Intraseasonal Variability in MERRA Energy Fluxes over the Tropical Oceans

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ABSTRACT

This paper investigates intraseasonal variability as represented by the recent NASA Global Modeling and Assimilation Office (GMAO) reanalysis, the Modern-Era Retrospective analysis for Research and Applications (MERRA). The authors examine the behavior of heat, moisture, and radiative fluxes emphasizing their contribution to intraseasonal variations in heat and moisture balance integrated over the tropical oceans. MERRA successfully captures intraseasonal signals in both state variables and fluxes, though it depends heavily on the analysis increment update terms that constrain the reanalysis to be near the observations. Precipitation anomaly patterns evolve in close agreement with those from the Tropical Rainfall Measuring Mission (TRMM) though locally MERRA may occasionally be smaller by up to 20%. As in the TRMM observations, tropical convection increases lead tropospheric warming by approximately 7 days. Radiative flux anomalies are dominated by cloud forcing and are found to replicate the top-of-the-atmosphere (TOA) energy loss associated with increased convection found by other observationally based studies. However, MERRA’s convectively produced clouds appear to deepen too soon as precipitation increases. Total fractional cloud cover variations appear somewhat weak compared to observations from the Moderate Resolution Imaging Spectroradiometer (MODIS). Evolution of the surface fluxes, convection, and TOA radiation is consistent with the “discharge–recharge” paradigm that posits the importance of lower-tropospheric moisture accumulation prior to the expansion of organized deep convection. The authors conclude that MERRA constitutes a very useful representation of intraseasonal variability that will support a variety of studies concerning radiative–convective–dynamical processes and will help identify pathways for improved moist physical parameterization in global models.

1. Introduction

Intraseasonal variability (ISV) of tropical convection (e.g., Madden and Julian 1972, 1994; Lau and Chan 1986; Stephens et al. 2004) remains a compelling and enigmatic science target for several reasons. First, the role of various physical processes that govern organization of tropical convection remains unclear. Among the key mechanisms that have been invoked as crucial to the existence, organization, and pacing of ISV are variants of wave-conditional instability of the second kind (wave-CISK; Lau and Peng 1987; Wang and Rui 1990; Salby et al. 1994), stochastic forcing by tropical (Salby and Garcia 1987), and extratropical (Blade and Hartmann 1993; Compo et al. 1999) sources, and an array of air–sea interaction processes controlling lower-tropospheric moisture (Emanuel 1987; Neelin et al. 1987; Blade and Hartmann 1993; Kemball-Cook and Weare 2001; Raymond and Fuchs 2009). Second, even acknowledging recent novel advances in climate modeling (e.g., Khaireoutdinov and Randall 2001; Miura et al. 2007), there has been rather modest improvement in simulating intraseasonal variability in global models dependent upon parameterized convection (Lin et al. 2006; Kim et al. 2009). A variety of theoretical studies (e.g., Raymond 2001; Raymond and Fuchs 2009; Majda and Stechmann 2009) and global numerical modeling experiments (Wang and Schlesinger 1999; Maloney and Hartmann 2000; Thayer-Calder and Randall 2009; Frierson et al. 2011) have pointed to fundamental problems with convective parameterization closure and the way these schemes interact with moisture. The issues surrounding ISV are at the very heart of the larger

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question regarding the ability to parameterize moist convection in global models. Climate variability on these time scales therefore represents an untapped source of predictability that could be realized if improved understanding of the key processes and their representation in global models could be obtained. Related to this parameterizable uncertainty is the degree to which improved data assimilation methodologies and new observations, particularly from satellite remote sensing, can overcome model physics deficiencies and provide depictions of ISV useful in advancing our understanding. Finally, the degree to which ISV in convective intensity and organization is important in enabling the quasi-steady-state energy balance of the tropics as a whole remains uncertain. Even though ISV in cloudiness and convective organization is localized primarily in the Eastern Hemisphere, humidity, temperature, and momentum signals extend globally around the tropics (Salby and Hendon 1994; Bantzer and Wallace 1996; Spencer et al. 2007). How these near-global adjustments figure into tropical energy balance appears to have received little attention.

Two of these points raised here provide motivation for this paper. To what degree can a state-of-the-art reanalysis capture tropical ISV? Here we focus specifically on the Modern-Era Retrospective analysis for Research and Applications (MERRA; Rienecker et al. 2011). More specifically, we ask what insights can MERRA, along with other observations, provide regarding the role of ISV in near-global radiative–convective adjustment and energy balance? Our focus therefore is not just on the state variables, but also on the energy fluxes. As noted by Kalnay et al. (1996) precipitation, turbulent fluxes, and cloud radiative effects derived from reanalyses possess greater uncertainty because they are strongly influenced by model physics.

Despite persistent challenges regarding the multiscale aspects of convective parameterization, global reanalysis efforts, whereby dynamical atmospheric models are constrained by the assimilation of diverse observations, have become increasingly useful to ISV studies, primarily by providing accurate representations of wind, temperature, and humidity (e.g., Shinoda et al. 1998; Annamalai et al. 1999; Kemball-Cook and Weare 2001; Jiang et al. 2011). Advances in data assimilation methods as well as a wealth of new data sources from satellite platforms have improved not only state variables, but also quantities like precipitation (e.g., Wang et al. 2012).

