosphere. This is equivalent to a surface enthalpy flux, as far as the vertically integrated moist static energy budget of the atmosphere is concerned. QE theory provides a useful context in which to frame this equivalence.

If the vertical structure of the atmospheric flow is assumed fixed (for example, a first baroclinic mode), and if we assume steady state and neglect horizontal advection, the budget of moist static energy requires that the large-scale vertical motion, or net vertical mass flux, is proportional to the net convergence of the vertical flux of moist static energy into the tropospheric column (e.g., Neelin and Held 1987; Raymond 2000; Neelin 1997; Neelin 2007; Sobel 2007). The latter is the sum of the net turbulent latent and sensible surface heat fluxes plus the vertically integrated radiative heating of the troposphere (or minus the radiative cooling).

The proportionality factor which relates the moist static energy (or moist entropy) flux to the mass flux is known as the gross moist stability (GMS), following Neelin and Held (1987). There is no very good theory for the value of the GMS, though some observational estimates have been made (Yu et al. 1998; Back and Bretherton 2006). Raymond et al. (2009) review recent thinking on the mechanincs of the GMS. The first baroclinic mode assumption is restrictive, perhaps even qualitatively misleading in some circumstances (e.g., Sobel 2007), but no better idea of comparable simplicity has yet appeared. In general, the GMS need not be a constant or a simple function of the temperature and humidity profiles alone (as in first baroclinic mode QE theory), because it is quite sensitive to the vertical profile of the divergent circulation (Sobel 2007). Since the latter can vary dynamically on a range of space and time scales, the GMS can as well. In simulations in a GCM with simplified physics (Frierson 2007b) the GMS is strongly influenced by properties of the convective parameterization (Frierson 2007a).

For our immediate purpose, what matters most is that GMS be positive on average on intraseasonal time scales, so that increases in net vertical moist static energy flux convergence into the column lead (with a time lag that is either negligible or at least short by comparison to the intraseasonal timescale; storage on timescales of a few days does not significantly complicate the argument) to increases in vertical mass flux, which in turn imply increases in deep convection. This is a weaker constraint than usually assumed in QE theory, though the difference is one of degree rather than kind. Even the positivity of the gross moist stability is questionable in observations, particularly in the eastern Pacific ITCZ (Back and Bretherton 2006), but it appears to be a reasonable assumption in the Indian and western Pacific regions for the time mean. In some models, the transient occurrence of negative GMS appears to be important to the dynamics of the simulated MJO (Raymond and Fuchs 2009), though even there enhanced surface fluxes appear to be associated with enhanced precipitation.

We assume that the difference in the cloud field between convectively active and suppressed precipitation regimes consists predominantly of the presence vs. absence of high clouds. Satellite observations have shown that, in the mean, these clouds produce perturbations in the net radiative energy flux at the top of the atmosphere which are small compared to their largely cancelling shortwave and longwave components (Ramanathan et al. 1989; Harrison et al. 1990; Hartmann et al. 2001). Lin and Mapes (2004) found that this cancellation is less close on intraseasonal time scales, with MJO-related shortwave anomalies being larger than longwave ones by as much as 30%. This is a significant difference, but still the cancellation substantially exceeds the remainder. The implication is that any anomalous radiative heating of the atmosphere due to these clouds, whether occurring in the longwave or shortwave bands, is approximately compensated by anomalous radiative cooling of the ocean. In the vertically integrated moist static energy budget, cloud-radiative heating anomalies due to deep convection are essentially similar to convectively induced perturbations to turbulent surface heat fluxes, as both amount to a net transfer of enthalpy from ocean to atmosphere in a convectively active phase. When convection is active, there is a net decrease of radiative energy flux into the ocean, accompanied by a significantly smaller change in the top-of-atmosphere balance. Thus we use the phrases ''surface fluxes'' or ''surface flux feedbacks'' to include radiative cooling feedbacks.

2.4. Ocean coupling

A substantial body of work over the last decade or so argues that intraseasonal SST variability is not only driven by the atmosphere, through intraseasonal variations in surface enthalpy fluxes, but that SST variability also influences the atmosphere through the influence of SST anomalies on column stability and deep convection. To the extent that these feedbacks are significant, intraseasonal variability is coupled. Most GCM studies addressing this in the context of the MJO have shown some enhancement of the simulated variability in experiments with atmospheric models coupled to a mixed-layer ocean models, as compared to models with fixed SST (Waliser et al. 1999; Kemball-Cook et al. 2002; Zheng et al. 2004; Fu et al. 2007), although at least two studies found no enhancement (Hendon 2000; Grabowski 2006) and and others found small enhancements (e.g., Maloney and Sobel 2004) or mixed results, with differences in the mean climate between coupled and uncoupled runs complicating the interpretation (Inness and Slingo 2003). This evidence suggests that the MJO is enhanced by coupling, but is not fundamentally dependent on coupling for its existence. In virtually all models tested in this way, a simulated MJO is present to some degree without coupling.

Observations suggest that coupling has a qualitatively similar impact on intraseasonal variability of the Asian monsoon in northern hemisphere summer, including northward-propagating rainbands and SST variability in the Arabian sea and Bay of Bengal (Vecchi and Harrison 2002; Wang et al. 2006; Roxy and Tanimoto 2007). In GCM studies, ocean coupling enhances northward-propagating intraseasonal variability in the Indian Ocean to varying degrees (e.g. Zheng et al. 2004, Seo et al. 2007, Fu et al. 2007, Fu and Wang 2004, Kemball-Cook et al. 2002). One recent study with an idealized axisymmetric model suggests that the SST variability is largely passive, being forced by the atmosphere but having only a modest impact on the atmospheric mode (Bellon et al. 2008).

The question of the importance of ocean coupling is related to but not the same as that of the importance of surface fluxes to the dynamics of intraseasonal variability. If ocean coupling is important, surface fluxes must be involved, since only through those fluxes can the ocean influence the atmosphere. The converse is not true: an important role for surface fluxes does not necessarily imply that coupling is important. Surface flux feedbacks can operate in models which assume fixed SST. Such models do not satisfy a surface energy budget, but their surface fluxes can still vary interactively and influence the atmosphere. Coupling can either amplify or damp intraseasonal variability, depending on the phasing of the SST anomalies relative to anomalies in atmospheric variables. For example, Shinoda et al. (1998) found that observed SST anomalies slightly reduced the amplitude of MJO-related surface latent heat fluxes compared to what they would have been for fixed SST.

Our interest here is in the role of surface fluxes in the dynamics of atmospheric intraseasonal variability. Ocean coupling, while also arguably important, is secondary in this discussion. However, as discussed next, the nature of the underlying surface is important to the extent that it must have sufficiently large heat capacity to allow substantial fluctuations in the net surface enthalpy flux on the intraseasonal time scale.

2.5. Single column dynamics

2.5.1. Single column dynamics inferred from observations

The MJO is commonly defined as having large spatial scales. However, plots of intraseasonal variance in quantities related to deep convection also show relatively small-scale features as described in section 3. These smaller-scale features appear to be related to the nature of the underlying surface, and thus to be a result of the local interaction of that surface with the atmosphere. A simple framework within which to grasp these local interactions may be the idealized dynamics of a single column, consisting of the atmosphere and ocean in a relatively small horizontal area.

The single-column view is taken in the observational study of Waliser (1996), who showed in a composite analysis that an oceanic ''hot spot'', defined as a region of at least 1×10^6 km^2 in which the sea surface temperature (SST) exceeds 29.75° for at least a month, typically appears after period of calm surface winds and clear skies. Anomalously strong surface winds and enhanced high cloudiness develop after the time of peak SST. Associated surface latent heat flux and shortwave radiative flux anomalies lead to the decay of the hot spot.

The conceptual model resulting from Waliser's study, as well as similar ones articulated by subsequent studies (Fasullo and Webster 1999; Stephens et al. 2004), describe coupled oscillations occurring in a single column. They leave open to what extent the oscillation can be self-contained in a single region, as opposed to being fundamentally driven by the passage of large-scale disturbances. Observations suggest that the latter is a better description of the MJO (e.g., Hendon and Glick 1997; Zhang and Hendon 1997), as well as the northward-propagating Asian monsoon mode, since they have large-scale spatiotemporal structure and propagation within which regional-scale features are embedded. Nonetheless, the single-column view is useful for understanding some aspects of the controls on deep convection, and may be particularly relevant to understanding the smaller-scale regional features shown below.

2.5.2. A simple coupled single-column quasi-equilibrium model

Sobel and Gildor (2003, SG03) presented an explicit single column dynamical model of a simple atmosphere coupled to a slab mixed-layer ocean of constant depth. This model was designed to capture the behavior described by the observational studies described in section i. above. Their model incorporates the standard assumptions of first baroclinic mode QE theory, as expressed in the "quasi-equilibrium tropical circulation model" (QTCM) formulation (Neelin and Zeng 2000; Zeng et al. 2000), plus a few additional assumptions. The most important additional assumption is that the local temperature profile — that is, not only the vertical structure of the temperature field, but also its value — is fixed. In this "weak temperature gradient" (WTG) approximation (e.g., Held and Hoskins 1985; Neelin and Held 1987; Mapes and Houze 1995; Zeng and Neelin 1999; Sobel and Bretherton 2000; Sobel et al. 2001; Majda and Klein 2003) large-scale vertical motion is assumed to occur as needed in order for adiabatic cooling to balance diabatic heating, and the local temperature profile is held close to that of surrounding regions by a process which is essentially geostrophic adjustment with a small Coriolis parameter (Bretherton and Smolarkiewicz 1989). This parameterization of large-scale dynamics allows the precipitation to vary strongly in response to variations in SST, surface turbulent fluxes, and radiative cooling. In the absence of a large-scale circulation — for example, in radiative-convective equilibrium — large precipitation variations cannot occur, because any changes in convective heating have to be balanced by changes in radiative cooling, which cannot become too large. By employing the WTG approximation rather than an assumption of no large-scale circulation (as in, for example, Hu and Randall 1994, 1995), the SG03 model allows physically plausible, if highly parameterized, interactions between convection, radiation and large-scale dynamics to occur in a single column.