A unique aspect of the MERRA dataset is availability of the analysis increments that constrain state variable evolution. All reanalyses must blend noisy and incomplete observations with less-than-perfect model physics. Through the incremental analysis update (IAU) procedure (Bloom et al. 1996), the analysis correction is applied to the forecast model gradually through an additional tendency term in the model equations during the corrector segment of the analysis cycle. This methodology minimizes spin-up–down issues with fluxes and enables closed budgets of model variables. With MERRA in particular, the representation of the MJO in the assimilating model in free-running mode is extremely weak (Kim et al. 2009). Mapes and Bacmeister (2012) point out that this fact, coupled with the availability in the dataset of the corrections made to the evolving state variables (temperature, wind, and humidity), enables insights to be gleaned regarding model physics biases.

The vast majority of investigations of ISV, particularly in models, have focused on the mechanisms governing horizontal and vertical structure, phase speed, and growth rate. Our investigation takes a different tack and addresses the rectification of various fluxes on the net heat and moisture balance over the tropical oceans. Bantzer and Wallace (1996) dissected the strong linkage between MJO-like precipitation variations across the Indo-Pacific warm pool and resulting lagged signals of tropics-wide deep tropospheric temperature response using the Microwave Sounding Unit (MSU) channel data. In a study of ISV at Manus Island in conjunction with field studies conducted by the Department of Energy Atmospheric Research Measurement Program (DOE ARM), Lin and Mapes (2004) determined that top-of-the-atmosphere (TOA) fluxes during the convectively active phases of composited intraseasonal events resulted in systematic anomalous net loss of energy to space by the atmosphere–ocean system. Spencer et al. (2007) explored the systematic coupling on intraseasonal time scales between tropical mean deep convection, TOA fluxes, and tropical mean atmospheric temperature variations. They noted the reduction in TOA net radiative fluxes over the global tropical ocean domain following the enhanced convective phase. These results suggest a potentially significant role for intraseasonal convective events in mediating variations in the tropical energy balance through various feedbacks involving the radiative and turbulent fluxes. Whether these feedbacks have any parallels to mechanisms at play in externally forced climate change is highly uncertain; variations in external radiative forcing by CO₂ are essentially irrelevant on intraseasonal scales. Still, the nonsteady-state behavior of tropical or global mean atmospheric heat content, TOA radiative fluxes, and tropical precipitation remains ill understood.

Here we seek to determine the degree to which intraseasonal variability in MERRA replicates key observations when the fluxes are integrated to the tropical domain. How consistent are precipitation, radiative, and
of vertically integrated or single-level data (products \texttt{tavg1\_2d\_slv\_Nx}, \texttt{tavg2\_2d\_int\_Nx}, and \texttt{tavg1\_2d\_rad\_Nx} available at a spatial resolution of $\frac{1}{2}^\circ$ latitude $\times \frac{1}{2}^\circ$ longitude). Three-dimensional instantaneous assimilated virtual air temperature and geopotential height data on 42 pressure levels were also obtained (\texttt{inst3\_3d\_asm\_Cp}). Of principal interest are the mean tropospheric temperature (TT) taken as the simple average of the virtual temperature at 850-, 700-, 500-, and 300-hPa levels, column-integrated water vapor (CWV), precipitation as the sum of the grid scale, convective and anvil components, various TOA and surface radiative fluxes, turbulent fluxes of latent and sensible heat at the surface, and near-surface meteorology (10-m temperature, specific humidity, and wind speed). All quantities were acquired as daily means on a $1.0^\circ \times 1.25^\circ$ latitude $\times$ longitude grid via an online subsetter.

### b. Other satellite and reanalysis data

A diverse collection of products has been used to diagnose and validate MERRA. Tropospheric air temperature estimates come from the 53.596-GHz channel 5 of the Advanced Microwave Sounding Unit (AMSU-A) flying on the \textit{National Oceanic and Atmospheric Administration-15} (NOAA-15) polar-orbiting satellite. These measurements, henceforth $T_{\text{CH5}}$, are representative of deep layer tropospheric means with peak weighting near 600 hPa (Christy et al. 2003). Daily ascending and descending swath data available at 1.0$^\circ$ resolution from the Global Hydrology Resource Center (http://ghrc.nsstc.nasa.gov) were regridded and averaged to make daily maps at 2.5$^\circ$ resolution.

The objectively analyzed air–sea fluxes for the global ocean (OAFlux) dataset (Yu and Weller 2007; Yu et al. 2008) uses a variational objective analysis technique with optimal weighting based on buoy and ship observations to combine satellite and reanalysis data. With these analyses, fluxes are derived using the Coupled Ocean–Atmosphere Response Experiment (COARE) bulk flux algorithm 3.0 (Fairall et al. 2003). From this dataset we took daily values of latent (LHF) and sensible heat fluxes (SHF) as well as 2-m specific humidity and temperature, 10-m wind speeds, and SST.

TOA and surface radiative fluxes produced as part of the National Aeronautics and Space Administration (NASA) Global Energy and Water Cycle Experiment, Surface Radiation Budget Project (GEWEX SRB) were obtained from the NASA Langley Research Center Atmospheric Sciences Data Center. Daily all-sky and clear-sky products available at 1.0$^\circ$ latitude $\times$ longitude resolution from the \texttt{SRB\_REL3.0\_LW\_DAILY} and \texttt{SRB\_REL3.0\_SW\_DAILY\_UTC} collections were used. These derived fluxes use cloud retrievals from the
International Satellite Cloud Climatology Project (ISCCP; Schiffer and Rossow 1983; Rossow and Schiffer 1999) and the atmospheric temperature and water vapor profiles from the GEOS Data Assimilation System, version 4 (Bloom et al. 2005). The primary shortwave (SW) algorithm is adapted from Pinker and Laszlo (1992) and the longwave (LW) algorithm is an adaptation of Fu et al. (1997).