SG03 also assumed that deep convective clouds induced radiative perturbations which reduced the longwave cooling of the atmosphere and shortwave warming of the ocean surface in equal measure, so that the cloud-radiative perturbations play a role very similar to that of surface flux perturbations, as discussed above. SG03 very crudely represented the effect of surface flux feedbacks by assuming them to be local and lumping them together with radiative feedbacks, which in turn were parameterized as proportional to precipitation. They did this by increasing the proportionality coefficient, r, relating radiative anomalies to precipitation anomalies compared to that estimated from observations (Bretherton and Sobel 2002; Lin and Mapes 2004), arguing that the parameterized flux anomalies represented those in both radiative and wind-induced turbulent surface fluxes. Recent work suggests such increases are justifiable. Araligidad and Maloney (2008) found that intraseasonal latent heat flux anomalies alone are about 20% of precipitation anomalies in the west Pacific warm pool, which appears broadly consistent with the wind speed-precipitation relationship found on daily time scales by Back and Bretherton (2005).

With a proportionality coefficient of around 0.25 or greater — corresponding to a net 25 W m^{-2} transfer of enthalpy from ocean to atmosphere for each 100 W m^{-2} of column-integrated latent heating — and other parameters set at typical control values, the model of SG03 is linearly unstable to free oscillations which qualitatively resemble those found in the observational studies (Waliser 1996; Fasullo and Webster 1999; Stephens et al. 2004). The growth rate of the oscillations is sensitive to several parameters in the model, including the surface enthalpy flux feedback parameter r, the mixed layer depth, the time scale for convective adjustment, and the GMS (which is assumed constant), but the period is robustly in the intraseasonal range.

SG03 argued that their model on its own was not adequate to represent the MJO, as its single-column structure makes it incapable of capturing the MJO's horizontal structure and propagation. They argued instead that their model might better depict how a small horizontal area responds to the passage of an MJO disturbance, with that disturbance viewed as an external forcing. With r small enough to render the model stable — such as is appropriate if it represents radiative feedbacks alone — the model solution is the forced response of a damped oscillator. SG03 imposed the forcing through the atmospheric temperature field, which they took to have a sinusoidal variation with intraseasonal period. Maloney and Sobel (2004) instead imposed an intraseasonally fluctuating surface wind speed forcing in their application of the SG03 model, making it appropriate to set $r \sim 0.1$, representing radiation only.

A parameter of particular interest in this model is the mixed layer depth. The amplitude of the model oscillations in precipitation as a function of mixed layer depth is presented in Figure 1, reproduced from Maloney and Sobel (2004). In this curve, the precipitation amplitude has a maximum at a particular value of the mixed layer depth, around 10-20 meters. It falls off slowly as mixed layer depth increases past the maximum, and rapidly as the mixed

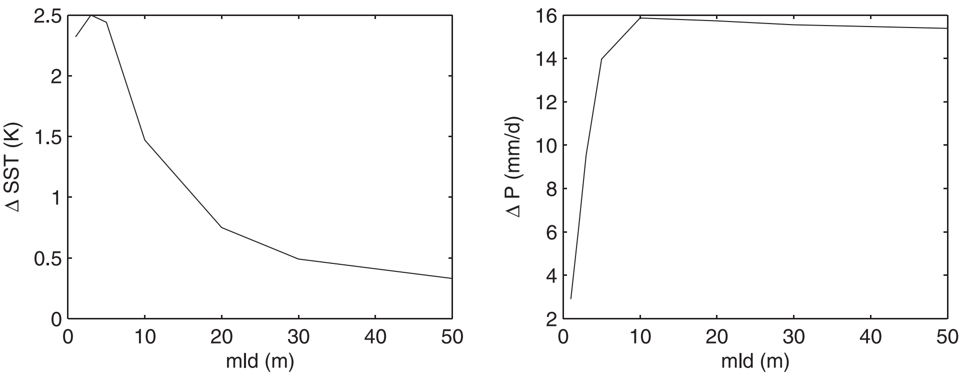

Figure 1. Peak-to-peak amplitude of oscillations in SST (*K*, left) and precipitation ($mm\ d^{-1}$, right) vs. mixed layer depth (m), SG03 model. From Maloney and Sobel (2004).

layer depth approaches zero. A similar (if weaker) amplitude maximum was found in the GCM results of Maloney and Sobel (2004) and is supported by a recent analysis of spatial and seasonal variability in the amplitude of intraseasonal variability compared to that in mixed layer depth (Bellenger and Duvel 2007).

The amplitude decrease for mixed layers deeper than that at which the maximum occurs indicates that in this model, ocean coupling can modestly enhance intraseasonal variability, since infinite mixed layer depth corresponds to fixed SST. This decrease was also found in the GCM study of Watterson (2002), and is implied in those studies which find stronger intraseasonal variability in coupled models than in atmospheric models over fixed SST.

The vanishing of the response as mixed layer depth goes to zero reflects the fact that surface enthalpy fluxes are critical to the oscillations in this model. As the mixed layer depth approaches zero, the net surface enthalpy flux must also vanish. This kills the oscillation because the gross moist stability is positive, requiring net moist static energy input into the column (which is equivalent to net surface enthalpy flux under our assumptions) in order to generate circulation anomalies.

A mixed layer of zero depth, or ‘‘swamp’’, with zero heat capacity but an infinite moisture supply, may be thought of as a crude representation of a land surface in a tropical region during its monsoon season, although it is not a good representation of land surface processes in general. When soil becomes subsaturated, variations in the Bowen ratio (ratio of sensible to latent heat flux) can result in ‘‘soil moisture memory’’ by which the land-atmosphere interactions have intrinsic time scales of up to several months. This effect appears most important in semi-arid regions (Koster and Suarez 2001). In active monsoon regions, soil moisture memory is less important, and we assume that modeling the land surface dynamics by a swamp, with zero heat capacity but infinite available moisture, is adequate for purposes of understanding the qualitative dynamics of the coupled system. Treating each horizontal location as represented by an independent single column model under WTG and SQE (SG03), we arrive at the prediction that *the amplitude of intraseasonal variability in deep convection can vary locally depending on surface type, and should be small over land and larger over ocean.*

3. Observations

3.1. Intraseasonal variance maps

3.1.1. Results

Figures 2 and 3 show maps of 30-90 day variance in precipitation from the TRMM 3B42 precipitation data set and outgoing longwave radiation (OLR) from the NOAA interpolated OLR data set, respectively, for the months November-April and May-October. Similar maps are shown in previous studies (e.g., Weickmann et al. 1985; Zhang and Hendon 1997; Vincent et al. 1998; Fasullo and Webster 1999; Sperber 2004, Duvel and Vialard 2007). Daily-averaged TRMM precipitation data during 1998-2005 averaged to a $1^\circ \times 1^\circ$ grid are used. The TRMM 3B42 product we use here incorporates several satellite measurements, including the TMI and TRMM precipitation radar to calibrate infrared measurements from geostationary satellites (Adler et al. 2000). The daily-averaged NOAA interpolated OLR product is used during 1979-2005 on a $2.5^\circ \times 2.5^\circ$ grid (Liebmann and Smith 1996). Unless otherwise stated, intraseasonal bandpass filtering is conducted using two 60-point non-recursive digital filters with half-power points at 30 and 90 days.

During southern hemisphere summer, intraseasonal variability is dominated by the canonical MJO, which has very large horizontal scales (Figs. 2a and 3a). However, the variance maps also exhibit prominent smaller-scale patterns. These small-scale patterns consist primarily of enhanced intraseasonal variance over the oceans and reduced variance over land. Particularly striking is the land-sea contrast in the Maritime Continent region. This land-sea variance contrast

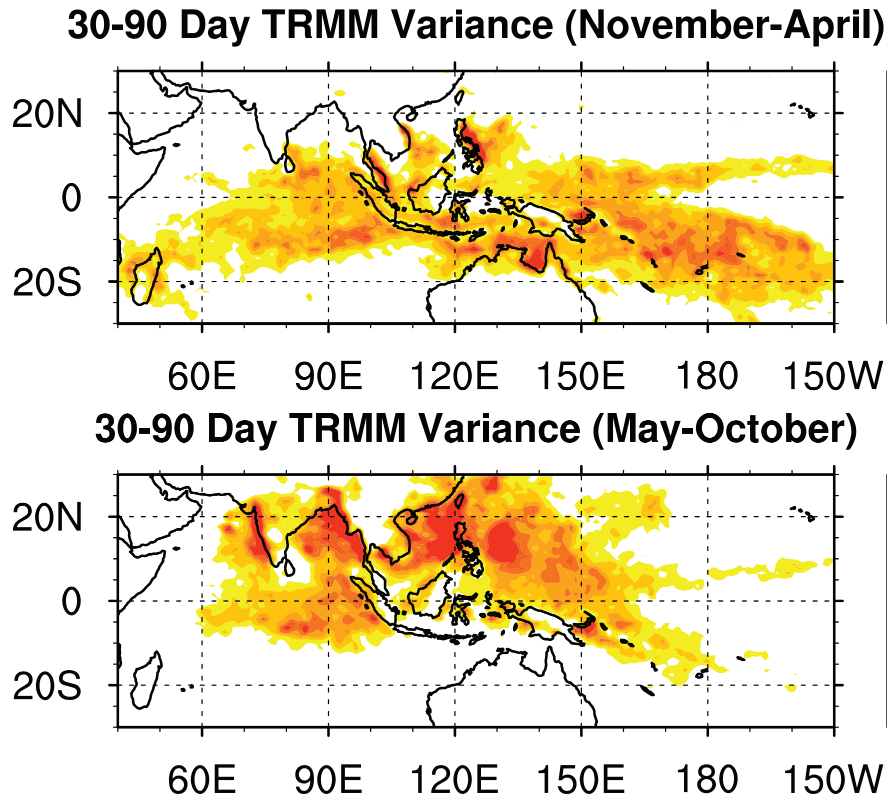

Figure 2. Intraseasonal variations in rainfall for a) November-April and b) May-October (mm^2 d^{-2}).

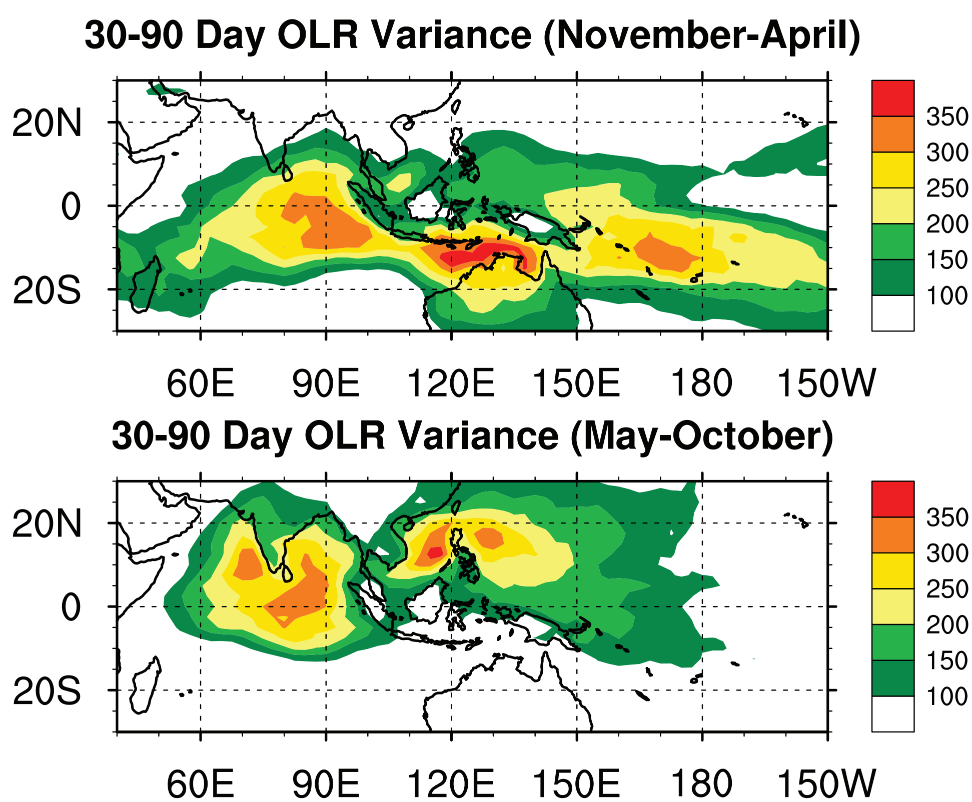

Figure 3. Intraseasonal variations in OLR for a) November-April and b) May-October (W^2 m^{-4}).

is also prominent in northern summer (Figs. 2b and 3b), and is evident down to the smallest scales resolved by the data.

Figures 4 and 5 show the climatological mean precipitation and OLR, respectively for May-October and November-April. The patterns of the southern and northern hemisphere monsoons are evident. In May-October (Figs. 4b and 5b), the climatological precipitation resembles the intraseasonal variance in its horizontal structure, with maxima in rainfall over the oceans and minima over land. In November-April, however, the same is not true (Figs. 4a and 5a). Climatological convection maximizes over the large islands of the maritime continent region, while intraseasonal variance minimizes there. This tendency is most striking when examining the OLR product, although neither the patterns of intraseasonal variance nor those of climatological precipitation shown above are sensitive to the choice of data set. Similar patterns are apparent, for example, in the CMAP (Xie and Arkin 1997) precipitation data set (not shown).

3.1.2. A proposed explanation of observed variance patterns

Despite the extreme simplicity of the SG03 model, the small-scale features in the observed patterns of intraseasonal precipitation variance are consistent with it, and thus with the assumptions of convective quasi-equilibrium and WTG which it incorporates. This is true in at least two important respects:

1. The fine-scale structure in precipitation variance suggests that, despite the large-scale structure of the flow features associated with intraseasonal variability, it may be appropriate to consider variations in convection in terms of a local picture, which can be captured by a single-column model using WTG [or perhaps also by other single-column parameterizations of large-scale dynamics (e.g., Bergman and Sardeshmukh 2004; Mapes 2004)].

2. The fact that intraseasonal precipitation variance maximizes over the ocean and minimizes over the land is consistent with an important role for interactive variations in the net surface enthalpy flux in generating the variance, since such variations can have significant amplitude over ocean but not over land. The variations in net enthalpy flux most likely have turbulent and radiative components, corresponding to wind-evaporation and cloud-radiative feedbacks. It is not possible to determine which is more important on the basis of either the idealized model of SG03 or the patterns of variance alone. The observational analysis of Waliser (1996) suggests that the two components may be of

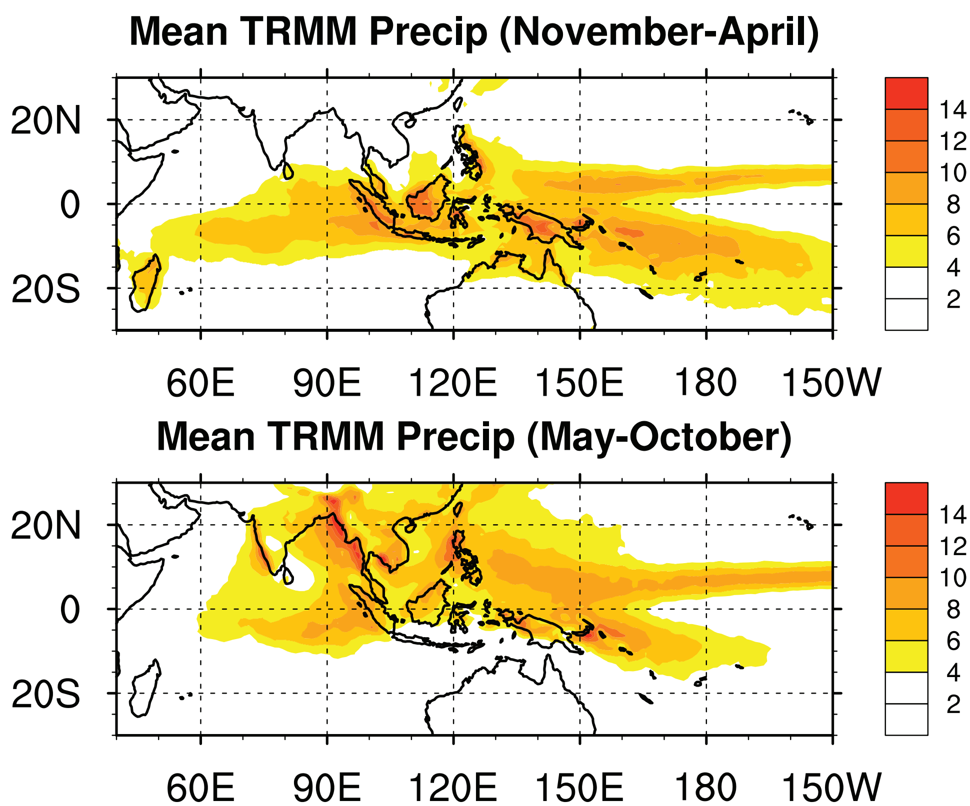

Figure 4 Climatological rainfall for a) November-April and b) May-October (*mm d*$^{-1}$).

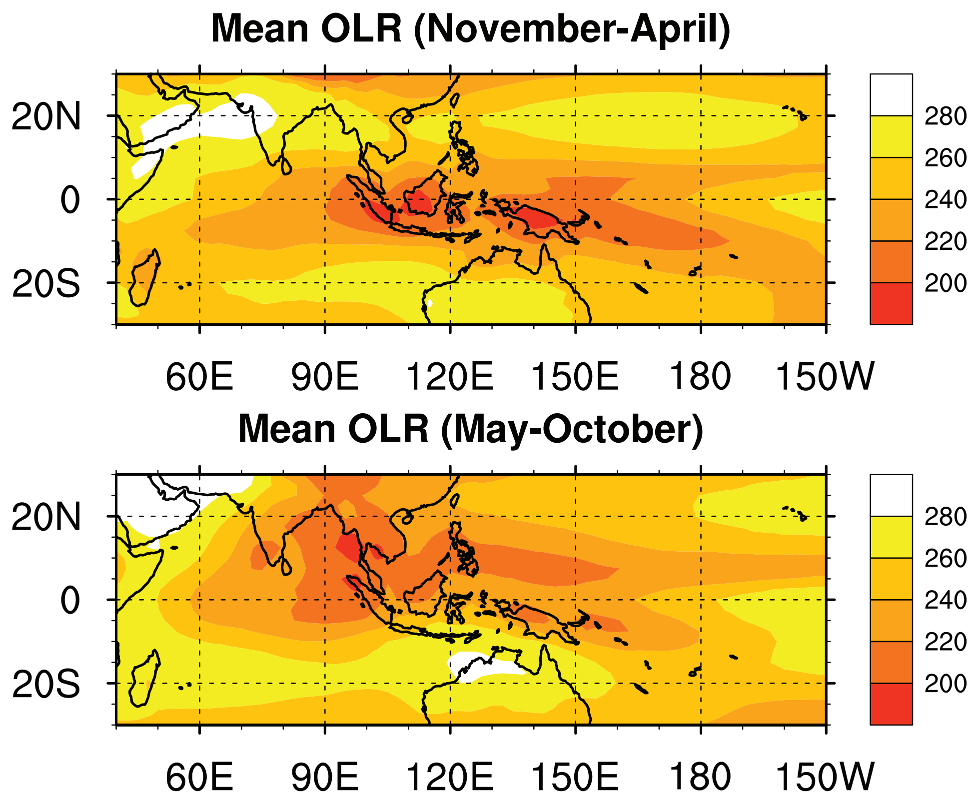

Figure 5. Climatological OLR for a) November-April and b) May-October ($W\ m^{-2}$)

comparable magnitude, while Hendon and Glick (1997) suggest that the relative importance of the two may vary with location.