Precipitation data are from two sources. The Tropical Rainfall Measuring Mission (TRMM) 3G68 dataset is a 0.5°, gridded product with hourly resolution daily files. We extracted the TRMM version 6, 2A25 (Iguchi et al. 2000), which are rain estimates that are based on the TRMM Precipitation Radar. Because precipitation is a very granular quantity with sensor specific biases, we also use the TRMM Multisatellite Precipitation Analysis, 3B-42 (Huffman et al. 2001). This product intercalibrates the TRMM Microwave Imager (TMI), the Special Sensor Microwave Imager (SSM/I), the Advanced Microwave Scanning Radiometer (AMSR), and AMSU passive microwave estimates using a joint TRMM radar–TMI product (2B31) and is subsequently used to calibrate additional infrared brightness temperature observations from geostationary satellite. The 0.25° × 0.25°, 3-h resolution native product was averaged to 2.5° × 2.5° daily means for this study. Other TRMM-related data include column-integrated water vapor from the TRMM TMI sensor (Hilburn and Wentz 2008).

Cloud fractional cover is taken from the Moderate Resolution Imaging Spectroradiometer (MODIS) collection 005 MOD08_D3 dataset. Cloud optical thickness counts at various cloud-top pressure bins were used to generate cloud fractional coverage at daily resolution on the native 1.0° latitude × longitude grid. MERRA cloud fraction as used by the model’s radiative parameterization is available in layers surface to 700, 700–400, and above 400 hPa. Since precipitating convection dominantly affects the upper two layers, we focus on cloud cover above and below 700 hPa. Accordingly, the MODIS cloud fraction was aggregated to those layers.

c. Compositing methodology

The compositing strategy used here is very similar to that of Spencer et al. (2007) whereby indexing of intraseasonal events is defined in terms of the tropospheric temperature averaged over the tropical ocean band (20°N–S); all other variables and fluxes are referenced in time to dates of extreme temperature events. The AMSU-A channel-5 temperature data were averaged over the tropical band as a daily time series. A daily resolved annual cycle, smoothed by a running 21-day filter was then removed and a discrete Fourier transform filter was applied to this anomaly time series to retain only frequencies between 20 and 90 days. A running seeker function was used to find dates of maxima and minima in the filtered time series; extrema larger than 1.0 standard deviation were chosen as reference dates. Because of the quality of the MERRA temperature assimilation, the 20–90-day filtering as applied to MERRA data yields almost identical dates. The results of this process on tropospheric temperature can be seen in Fig. 1 where the MERRA filtered data are shown with the AMSU-A data with just the annual cycle removed. Elimination of lower-frequency interannual variability by the bandpass filter is readily seen. This procedure differs somewhat from Spencer et al. (2007) who chose their dates based on the strongest 15 positive intraseasonal events that were taken from unsmoothed daily anomalies after the smoothed annual cycle was removed. Our approach has been designed to eliminate synoptic variability in the selection of dates and more accurately identify the dates of the intraseasonal temperature maxima and minima. For all other datasets, daily anomalies at each grid point were constructed by removing their respective annual cycles, which were similarly defined as above. A high-pass Fourier filter to retain signals shorter than 90 days was applied to the anomalies at each grid point in order to remove interannual signals, but all frequencies greater than .05 day⁻¹ were retained. Thus, we have a measure of
how the intraseasonal signals rise above the synoptic “noise” in our composite. There were 39 positive and 42 negative events whose amplitude exceeded the 1.0 standard deviation threshold. These days, which define day lag 0 for each event, then served as a reference day to build the composites. Averages for the positive group minus the negative group, divided by 2, were then constructed for all quantities.

3. Variability of key quantities

Time series composites at various daily lags between plus or minus 30 days relative to the tropospheric temperature maximum (Fig. 2) emphasize the net intraseasonal signals integrated over the tropical ocean domain. By definition the tropospheric temperature, MERRA TT or AMSU TCH5, maximizes at lag 0 days and the similar amplitude and structure between the two is expected given that the former assimilates passive microwave brightness temperatures directly. TMI CWV, though not assimilated, agrees well with MERRA CWV since the reanalysis is constrained by assimilation of similar SSMI and AMSU-A passive microwave frequencies that respond strongly to water vapor overburden. Perhaps the most important comparison in Fig. 2 is the agreement between MERRA precipitation and that from the two TRMM estimates. The 3B42 product has much better space–time sampling characteristics (wider swath and multi-instrument measurements; Huffman et al. 2001) while the 3G68 is a direct radar reflectivity measurement of the precipitating drops, but only over a 240-km swath width (Iguchi et al. 2000). Their difference provides a measure of observational uncertainty. In the assimilation, precipitation is essentially the result of model physics, especially the convective parameterization, responding to the evolution of state variables and the increment forcings of moisture, heat, and momentum. As in the observations, MERRA precipitation leads tropospheric mean temperatures by about 7–8 days, but the temperature wave peak is very flat and starts to saturate at about 5 days after the precipitation maximum. This phase relationship is consistent with observational studies (Milliff and Madden 1996; Bantzer and Wallace 1996) as well as numerical modeling evidence (Jin and Hoskins 1995) showing role of Kelvin waves in rapidly dispersing an impulse of convective heating around the tropical domain. The amplitude of the MERRA precipitation signal, though smaller than that of TRMM, has been shown to be improved over previous reanalyses (Wang et al. 2012). For example, Shinoda et al. (1999) found that in the reanalysis produced by the National Centers for Environmental Prediction–National Center for Atmospheric Research (NCEP–NCAR) reanalysis, MJO-related precipitation variance was about one-half of that observed. We note also that 3B42 precipitation phase is slightly ahead of the 3G68 estimate and also larger in amplitude, signifying the level of uncertainty in precipitation estimates.