3.1.3. Alternative explanations, 1: Orography

An alternative explanation for the patterns shown in Figs. 2 and 3 is that the patterns are controlled by orographic effects rather than by land-sea contrasts in surface enthalpy fluxes. For example, in May-October, the patterns of intraseasonal precipitation variance (especially away from the maritime continent) resemble the patterns of mean rainfall, which are certainly influenced by orography. The orographic influence most likely is largely due to the dynamical forcing of upslope flow as monsoon winds impinge on mountain ranges, and is thus dynamically distinct from the thermodynamic effects of land-sea contrasts. Focusing on the Indian and Southeast Asian regions, one maximum (in both variance and mean rainfall) occurs over and just upstream of the Western Ghats, while another occurs over the Bay of Bengal, upstream of the mountains on the Cambodian coast. It might be argued that the variance minimum in between (fig. 2b), over the Indian subcontinent, owes its existence to the minimum in mean rainfall, and that the latter minimum owes its existence to orography. Southern India lies in the rain shadow of the Western Ghats, making it drier than the regions upstream and downstream. In general, where mean rainfall is smaller variability will also be smaller. A related argument follows from the analysis of Hoyos and Webster (2007), who present evidence both that much of the total precipitation falling in the Asian monsoon is associated with intraseasonal events, and that the precipitation distribution in these events is modulated by orography.

While orography undoubtedly influences the rainfall patterns shown above, orographic effects alone cannot explain all aspects of the intraseasonal variance maps shown in Figures 2 and 3. We contend that these patterns can be explained more generally by the land-sea difference in heat capacity, which in turn suggests a role for surface fluxes. This is particularly apparent when we consider the November-April and May-October results together, and look for the most general explanation for both. Consider the maritime continent region in November-April. Grossly speaking, intraseasonal variance maximizes over ocean and minimizes over land. Mean rainfall does the opposite, particularly over the largest Indonesian islands (though there is also more complex structure within individual islands which is surely influenced by topography). The large mountains on these islands most likely play a role in inducing the mean precipitation maxima (e.g., Qian 2008). If the structure of the intraseasonal variance were determined by the structure of the mean rainfall, we would expect to see variance maxima over these large islands, coincident with the mean rainfall maxima, but instead these are regions of relatively low variance. The dominance of surface type over orography in the determination of the intraseasonal variance patterns is also suggested by the pattern over and around northern Australia, where intraseasonal variance also maximizes over ocean to a greater extent than mean rainfall does. Northern Australia lacks significant orography, so it seems almost certain that this difference is due to land-sea contrast. Even in May-October, a primary role for land-sea difference in heat capacity rather than orography is suggested by the maximum variance to the east of the Philippines, which lies neither over nor immediately upstream of any mountains.

3.1.4. Alternative explanations, 2: Diurnal cycle

Besides orography, another explanation might be that convection over land is dominated by the diurnal cycle, and this disrupts the variability at intraseasonal time scales (e.g., Wang and Li 1994). The diurnal cycle in precipitation persists over land even in suppressed phases of the ISOs, and since precipitation cannot be negative this implies larger precipitation than over ocean, where the diurnal cycle is weaker and suppression of precipitation can be more complete (e.g., Ichikawa and Yasunari 2008). This is consistent with the explanation that we present in section 2e. Over land, due to the small heat capacity of the surface, the preferred frequency for coupled single-column oscillations is much higher than over ocean. Thus, the system responds more strongly to diurnal solar forcing and less strongly to intraseasonal forcing by wind or atmospheric dynamics. This is shown in figure 6, which displays results from a set of calculations with the SG03 model forced by sinusoidal variations in solar insolation with a period of one day and amplitude of 40 $W\ m^{-2}$. The peak-to-peak precipitation amplitude is plotted as a function of ocean mixed layer depth. Maximum amplitude occurs for mixed layer depth of about 1 m, still perhaps large as a value representative of the heat capacity of land, but much smaller than typical ocean values.

This is nothing but a simple linear argument based on the single column framework discussed above. Considering each horizontal location to support a local recharge-discharge oscillation which is forced by a larger-scale intraseasonal oscillation, locations over land respond weakly while locations over land respond strongly, due to the difference in heat capacities. The same difference in heat capacities leads to a weak response to the diurnal cycle in solar forcing over ocean compared to that over land. This explains both the stronger diurnal cycle over land and the stronger intraseasonal variability over ocean without requiring any direct interaction between the diurnal cycle and the intraseasonal oscillation (other than that associated with the inability for precipitation to be negative; this does lead to a weakening of the intraseasonal oscillation over land as discussed above, but the ultimate cause of this is still the difference in the linear dynamics between the local responses to diurnal and intraseasonal forcings).

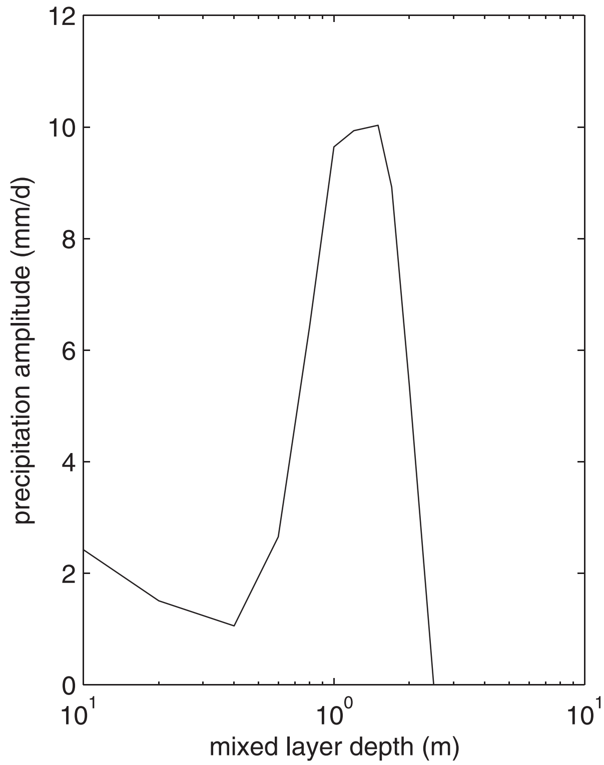

Figure 6. Response of the Sobel-Gildor (2003) model to ''solar'' forcing with a diurnal period. Peak-to-peak precipitation amplitude is plotted as a function of ocean mixed layer depth, for a set of nonlinear integrations in which solar insolation varies sinusoidally with a period of one day and an amplitude of 40 $W\ m^{-2}$.

It is also possible that the stronger diurnal cycle could interfere directly with intraseasonal variability over land, thus explaining the land-sea contrast in intraseasonal variance through some mechanism (as yet unexplained) which would not require that the intraseasonal oscillations depend on surface flux feedbacks. Since there is a substantial gap between diurnal and intraseasonal frequencies, such interference would have to be inherently nonlinear. We cannot rule this out, but Occam's razor seems to favor the linear explanation involving surface fluxes.

3.2. Correlation between surface latent heat flux and precipitation

Surface fluxes can be important to intraseasonal variability only if anomalous surface fluxes are able to influence the occurrence or intensity of deep convection. If this is the case, we might reasonably expect surface fluxes and precipitation to covary in space and time. The degree of covariance has been assessed in a couple of recent studies. Back and Bretherton (2004) showed that there is a small but significant correlation on the daily time scale between surface wind speed (which plays the dominant role in controlling the surface turbulent flux variations over tropical oceans) and precipitation, with the correlation being stronger for regions of high column water vapor content.

On the intraseasonal time scale, surface latent heat flux and precipitation (or quantities related to it, such as OLR) have been found to be locally correlated. The peak correlation is typically found when latent heat flux lags convection by a week or so, though that optimal lag varies slightly from one study to the next (e.g., Hendon and Glick 1997, Shinoda et al. 1998, Woolnough et al. 2000). Araligidad and Maloney (2008) demonstrated a strong instantaneous correlation (0.7) between November-April 30-90 day QuikSCAT wind speed and TRMM precipitation within the west Pacific regions of strong intraseasonal precipitation variance shown in Figures 2 and 3. Araligidad (2007) demonstrated similar strong correlations in the Indian Ocean during both summer and winter. Consistent with a strong covariance of precipitation and wind-driven fluxes, Araligidad and Maloney (2008) showed a significant correlation between intraseasonal TRMM precipitation anomalies and Tropical Atmosphere Ocean buoy latent heat flux anomalies. For example, Figure 7 is derived from Araligidad and Maloney (2008) and shows a scatterplot of intraseasonal latent heat flux versus precipitation anomalies at 8S, 165E during November-April of 1999-2005, within the band of strongest intraseasonal precipitation and OLR variance of Figures 2 and 3. If only the wind-driven portion of the latent heat flux anomaly is retained in this analysis, the correlation is about

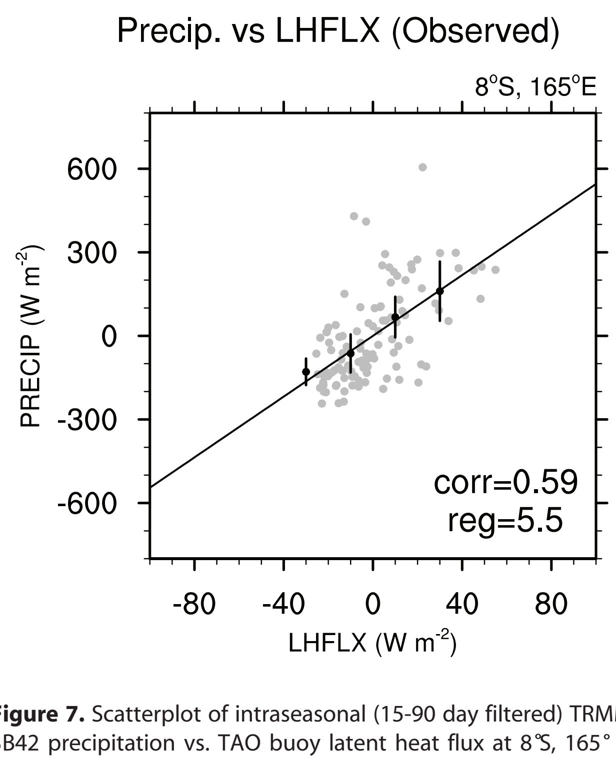

Figure 7. Scatterplot of intraseasonal (15-90 day filtered) TRMM 3B42 precipitation vs. TAO buoy latent heat flux at 8°S, 165° E, from Araligidad and Maloney (2008). Regression and correlation coefficients are indicated on the plots, and the black points represent binned averages, with bars depicting the 90% confidence limits about those averages.

0.1 higher, indicating that intraseasonal anomalies in air-sea humidity difference (forced primarily by SST variations) act to reduce the correlation of latent heat flux and precipitation versus if the wind-driven component were acting alone.

A positive covariance between precipitation and latent heat flux is encouraging regarding the ability for wind-evaporation feedbacks to support the MJO, in that it suggests that surface latent heat fluxes can influence convection. However, for surface flux feedbacks to drive the MJO, it is essential for wind-induced fluxes to engender a positive covariance of intraseasonal tropospheric temperature and diabatic heating. Such a positive correlation would indicate eddy available potential energy (EAPE) generation, with subsequent conversion of EAPE to eddy kinetic energy supporting the large-scale MJO circulation against dissipation. It is difficult to diagnose these energy conversions accurately from observations, but the evidence from studies done to date suggests that the phase relationships are consistent with a role for surface enthalpy fluxes in the instability of the MJO. Hendon and Salby (1994) used satellite observations of OLR and tropospheric temperature to show that heating and temperature are positively correlated over the region of strong intraseasonal convective activity, being almost perfectly in phase in the Indian ocean where the MJO is growing in amplitude. Similar results were found in a more recent observational study by Yanai et al. (2000). Since surface latent heat flux lags precipitation only by a small amount (compared to the 30-60 day period of the mode), and radiative heating is exactly in phase with precipitation (within the accuracy of the observational estimates) the total surface enthalpy flux anomaly is positioned to induce convective heating anomalies with the correct phase to generate EAPE, particularly in the growing phase of the MJO life cycle.

The fact that surface flux anomalies lag precipitation anomalies on intraseasonal timescales indicates that the fluxes do not play a role in the propagation of the ISOs. Rather, it appears that the fluxes retard propagation. This is not at all inconsistent with the claim that the fluxes are important to the destabilization of the modes. It is true that in the original linear models (Emanuel 1987; Neelin et al. 1987) fluxes lead precipitation, aiding in propagation. However, there are clear examples of other (more complex) models in which fluxes have been explicitly shown to be destabilizing despite lagging precipitation, as observed (e.g., Maloney and Sobel 2004, Bellon and Sobel 2008b).

4. Numerical model results

4.1. Previous work

The hypothesis that interactive surface enthalpy flux feedbacks are essential to the dynamics of intraseasonal variability is testable in numerical models, under a perfect model assumption. This can be done by overriding the parameterizations which determine the net surface enthalpy flux, or its individual components, and forcing the fluxes to equal those from a climatology. The climatological fluxes can be taken from a control run of the same model having interactive fluxes. Prescribing fluxes in this way renders the surface flux feedbacks inactive, as the surface fluxes in the model are no longer a function of the instantaneous model variables. If surface flux feedbacks are essential to the model's intraseasonal variability, that variability should be eliminated, or at least significantly reduced in amplitude. A number of variations on these experiments may be useful, such as one in which surface wind speed, rather than the turbulent surface fluxes themselves, is prescribed. Wind speed can either be set to a climatology or, in a simpler but less clean experiment shown below, to a spatially and temporally constant value.

To our knowledge, experiments of this type have been done only with a couple of recent-generation general circulation models using realistic basic state SST. Maloney (2002) performed an experiment in which the surface wind speed was set to its climatological value in the computation of surface turbulent heat fluxes. As intraseasonal variations in these fluxes are largely controlled by wind speed variations — the WISHE feedback — this eliminated most (but not all) intraseasonal flux variations. The surface latent heat flux itself was set to its climatological seasonal cycle in one simulation in Maloney and Sobel (2004). In that study, eliminating surface flux feedbacks significantly reduced the amplitude of the simulated MJO, indicating an important role for surface flux feedbacks. In Maloney (2002), on the other hand, the elimination of WISHE actually increased the amplitude of eastward-propagating wind and precipitation variability. The complete disagreement between the results of these two studies is at first perplexing. The models used in them were rather similar. Both used the relaxed Arakawa-Schubert convection parameterization in successive versions of the NCAR Community Atmosphere model; the two models differed primarily in the treatments of convective downdrafts and cloud microphysics. However, the resulting relationships between intraseasonal convection and the large-scale anomalous circulation in these two models were significantly different, with enhanced convection occurring in anomalous easterlies and suppressed latent heat fluxes in the model of Maloney (2002), and in anomalous westerlies and enhanced latent heat fluxes (as observed) in the study of Maloney and Sobel (2004). Thus, removing wind-evaporation feedback might be expected to have different effects in these two models.

A few earlier GCM studies also tested the importance of surface turbulent and cloud-radiative feedbacks to intraseasonal variability (e.g. Hayashi and Golder 1986) , but given the considerable advances in simulation capability in the last two decades, it may be most productive to focus on results from more recently developed models. In some relatively recent GCM studies using zonally-symmetric SST distributions strong sensitivity to WISHE has been found (e.g. Hayashi and Golder 1997, Colon et al. 2002). It might be argued that the differences between the basic state wind fields in these calculations and the observed wind fields render their relevance to real intraseasonal variability somewhat indirect. On the other hand, we do not understand that variability well enough to be sure what the role of the basic state is.

Given the sensitivity of GCM results to convective parameterization, it is desirable to use cloud-resolving models to examine the mechanisms of intraseasonal variability. Until very recently it was not computationally feasible to do this because of the large domain sizes necessitated by the scales of intraseasonal variability. Grabowski and Moncrieff (2001) reduced the computational burden by considering a two-dimensional domain, which may be considered to represent the longitude-height plane along the equator. In their simulations, surface flux feedbacks were not important to large-scale organization of convection. They did not argue that their results were relevant to the MJO, focusing their discussion instead on convectively coupled Kelvin waves.

More recently, improvements in both computational power and simulation technologies allow for more direct assaults on the intraseasonal variability problem with models that do not require convective parameterization. These technologies are only beginning to be used to unravel the mechanisms of intraseasonal variability. Grabowski (2003) found that interactive surface latent heat fluxes were essential to the development of the MJO in simulations using the multiscale modeling framework (MMF; Randall et al. 2003). Z. Kuang (pers. comm.) has found a similar result in global cloud-resolving simulations whose computational expense was reduced using the ''diabatic acceleration and rescaling'' methodology (Kuang et al. 2005, Pauluis et al. 2006). These early results are in agreement with the other arguments and evidence we present here, but are not conclusive. Many different model configurations and choices are possible even when those associated with convective parameterization are eliminated, so that different models can still yield different results. Much more work along these lines is needed in order to make the most of the new capabilities associated with global cloud-resolving models and the MMF.

4.2. Results with the GFDL AM2

In this section we present results from new simulations with the Geophysical Fluid Dynamics Laboratory's Atmospheric Model 2.1 (AM2.1), which is the atmospheric component of the coupled climate model CM2.1.

4.2.1. Model description

With the exception of the modification to the convection scheme that we describe below, the model used here is identical to that presented by the GFDL Global Atmospheric Model Development Team (Anderson and Co-authors 2004). It has a finite volume dynamical core, with $2^\circ \times$

2.5° horizontal resolution, and 24 vertical levels. The model is run over realistic geography and climatologically varying SSTs. The simulations are run for 11 years, with statistics taken over the last 10 years.

The convection scheme is a version of the Relaxed Arakawa-Schubert (RAS) scheme (Moorthi and Suarez 1992). In the RAS scheme, convection is represented by a spectrum of entraining plumes, with a separate plume corresponding to each model level that can be reached by convection. The entrainment rates in these plumes are then determined by the requirement that the levels of neutral buoyancy of the plumes correspond to model levels. The convection scheme in AM2 also uses the modification of Tokioka et al. (1988), in which convection is not allowed to occur when the calculated entrainment rates are below a critical value λ_0 determined by the depth of the subcloud layer z_M, with $\lambda_0 = \alpha/z_M$. Thus with larger values of the Tokioka parameter α, convection is prevented from reaching as deeply. Inspired by the results of (Tokioka et al. 1988) and (Lin et al. 2008) showing that larger values of the Tokioka parameter lead to stronger and more realistic MJO variability, we change α from its standard AM2 value of $\alpha = 0.025$ to the larger value $\alpha = 0.1$ in order to increase the MJO variance in the model. We also show results using the standard AM2 value of the Tokioka parameter, in which the simulated MJO is weak.

In order to identify the importance of WISHE to the model MJO, we construct no-WISHE simulations by replacing the wind speed dependence in the surface flux formulation with a constant value representative of typical values over the tropics, 6 *m*/*s* everywhere. This modification is very simple to implement, as it only requires changing one line of code. However, in addition to preventing intraseasonal variations in turbulent surface fluxes (the intended effect) this modification also alters the climatology of the model, since the model's actual climatological surface wind speed is not constant in either space or season. We therefore must confirm that the climatological precipitation distribution in our no-WISHE simulation does not become so different from that in the control model (or in observations) as to render the experiment irrelevant. The annual mean precipitation distributions for the control and no-WISHE cases are plotted in Figure 8. This figure shows that while there are some important changes in the precipitation distribution when WISHE is removed (e.g., less precipitation in the NW Pacific and more precipitation in the Indian Ocean), the distributions remain qualitatively similar enough to merit comparison of the MJO characteristics.