The propagation and spatial organization of the MERRA and TRMM 3B42 precipitation are provided in Fig. 3 (left panel) using pentad-averaged values centered on various lags. TRMM 3B42 values contoured over the shaded MERRA quantities show excellent spatial agreement, though occasionally the observations can be up to 20% larger. Positive precipitation anomalies develop first in the western Indian Ocean at lag ~20 days with negative anomalies in the western Pacific and eastward along the ITCZ. Positive anomalies intensify over the Indian Ocean and spread eastward and southward along the SPCZ over the next 10 days. This precipitation pattern migrates northward and weakens substantially by lag 5 days. Negative anomalies again become established in the western Indian Ocean by lag 0 days and strengthen into a dominantly negative pattern.

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1 Although some TRMM precipitation observations are assimilated, they are severely constrained in amplitude and have been found to have little effect on the resulting precipitation fields.
FIG. 3. (left) MERRA (shaded) and TRMM 3B42 (contoured) precipitation anomalies (mm day$^{-1}$) at (top to bottom) lags $-20$ to $+10$. Lag 0 is centered on date of maximum tropospheric temperature. All quantities are 5-day means. Contour increments are 0.5, 1.5, and 2.5 mm day$^{-1}$ with positive (negative) anomalies dark (light) green. (right) MERRA SST ×10 (shaded) and mean tropospheric temperature (contoured) anomalies at (top to bottom) lags $-20$ to $+10$. Tropospheric temperature contours are 0.05°, 0.15°, and 0.25°C with positive (negative) anomalies contoured black (gray).
by lag 10 days. Note also the small but distinct precipitation center over the tropical Americas that is out of phase with activity over the warm pool. This may be evidence of the large-scale dynamical connections that extend beyond the immediate precipitation zones across the Indo-Pacific warm pool. It is important to note that since we have not stratified by season, the composite events include the eastward-propagating MJO as well as northward-propagating precipitation patterns associated with intraseasonal variability of the Asian monsoon trough. The presence of the northward propagation in Fig. 3 suggests that sampling of this latter mode dominates in our dataset.

SST signals in Fig. 2 have much smaller amplitude relative to the tropospheric temperature and the two quantities are anticorrelated. Evolution of the mean SST spatial structure (Fig. 3, right) tends to be very regionally focused, particularly over the warm pool. SST anomalies in the equatorial Pacific also tend to be offsetting in sign with tropospheric temperature that generally exhibits much broader scale. These relationships are consistent with the importance of SST “hot spots” developing prior to convective flare-ups as surface SW heating maximizes and LHF is weakest (Waliser 1996). The “flatness” of the tropical tropospheric temperature field reflects the effectiveness of gravity waves in dispersing energy away from convectively heated regions due to the smallness of the Coriolis parameter in the tropics.

Returning to the time series quantities in Fig. 2, a subtle but potentially important difference between the TRMM observations and MERRA is apparent. During the buildup of CWV (lag −25 to lag −15 days) the onset of MERRA precipitation is a few days too early compared to that of TRMM. The result is a shorter lag between the buildup of CWV and precipitation compared to that observed. The importance of a period of increasing low-level moisture to support convection has been widely noted observationally (Kemball-Cook and Weare 2001; Kiladis et al. 2005; Agueldes et al. 2006; Benedict and Randall 2007) and is a crucial mechanism in the discharge and recharge hypothesis (hereafter, D/R) argued as central to ISV by Blade and Hartmann (1993). This earlier rise in MERRA precipitation suggests that the Relaxed Arakawa–Schubert (RAS) convective parameterization (Moorthi and Suarez 1992) responds somewhat too quickly to the moisture recovery in the subcloud layer and fails to let moisture build upward sufficiently in the lower troposphere and elevate convective available potential energy (CAPE). This problem is common to most all convective parameterizations in climate models (Wang and Schlesinger 1999; Kim et al. 2009; Thayer-Calder and Randall 2009).

Approximately 80%–90% of the variability in $\theta_e$ for MERRA and OAFlux (Fig. 2) is due to moisture. The $\theta_e$ in OAFlux lags the CWV but is almost in phase for MERRA. Observed near-surface $\theta_e$ variability is also slightly smaller in MERRA. Whether this phase difference reflects the lack of convective downdrafts drying the subcloud layer in MERRA is not yet clear. The $\theta_e$ time series also leads the tropospheric temperature by 5 days in the observations and perhaps by 3 days in MERRA.

4. Top-of-the-atmosphere fluxes and the role of clouds

Intraseasonal variations in net TOA flux (Fig. 4, left) comprise significant contributions from SW and LW components of cloud radiative effect (CRE). Here CRE is defined for shortwave as TOA net all-sky SW minus net clear-sky SW, and CRE for longwave as TOA net clear-sky LW minus net all-sky LW. MERRA and SRB both show anticorrelated behavior in SW and LW CRE, but these components are not directly opposite in phase, particularly for MERRA. As MERRA precipitation amounts decline after lag −7 days, LW CRE also weakens and eventually becomes an anomalous loss after lag −5 days. But the SW CRE anomalous loss maintains its intensity, turning positive only after about lag 2 days. This behavior results in a minimum value of net TOA absorbed by the tropical regions centered near lag −2 days. Though cloud-induced reduction in net energy input to the tropics dominates the net radiative flux changes, the clear-sky component is not negligible, contributing roughly one-third of the net radiative effect. The clear-sky contribution is due almost entirely to the increased LW emission by the warmer atmosphere centered near lag 0 days. (A note of caution is warranted here since the SRB clear-sky fluxes are determined by temperature and moisture taken from the GEOS-4 assimilation—a direct predecessor of MERRA.) In terms of differences, the SRB-observed fluxes exhibit less out-of-phase behavior and produce a net anomalous TOA loss of energy that is strongest near lag 2 days, roughly 4 days later than does MERRA. This difference is due principally to the MERRA LW CRE signal maximizing almost 10 days earlier than is observed in SRB. Since the phasing of MERRA’s buildup of precipitation is only a few days earlier than observed, one plausible explanation is that MERRA convective clouds may be too deep (cold) at this stage. Populations of convective clouds are known to lack sufficient variability in their depth (Tokioka et al. 1988; Wang and Schlesinger 1999; Frierson et al. 2011).

In Fig. 5 the spatial distribution of MERRA TOA flux components and their net at lag 0 days shows reasonably
good agreement with SRB everywhere, except perhaps in the eastern Pacific. MERRA fluxes in the Indo-Pacific region also tend to be weaker than those of SRB. Much of this effect on the area-mean value is mitigated by the opposing signs of flux north and south of the ITCZ. These radiative effects can be traced to variations in cloud amount patterns (Fig. 6). MERRA low cloud amount anomalies are too negative along 15°N, but too positive along 15°S. The larger amplitude of high plus middle fractional coverage in MERRA compared to SRB suggests that differences in high clouds and their shielding of low clouds may explain these problems. For both MERRA and observations, the spatial distributions of high plus middle cloud anomalies in Fig. 6 correlate (anticorrelate) well with the SW (LW) component anomalies in Fig. 5. Strong cancellation of the SW and LW effects of higher clouds thus also extends to intraseasonal scales.

There is evidence that low-level cloudiness variations play a significant role, too. In both MERRA and MODIS estimates of cloud fractions, high and middle cloudiness associated with deep convection peaks earlier than does low cloud fraction (Fig. 4). Note that the net CRE effect for MERRA and MODIS becomes negative when low-level cloud abundances become positive near lag 210 days. It is also evident for each that SW CRE tends to be

![Fig. 4.](image)

**Fig. 4.** (left) Time series of intraseasonal composite TOA radiative flux anomalies (W m$^{-2}$; mean annual cycle removed) averaged over the tropical ocean domain as a function of lag from day 0: (top) SRB and (bottom) MERRA. Negative lags denote time prior to maximum tropospheric mean temperature. (right) Cloud fraction anomalies (percent): (top) MODIS and (bottom) MERRA. See text for additional details on cloud forcing definitions and determination of high, middle, and low clouds.

![Fig. 5.](image)

**Fig. 5.** Lag 0 composite anomaly maps of TOA LW, SW, and Net radiative fluxes (W m$^{-2}$) for GEWEX SRB and MERRA. Note change in scale for net TOA flux.
anticorrelated with the total cloud fraction, whereas the LW CRE follows high plus middle cloud fraction. Low clouds tend to have weaker effects on the OLR since their effective cloud-top temperatures are typically much closer to the effective radiating temperature of the tropospheric air compared to that of deep convective clouds. Low clouds can also have high albedos similar to those of thick high clouds. Note that although the phasing of low clouds versus high plus middle clouds is captured reasonably well by MERRA (Fig. 4), its larger phase difference results in a smaller variation of net cloud fraction compared to MODIS. This tends to weaken the TOA net SW CRE. But the early growth of MERRA high plus middle clouds into a cold atmosphere (Fig. 4) means that the LW CRE peaks earlier than is seen in SRB. These two offsetting factors contribute to the agreement in net TOA flux between MERRA and the GEWEX SRB observations.

The net CRE is obviously sensitive not just to the amount of cloudiness present, but also to the height (temperature) and optical properties of the clouds. To assess the intraseasonal CRE of cloud types in MERRA we have stratified TOA net cloud radiative fluxes (SW CRE and LW CRE) according the fractional presence of high plus middle versus low clouds (Fig. 7). High and middle clouds are those defined to have tops above 700 hPa and low clouds are those with tops below 700 hPa. For both MERRA and MODIS, cloudy grid points on each of the 61 composite days were classified as a high plus middle scene if the low cloud fraction cloud anomaly did not exceed 0.05. Similarly, low cloud scenes were required to have less than 0.05 coverage of high plus middle clouds. The plot shows that for MERRA and the observations, the net effect of all cloud types is to produce a net anomalous cooling effect to the planet at TOA. Low clouds, as expected, provide the strongest cooling, over 0.8 W m\(^{-2}\) per percent cloud cover for both datasets. High and middle clouds too provide net cooling, though the MODIS estimate (−0.38 W m\(^{-2}\) per percent unit cloud fraction) is much stronger compared to MERRA (−0.13 W m\(^{-2}\) per unit cloud fraction). Cloud amounts and optical properties (e.g., particle effective radius and its effect on optical thickness) are strongly tunable parameters in climate models, but they are typically adjusted to obtain reasonable TOA climatological balance, not in terms of short-term variations on intraseasonal and shorter scales.