4.2.2. Results

As a first measure of the strength of the MJO with and without WISHE in these simulations, we show in Figure 9

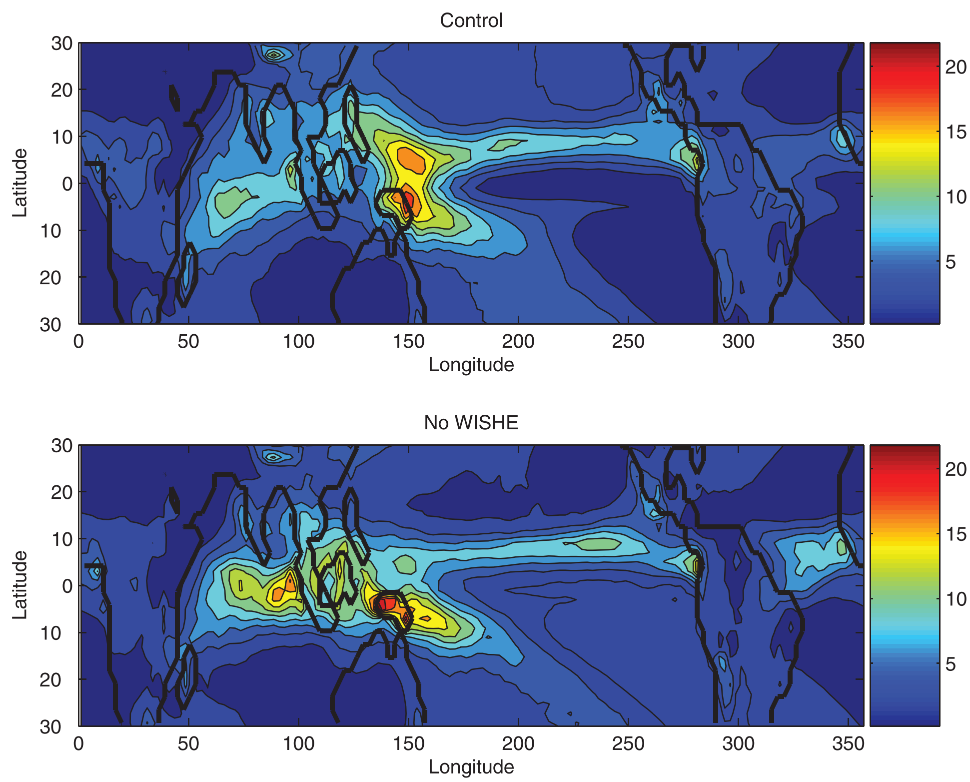

Figure 8. Annual mean precipitation (*mm* d^{-1}, contour interval 2 *mm* d^{-1}) for the control and no-WISHE simulations GFDL AM2 with Tokioka parameter $\alpha = 0.1$ (see text for details).

lag-regression plots for 30-90 day filtered equatorial ($10^\circ N$–$10^\circ\ S$ averaged) zonal wind at 850 *hPa* (U850) for the months of November-April for observations, the control case (with $\alpha = 0.1$), and the control case without WISHE. Regression coefficients are scaled by the 1σ value of the reference $156^\circ E$ time series, and stippling indicates where the correlation coefficient is significantly different from zero at the 95% confidence level. The control case MJO propagation is quite similar to observations in many aspects, including implied phase speeds of approximately 5 *m*/*s*, large variance over the Indian and Pacific Ocean, and faster propagation over the central/eastern Pacific, though the regression coefficients are generally a bit weaker than those in observations, especially just to the west of the reference point. When WISHE is suppressed, the amplitude of the MJO is reduced significantly. Only over the Pacific does any significant correlation exist away from the reference point. This clearly demonstrates that the MJO in this model is strongly influenced by WISHE.

As an alternative measure of the MJO intensity, we examine the intraseasonally averaged (30-90 days) wavenumber spectrum for U850, separated into eastward and westward propagating components. The ratio of eastward to westward variance at wavenumber 1 is often used as a measure of the MJO strength (see, e.g., Zhang et al. 2006). The ratio of eastward to westward variance at wavenumber 1 in the control case is 2.60, which is stronger than nearly all the atmosphere-only models in the Zhang et al. (2006) study, although weaker than observations, which have a value of 3.5. When WISHE is removed, this E/W ratio is reduced to 1.16, corroborating the result that WISHE is fundamentally important to the MJO in this model.

As a test of robustness to changes in model physics, we examine the same MJO diagnostics for the standard version of AM2.1, in which the Tokioka parameter $\alpha = 0.025$. Examining lag correlations for this configuration in Figure 10a, one can clearly see weaker MJO propagation everywhere as compared to the Tokioka-modified control case in Figure 9b. Correlations are significantly weaker, especially in the Pacific basin. When WISHE is removed in this model configuration (Figure 10b), the model MJO is little affected. There is a small indication of decreased MJO correlations in the Indian Ocean and immediately downstream of the reference point, but these changes are subtle. The ratio of eastward to westward intraseasonal variance at wavenumber 1 for U850 is reduced from 1.31 to 1.08 when WISHE is removed, indicating a small decrease in MJO amplitude with WISHE in this diagnostic. Generally speaking, removing WISHE has a small effect on the MJO in this model configuration; but then, the MJO is weak to begin with.

The results from the model with $\alpha = 0.1$ bring to two the number of recent-generation models in which WISHE has been found to be important to the MJO in simulations with realistic basic states, the other being that used by Maloney and Sobel (2004). Both models use versions of the RAS convective parameterization, so they are not entirely unrelated, but most other aspects of the two models are different. In the models of Maloney (2002) and the AM2.1 with $\alpha = 0.025$, WISHE is not important to the simulated MJO. On the other hand, these two models have MJO simulations which resemble observations less closely than do those of Maloney and Sobel (2004) and the AM2.1 simulations with $\alpha = 0.1$. At least for this small sample of models, a better simulation of the MJO seems to be associated with an increased role for WISHE.

The simulations discussed above address only the role of the surface latent heat flux, and in the case of Maloney (2002) and the AM2.1 calculations described here, only the wind-induced component of that flux. A few studies with full-physics GCMs over realistic of continents and sea surface temperature have assessed the role of radiative flux perturbations in simulated MJO dynamics. Lee et al. (2001) found in an aqua-planet GCM that overactive cloud-radiative feedbacks degraded the MJO simulation by inducing spurious small-scale disturbances; reasonable changes to the model's physical parameterizations mitigated this degradation. Other studies have been done in more idealized frameworks. Raymond (2001) found that cloud-radiative feedbacks were essential to the MJO simulated in his intermediate-complexity model, with surface turbulent fluxes also playing a significant role. Grabowski (2003) found in aqua-planet simulations with what is now called the ''multiscale modeling framework'' (MMF) that cloud-radiative feedbacks were not important to his simulated MJO disturbances, while surface latent flux feedbacks were essential to the disturbances' development. Lin et al. (2008) found no effect of cloud-radiative feedbacks on the MJO in a model in which the MJO was weak to begin with.

4.3. Relation between MJO bias and mean precipitation bias

Systematic error in simulation of the MJO comprises only one out of several common errors, or ''biases'' in a typical model's simulation of the tropical climate; other common biases include the tendency towards double ITCZs and poor simulation of ENSO (e.g., Bretherton 2007). The various biases appear related to each other to some extent; changes in model physics or numerics often affect more than one. Improvement of the MJO simulation is often possible by methods which appear somewhat consistent across various studies and models (Tokioka et al. 1988; Wang and Schlesinger 1999; Maloney and Hartmann 2001); broadly, anything which makes it more difficult for the deep convective parameterization to fire tends to improve the MJO simulation. However, this improvement seems often to come at the cost of increased biases in other aspects of the simulated tropical climate.

Fig. 11 shows the biases in the climatological annual mean precipitation field (with respect to TRMM 3B42, 1998-2006) in four models. Two are the GFDL AM2, with the Tokioka parameter set to 0.1 and 0.025; these are the same calculations used in the preceding section. The other two are the NCAR CAM 3.1 with the relaxed Arakawa-Schubert convective parameterization, and convective rainfall re-evaporation fractions set to 0.6 and 0.05 (weak evaporation). In this version of CAM, a stronger MJO is produced by increasing evaporation fraction (Maloney 2008). Both CAM simulations have a Tokioka-like minimum entrainment threshold set to 0.1/(1000 m).. The models with stronger MJOs (the second and fourth panels) have large regions in which precipitation is stronger than observed in the off-equatorial western north Pacific, centered around 150E, while those with weaker MJOs (the third and fifth) have stronger than observed precipitation in the equatorial maritime continent region centered around 120E. The off-equatorial "red spots" in the stronger-MJO models are qualitatively consistent with similar structures found in simulations under the multiscale modeling framework ("SP-CAM"; Khairoutdinov et al. 2008), which also has a strong MJO compared to that in most GCMs.

Many other aspects of the bias patterns are not consistent between the stronger-MJO and weaker-MJO models (Fig. 11), and again it is not appropriate to draw strong conclusions from a small sample of models. Nor do we have any mechanistic hypothesis that might explain the relationships between mean climate biases and MJO biases which do appear in this limited sample. Given the importance of bias reduction in global climate simulation, the issue seems worth exploring with a larger number of models. If MJO bias and other tropical climate biases are systematically related, that information might usefully constrain thinking on the physical mechanisms behind both biases.