This enhanced net cooling effect associated with the effects of enhanced deep convective precipitation has been found other studies. Cess et al. (2001) in a study of the 1997/98 El Niño event noted the shift to shallower precipitating cloud tops and a net reduction in absorbed radiation over the western Pacific in conjunction with the reduced large-scale ascent there. On intraseasonal scales, Lin and Mapes (2004) found a systematic reduction in net TOA-absorbed energy over the western Pacific associated with the passage of enhanced deep convection. Spencer et al. (2007) showed this intraseasonal variability to be a robust property of heat balance fluctuations for the tropical ocean domain as a whole.

5. Energy budgets

In this section the fluxes and state variables are used to construct water and energy budget variations averaged over the tropical oceans (20°N–S). The vertically integrated moisture and dry static energy budgets are

\[
L \frac{\partial \tilde{q}}{\partial t} = L(E - P) - \nabla \cdot \tilde{q} \tilde{V} + \tilde{L} \tilde{q}_i \text{ana} \tag{1}
\]
and

\[
\frac{\partial \hat{s}}{\partial t} = LP + S_o + R_{TOA} - R_{SFC} - \nabla \cdot \vec{s} V + H_{ana},
\]

respectively, where \( \hat{q} \) and \( \hat{s} = \hat{c}_p T + \hat{\varphi} \) are the column-integrated specific humidity (CWV) and dry static energy (enthalpy plus geopotential energy) with the hat symbol denoting mass-weighted vertical integral through the atmosphere. Here LE and LP are the surface latent heat flux and precipitation in energy units with \( L \) being the latent heat of vaporization–condensation. The \( S_o \) is sensible heat flux and \( R_{TOA} \) and \( R_{SFC} \) are net radiative heat flux downward at TOA and the surface, respectively. Vertically integrated horizontal flux convergence of specific humidity and dry static energy are represented by \(-\nabla \cdot \vec{q} V\) and \(-\nabla \cdot \vec{s} V\). The terms \( Lq_{ana} \) and \( H_{ana} \) apply to MERRA budgets only and represent vertically integrated moisture and dry static energy increments, respectively.

Two points are in order regarding construction of the observational budgets. For dry static energy storage, the gravitational potential energy component has been estimated by using the hypsometric equation, assuming the temperature anomaly is uniform throughout the troposphere, and neglecting surface pressure variations. CWV are from Hilburn and Wentz (2008). One-sided time differences of CWV and \( s \) were used to obtain the tendencies. Though this assumption is crude, it is probably not too bad since the MERRA and observed \( s \) storage anomalies track each other well. Second, the observed moisture and flux convergence terms have each then been solved as residuals needed to balance the budget equations. For MERRA these assumptions are not needed since we have the archived terms. We have neglected MERRA budget contributions from numerical “fixer” terms.

### a. Atmospheric budgets

Components of \( \hat{q} \) and \( \hat{s} \) budgets (Fig. 8) show strong similarities between observed quantities and MERRA-assimilated counterparts. In the moist period prior to maximum tropospheric temperature (lag 0 days), precipitation obviously serves as a significant sink of moisture and a source of dry static energy. LHF by itself cannot supply the moisture needed for precipitation and in situ storage; thus, the import of moisture is a significant budget term. LHF and moisture flux convergence are of similar magnitude and in phase—more strongly in MERRA than in the observations. Additional analysis (not shown) has indicated that the correlation (cor) between these LHF variations and those of wind speed is quite high (tropical ocean area-average cor = 0.80; Indo-Pacific ocean area-average cor = 0.88). In MERRA the contribution by the moisture increment is significant, acting as a source of moisture as precipitation transitions to becoming a strong sink (before lag \(-10 \) days) and likewise serving as a drying influence during the period of precipitation decay (lag \(-10 \) to lag 10 days).

For the \( \hat{s} \) budget we have displayed the combined net radiative and sensible heat fluxes since the latter is a very small contribution. The net radiative effects are almost in quadrature with the precipitation, providing a net atmospheric heating as precipitation increases and a sink as precipitation anomalies are rapidly decreasing. In contrast, Lin and Mapes (2004) found an in phase
relationship between atmospheric radiative heating and precipitation. This difference appears due to our tropical ocean averaging framework in contrast to their single location (Manus Island) diagnostics. Strong OLR reductions dominate the net atmospheric radiative changes directly collocated with deep convective systems. However, for our tropical domain, clear-sky fluxes in non-convecting regions play a considerable role. The SRB net TOA LW cloud forcing in Fig. 4 is, in fact, in phase with TRMM precipitation in Fig. 2, suggesting that in cloudy, convecting regions, the behavior is similar to that of Lin and Mapes (2004). For MERRA, this explanation does not seem to hold; LW cloud forcing still leads maximum precipitation by about 10 days. The reason for this is not yet clear, but again we suspect that as precipitating convection anomalies become positive, the clouds produced by detrainment from the convective parameterization are too deep and cold compared to reality. The \( x \) increments vary systematically, providing a sink of energy during the buildup to peak precipitation and a source as precipitation declines. Interestingly, this means that the net effect of increments on moist static energy (vertically mass-integrated \( c_p T + g z \); \( \frac{dh}{dt} \) ) is rather small compared to LHF and net radiative fluxes. We explore these increment forcings in more detail in section 6.