The relationship between MJO bias and other tropical biases should also be kept in mind when we wish to assess the state of the art in MJO simulation. Model developers naturally focus their efforts on those biases most directly relevant to the primary task at hand. For global climate assessments such as that of the Intergovernmental Panel on Climate Change (IPCC), the focus is on key measures of the time-mean climate, or perhaps some gross measures (e.g., global mean surface temperature) of the climate's time evolution in the modern era. Relatively high-frequency

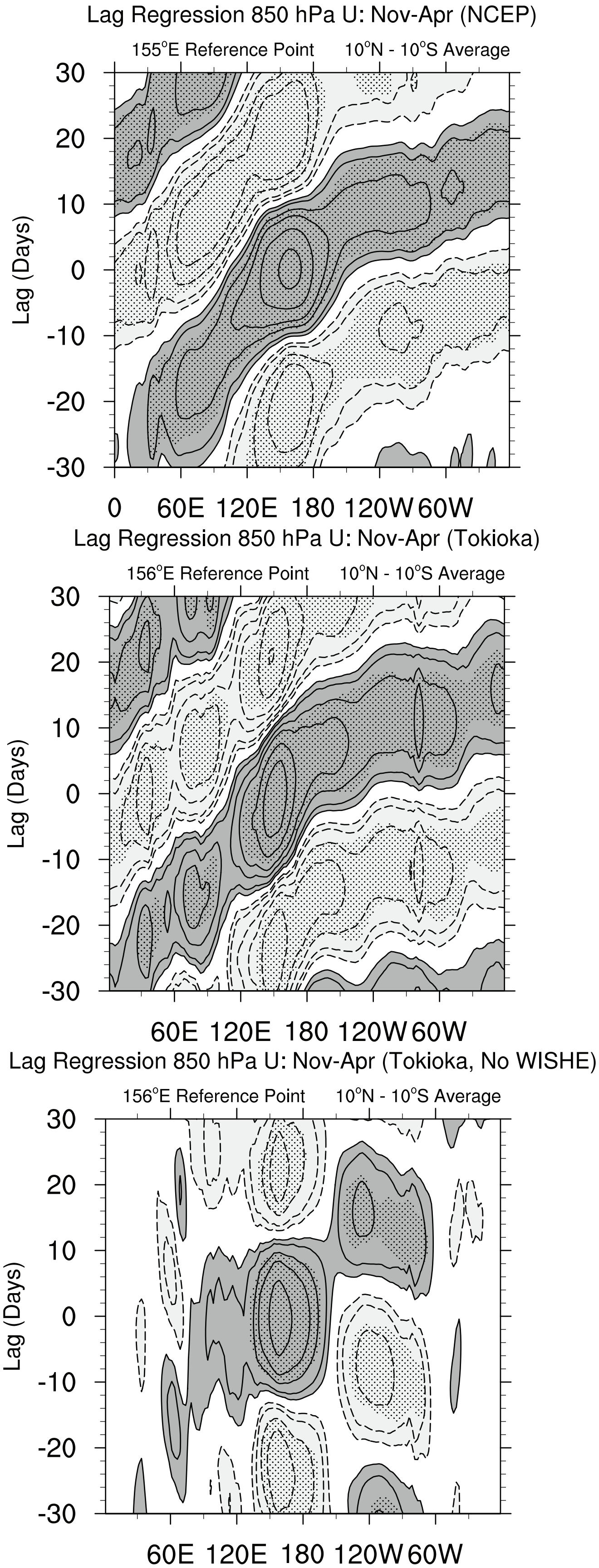

Figure 9. Lag-regression of $10^\circ S - 10^\circ N$ averaged, 30-90 day filtered zonal wind at 850 hPa (U850, ms^{-1}) against the time series of the same field at $156^\circ E$, in GFDL AM2 with Tokioka parameter $\alpha = 0.1$. The top panel shows results from the NCEP/NCAR reanalysis, the middle shows results from the model, and the bottom shows results from the model with no WISHE (see text for details). Regression coefficients are scaled by a 1σ value of the reference time series. Only data during November-April are used. Contours are plotted at $\pm$ 0.1, 0.2, 0.4, 0.6, 1.0, 1.5, 2.0, and 2.5 ms^{-1}. The zero contour is not shown. Values greater (less) than 0.1 (-0.1) are dark (light) shaded.

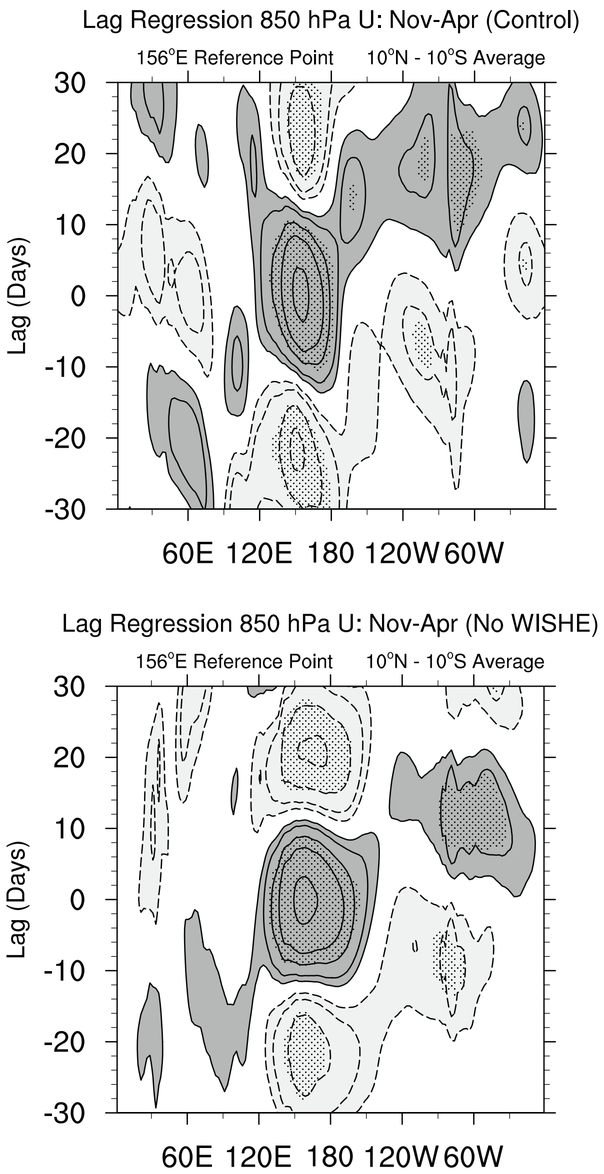

Figure 10. Lag-regression of U850 in GFDL AM2 , as in the lower two panels of fig. 9, but with Tokioka parameter $\alpha = 0.025$.

variability such as the MJO has not been a primary target of these assessments. To the extent that there are trade-offs in between mean biases and MJO bias, the MJO bias tends to be a lower priority. The intercomparison study of Lin et al. (2006) uses simulations performed for the IPCC's fourth assessment report (AR4), which have presumably been tuned for the mean climate rather than the MJO; in the case of the GFDL and NCAR models, we know from experience that the MJO simulation can be improved at some cost to other biases, and it seems likely that the same is true in other models. Our ability to simulate the MJO in the current generation of models is not necessarily optimally represented by the standard operational versions of those models.

5. Discussion

5.1. The crux of the matter, in theoretical context

The claim that surface enthalpy fluxes are essential to the dynamics of tropical intraseasonal variability is not new, going back at least 20 years to the studies of Emanuel (1987) and Neelin et al. (1987). We believe that, given the lack of broad agreement on the mechanisms of the MJO (despite decades of intense effort) and the evidence from both observations and GCMs discussed above to support the hypothesis that surface fluxes are important, the time has come to reassess this hypothesis in a more focused way.

We have not provided a comprehensive discussion of all theories for tropical intraseasonal variability. However, it should be uncontroversial to state that in many of these theories, interactive surface fluxes either are not essential or are absent altogether. We propose that it would be useful to divide the current set of theories into two subsets, one in which feedbacks involving surface moist enthalpy fluxes (including radiative fluxes) are essential and one in which they are not, and then attempt to eliminate one subset via focused numerical modeling studies, perhaps combined with further analysis of observations.

From a purely conceptual point of view, whether surface fluxes are essential to intraseasonal variability is a fundamental question. In extratropical dynamics, it has been found useful to divide the set of possible dynamical processes into those which are dry adiabatic and those which are not. If a phenomenon can be understood using adiabatic models, it is advantageous to do so. Similarly, it is natural when discussing tropical dynamics to divide the large set of possible processes into those which involve only deep convection and large-scale dynamics — that is, those which can be represented by moist adiabatic dynamics — and those in which diabatic processes external to both deep convection and large-scale dynamics, namely turbulent surface fluxes and radiative cooling, are involved. It has been a goal of theoretical tropical meteorology for several decades to determine whether the interaction of convection and large-scale dynamics alone can generate large-scale variability [as in early CISK models, as well as in more recent models with more complex physics (e.g., Mapes 2000; Majda and Shefter 2001a,b; Kuang 2008)] or whether interaction with diabatic processes external to convection and large-scale dynamics is necessary (as in first baroclinic mode QE models). Determining whether interaction with turbulent surface fluxes and radiation is essential to observed intraseasonal variability in particular (leaving aside other modes, for example higher-frequency convectively coupled waves) would be a major step forward in our understanding of the tropical atmosphere.