**b. Surface energy balance**

Ocean surface energy fluxes have significant amplitudes on intraseasonal time scales (Shinoda et al. 1998; Woolnough et al. 2000) and exert considerable control on SST variations. Figure 9 shows time series of the net surface heat flux into the ocean as a function of lag averaged over the 20°N–S domain. Also shown are the radiative SW and LW components and LHF. The strong

![Diagram](image-url)
similarities between MERRA and the observations include the dominance of LHF, offsetting SW and LW effects, and phasing of the net flux. Though the behavior of LW components is similar, MERRA SW forcing variations are weaker than those of SRB and, as expected, strongly correlated to the TOA SW in Fig. 4. Net heat flux out of the ocean is in quadrature with SST (Fig. 2) and also varies closely with precipitation, leading by about two days.

Of course, the spatial averaging in Fig. 9 eliminates important pattern structure and propagation information concerning LHF and net radiation at the surface; to see those relationships we examine MERRA lag maps in Fig. 10. LHF increasingly extracts energy from the ocean as precipitation increases and is predominantly positive everywhere (lag \(-20\) to lag \(-10\) days), but especially in the subtropical western Pacific with maximum values focusing in subtropical regions adjacent to precipitation zones. Given the strong influence of wind speed on LHF variations and the fact that LHF \(q\) import and \(s\) export phase similarly in time (Fig. 9) it is likely that LHF evolves in part as a dynamical circulation response to the heating. Variations in surface net radiation are much more focused in the northern Indian Ocean, over the warm pool, and along the South Pacific convergence zone and strongly correspond to the evolution of cloud cover generated by convection. The strong control on SST evolution exerted by these two fluxes (surface net radiative flux minus LHF) is shown in Fig. 11 where we have correlated SST tendency with net surface energy flux (into the ocean). Correlations exceed 0.7 over broad areas of the Indo-Pacific warm pool. Observational counterparts to the MERRA patterns (not shown) are similar, but with larger amplitude radiative flux variations.

c. Interpretation–synthesis

Collectively, the TOA and surface fluxes and atmospheric budgets suggest the following picture: after initial moisture availability becomes sufficient over the Indian Ocean, convection amplifies, deepens, and precipitation
increases. Associated cloudiness increases that suppress surface insolation act in concert with elevated LHF to increase the removal of energy from the ocean. The LHF increases result partially as a dynamical response to the convective heating. Tropically integrated, these processes drive an ocean energy sink beginning by approximately lag $-20$ days. Gradual energy input to the atmosphere from increased precipitation and LW trapping peaks near lag $-8$ days. As convection propagates northeastward and begins to decay, net CRE associated with the evolution of the precipitation is essential in rejecting some of this energy to space. Decay of the heating-driven circulation reduces LHF, reduces cloudiness, and results in a net source of energy to the oceans by day zero (maximizing at lag 10 days). This sequence of events is broadly consistent with previous findings (e.g., Waliser 1996; Sobel and Gildor 2003; Stephens et al. 2004) that antecedent solar heating of the surface is an important factor in establishing anomalously warm SSTs in advance of convection. The spatial relationships between evolving SST and convection can be readily seen in Fig. 3.

Dynamical links to the energy fluxes are significant with an anomalously strong thermally forced circulation importing $\dot{q}$ and exporting $\dot{s}$. Net export-import of moist static energy $\dot{h}$ associated with this thermally-driven circulation is much smaller (a noisy quantity in our budgets). As convection subsides the thermally direct circulation becomes anomalously weak. As noted in section 5 a strong correlation exists between LHF and the intensity of this circulation. Viewed from the context of $\dot{h}$ variations, the net transfer of energy from the ocean to the atmosphere results in both elevation of $\dot{h}$ as well as some loss of energy to space. This cycle then reverses with a net energy input at TOA, a reduction of atmospheric $\dot{h}$, and resupplying of the ocean mixed layer. Though the net lateral boundary flux of $\dot{h}$ is small in either phase, the individual $\dot{q}$ and $\dot{s}$ fluxes are large and strongly coupled to the circulation driven by the heating. This storage-release cycle of energy in which the ocean accumulates energy, turbulent fluxes recharge the atmosphere, and convective-radiative adjustment processes reject a portion of this energy to space is closely related to the D/R mechanism articulated by Blade and Hartmann (1993). The latter is focused over the Indo-Pacific warm pool and the Indian Ocean in particular and refers explicitly to deepening atmospheric moisture and buildup of CAPE localized to that region. The storage-release cycle described here is a tropically integrated signal that involves the ocean and atmospheric reservoirs and also extends to the radiative fluxes at the surface and TOA interfaces. Both the D/R sequence as well as the areal-integral format used here emphasize the “lagged capacitance” behavior in that energy is stored (in the ocean or in the atmosphere) and then released, in contrast to a steady-state configuration.

The overall role of MERRA heat and moisture increments compared to the physical fluxes is significant. They are only slightly smaller than the net atmospheric radiative heating and evaporation terms. Prior to the development of convection, the increments are an important moisture source and they provide a heat sink prior to the maximum in precipitation. As convection subsides, their roles are reversed with drying and heating contributions. This behavior is consistent with the tendency for convective parameterizations to link much too strongly with CAPE production, short-circuiting the D/R sequence (Thayer-Calder and Randall 2009).