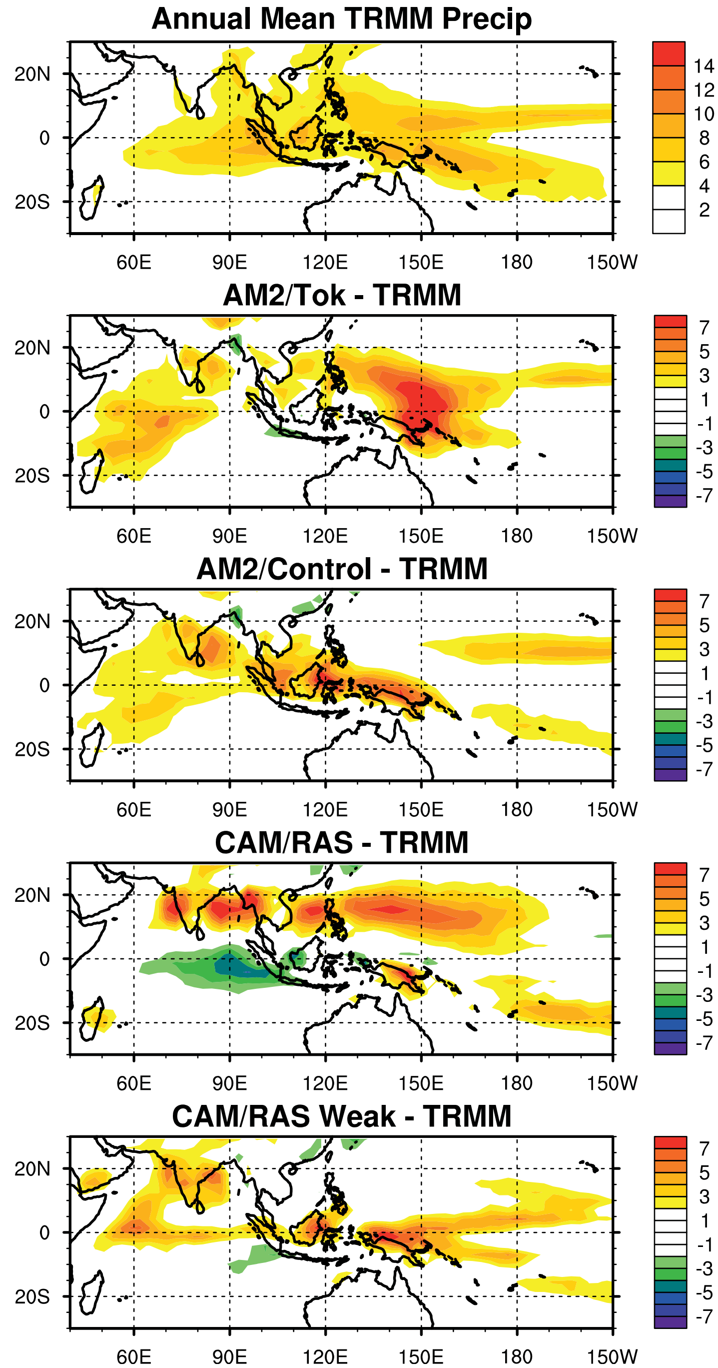

Figure 11. Rain biases (*mm* d^{-1}). The top panel shows annual mean rainfall climatology from the TRMM 3B42 data set, while the subsequent panels show the difference between a simulated GCM climatology and that, from, in descending order: the GFDL AM2 model with increased Tokioka parameter $\alpha = 0.1$, the standard GFDL model with $\alpha = 0.1$, the NCAR CAM 3.1with the RAS convection scheme and rain re-evaporation fraction of 0.6, and the NCAR CAM 3.1 with RAS and evaporation fraction of 0.05.

5.2. Proposal for further model intercomparison

Recent model intercomparisons (Zhang et al. 2006; Lin and coauthors 2006), have been performed which summarize the state of the art in simulating the MJO in general circulation models. Without restating the results of these studies in detail, the MJO simulations in the latest generation of models are on average superior to those in previous generations (Slingo et al. 1996; Sperber et al. 1997) in simulating eastward-propagating zonal wind variability in the tropics with a dominance of eastward vs. westward power, though even the best models still have deficiencies in their MJO simulations. While keeping model deficiencies in mind, we propose that interested modeling groups perform experiments like those described above, in which the total surface moist enthalpy flux, and ideally also its individual components, are set to climatology, eliminating feedbacks involving those fluxes. These experiments are likely to yield unambiguous information about the dynamics of a model's intraseasonal variability. The negative of the quantitative change in the strength of the intraseasonal variability in these experiments provides a direct estimate of the role of the eliminated feedbacks in the dynamics of the simulated variability in the control simulation. Besides GCMs with parameterized physics, these experiments can also be done in models with resolved convection such as the multiscale modeling framework (Grabowski 2003; Randall et al. 2003; Khairoutdinov et al. 2005; Khairoutdinov et al. 2008) or global cloud resolving models (Miura et al. 2007). Both of these technologies are showing great promise in simulating the MJO, and sensitivity experiments to determine the roles of surface turbulent and radiative flux feedbacks in their results would be particularly valuable.

Because simulations of intraseasonal variability are imperfect in all models, such experiments will not yield unambiguous information about the dynamics of intraseasonal variability in the real atmosphere. It is entirely possible that any given model, or even an entire generation of models (given the broad similarities of approach found in common physical parameterizations in climate models), is getting something close to the right answer for the wrong reasons, so that the results of these experiments would be misleading. It is perhaps also equally probable that different models will yield different results from these experiments.

Neither will such experiments provide any direct information about how to improve the simulation of intraseasonal variability in any given model. The importance of surface flux feedbacks to the dynamics of intraseasonal variability may not be related in any simple way to any particular property of the physical parameterizations of a model, nor to any other aspect of its construction (e.g., resolution or the dynamical core). These feedbacks are arguably a high-level, or ''emergent'' property of a given model. Even in the relatively simple models used in theoretical studies, it is often not apparent what determines the importance of surface flux feedbacks. We might expect it to be even less obvious in comprehensive GCMs.

If we were fortunate, the importance of surface flux feedbacks in a model would be related to that model's ability to simulate intraseasonal variability, as is the case in the very small sample of models discussed in section 4. We can imagine a scatter plot in which one axis is the fidelity of modeled intraseasonal variability to that observed (how to quantify this is a separate problem which we do not address here); the other axis is the importance of surface flux feedbacks to intraseasonal variability in the same model, as quantified by minus the change in MJO amplitude in an experiment where the surface flux (either total, or a given component) is set to climatology; and each point represents one model. A significant positive slope to the best-fit regression line would suggest that surface flux feedbacks are important to dynamics of intraseasonal variability in the real atmosphere, while a significant negative slope would suggest the opposite. Lack of any significant slope, of course, would be an ambiguous result.

In any case, knowledge of the role of surface fluxes in simulated intraseasonal variability would be useful to model developers. It seems likely that any increase in physical understanding of the dynamics of the modeled intraseasonal variability, such as quantification of the role of surface flux feedbacks, would help to guide in the formulation of hypotheses about how to improve a model. For example, an active role for surface fluxes in regulating intraseasonal variability may compel modelers to further develop parameterizations coupling mesoscale perturbations of moist entropy and gustiness to the boundary layer (e.g. Jabouille et al. 1996; Redelsperger et al. 2000), where they may significantly affect surface fluxes during MJO events.

5.3. Theoretical challenges

A determination that interactive surface fluxes are essential to the dynamics of intraseasonal variability would not constitute a complete theory for that variability. Even if we were able to resolve the importance of surface fluxes, questions that would remain unanswered include (among others): What is the relative importance of turbulent vs. radiative fluxes? How should the physics of deep convection and other unresolved processes be parameterized in order to yield the correct feedback between the fluxes and the large-scale dynamics of the mode? Are the large-scale dynamics essentially linear or nonlinear? What is the role of nonlinear energy transfers from synoptic- or mesoscales? Are extratropical influences important? What are the essential elements of the vertical structure? Is the structure of the basic state critical? What is the role of ocean coupling? Perhaps most importantly, what sets the phase speed of the disturbances? Theorists currently struggle with all of these questions. Proving or disproving the hypothesis that surface fluxes are essential to tropical intraseasonal variability would tell us that the ultimate energy source for the disturbances is or is not the ocean mixed layer, and in doing so would eliminate a large subset of theories, but much theoretical work would be left to do.

Assuming that the wind-induced component of the surface turbulent fluxes is important, the dynamics by which these fluxes interact with the dynamics of eastward-propagating MJO disturbances must be different in detail than that envisioned by Neelin et al. (1987) and Emanuel (1987). The real MJO is not a pure Kelvin wave (though it retains aspects of Kelvin wave dynamics), and the basic state surface winds in regions of strong tropical intraseasonal variability are westerly (e.g. Inness and Slingo 2003; Maloney and Esbensen 2007). One possibility is that nonlinear WISHE, rather than its linear counterpart, is acting. Studies which present numerical simulations with idealized or intermediate-complexity models (Raymond 2001; Sugiyama 2009a; Sugiyama 2009b) as well as comprehensive models (Maloney and Sobel 2004) provide suggestions of how this might work, but we do not have a simple analytical prototype model for nonlinear WISHE. Another possibility is that the dynamics are still fundamentally linear, but that changes to other aspects of the original E87 and N87 models (e.g., the identification of the MJO as a Kelvin wave, the assumption of a first baroclinic mode vertical structure, the simple quasi-equilibrium convection schemes, etc.) allow the requirement of mean easterlies to be relaxed. In the idealized model of Wang and Xie (1998), for example, the combination of coupling to a mixed layer ocean and parameterized radiative feedbacks is linearly destabilizing in the presence of mean westerlies.

In the case of the northern summer northward-propagating mode, there may be no fundamental theoretical problem. In at least one idealized model, a mode resembling that observed is linearly destabilized by WISHE (Bellon and Sobel 2008a, b).

6. Summary

We have argued that feedbacks involving the total surface enthalpy flux are important to the dynamics of tropical intraseasonal variability, possibly providing the primary energy source for intraseasonal disturbances. Observational evidence in support of this argument consists of maps of intraseasonal variance in precipitation and OLR as well as local correlations between precipitation and surface fluxes on intraseasonal time scales. Modeling evidence consists of results from several GCMs as well as idealized models in which surface flux feedbacks are demonstrably important if not essential to simulated intraseasonal variability.

Our understanding of the mechanisms responsible for intraseasonal variability is still poor after decades of study. We have argued that the time has come for a more systematic evaluation of the role of surface enthalpy fluxes, given all the tools at hand, with the aim of eliminating from consideration either those hypotheses in which surface fluxes are important or those in which they are not. Given the evidence presented here, the increasing fidelity with which comprehensive numerical models simulate intraseasonal variability, and the relative straightforwardness of assessing the importance of surface flux feedbacks in those models, we have argued that it would be particularly useful if a larger number of interested modeling groups were to perform the necessary assessments. Such efforts, combined with targeted observational and theoretical work, might enable the field to move forward in a more coordinated and productive way towards a better understanding of tropical intraseasonal variability.

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