6. Behavior of analysis increments

The consistency between MERRA quantities and various observed counterparts thus has to be viewed in context with the domain-averaged importance of the analysis increments (Fig. 8) Ideally, one would want the increments to be randomly distributed quantities, indicating that the model physics had no mean bias. Since the
previous budget discussion shows that this is not the case, here we explore some significant regional details. Figure 12 shows the vertically integrated increments of enthalpy and water vapor, dhdt-ana and dqvdt-ana, respectively. Implicit in the water and heat balance are also the effects of \( u \)- and \( v \)-component momentum forcing. Also shown in Fig. 12 is the divergence of the \( u \)- and \( v \)-wind component increments at 975 hPa, dgradV\(_{975}\)dt-ana. The 975-hPa level was taken as representative of the near-surface conditions. The maps are at various lags with contours of precipitation overlaid (omitted from dqvdt-ana for clarity). Also provided in Fig. 13 are fields of correlation between these increments and precipitation calculated with the 61-day time series. These plots give a more quantitative measure of the importance of the increments to the largest term in the budgets, one that is a direct response to model physical parameterizations.

The dqvdt-ana distribution has a very granular structure, but is organized as well on large scales. The gradual transition from general moistening to drying tendency takes place from lag \(-20\) to lag \(0\) days in the tropical Pacific between \(15^\circ N-S\). In adjacent subtropical regions the trend is reversed. Over the Indian Ocean the structure and trend is less clear. But there is a surprisingly broad and systematic structure to the correlation with precipitation (Fig. 13). Positive correlations (moistening in the presence of precipitation, drying in its absence) dominate the tropics except over the western equatorial Pacific where negative correlations are found. Poleward of \(30^\circ\), drying is present in most locations. However, these correlations are quite weak with most being less than \(0.30\) in amplitude. In contrast, dhdt-ana exhibits a smoother behavior in Fig. 12 and a more coherent pattern of correlation with precipitation over the warm pool region (Fig. 13) with anticorrelations exceeding \(-0.50\) there and over other convecting regions. This may be due in part to the local dominance of the precipitation forcing in the \( s \) budget compared to the more complex nature of the local moisture balance between precipitation, moisture convergence, evaporation, and storage.

In addition to the direct effects of heating and moistening increments, there is the implicit effect of momentum field forcing. The (gradV\(_{975}\))dt-ana provides some sense of how forcing of low-level convergence is acting to alter

**FIG. 12.** Composite MERRA increment anomalies for (left) vertically integrated specific humidity, (middle) divergence of 975-hPa velocity, and (right) dry static energy from (top to bottom) lag \(-20\) to lag 10 days. Lags are 5-day means. Overlaid are precipitation anomalies (positive, solid; negative, dotted) contoured at intervals of 0.5, 1.5, and 2.5 mm h\(^{-1}\).
the accumulation of moisture, low-level vertical velocity, and rainfall. The negative correlation with precipitation anomalies over the tropics in Figs. 12 and 13 shows important systematic modification of the low-level circulation to increase the amplitude of precipitation anomalies. We repeated this analysis at 850 hPa (not shown) and noted a substantially weaker relationship to precipitation. This suggests that surface observations (buoys, ships, and scatterometer data) may be having a substantial impact on the analysis, but largely near the surface.

Given the weakness of MJO signals in the GEOS-5 assimilating model (Kim et al. 2009) it is no surprise that the systematic influence of the increments in Figs. 8, 12, and 13 is strong. While it is beyond the scope of this paper to isolate details of the model physics shortcomings our inference that aspects of the parameterized convective processes are problematic is consistent with many other studies (e.g., Wang and Schlesinger 1999; Bechtold et al. 2008; Thayer-Calder and Randall 2009). Like other parameterizations, the RAS routine used in GEOS-5 has difficulty in producing adequate moistening at lower-tropospheric levels without triggering precipitating convection; this is despite incorporation of a constraint on minimum entrainment similar to Tokioka et al. (1988) and a critical relative humidity threshold at the lifting condensation level. The behavior of the heat and moisture increments seen in this study largely echoes the results of Mapes and Bacmeister (2012). Their analysis uses a different strategy to isolate ISV signals and focuses on boreal winter periods, but they find similar patterns of increments moistening and cooling during the buildup of convection and drying and warming as convection subsides.
7. Discussion and conclusions

We have investigated tropical intraseasonal variability in MERRA from the standpoint of area-integrated signals. This perspective allows insights as to the importance of ISV to the overall tropical energy balance and its mechanisms by which the climate oscillates around its current quasi–steady state. Despite the fact that the GEOS-5 system and its predecessors have, like other models, a very limited representation of intraseasonal behavior, particularly in the context of the MJO (Kim et al. 2009), MERRA has successfully captured a reasonable amplitude response and the phase relationships between key flux processes. Not surprisingly, state variables of atmospheric temperature and moisture are well represented in MERRA on intraseasonal time scales and thus at least give the model physics good forcing with which to operate. Resulting MERRA precipitation anomaly patterns evolve in close agreement with those of TRMM products. While locally MERRA anomalies can be up to 20% smaller, the maximum amplitude of anomalies averaged over the tropical oceans for MERRA and TRMM are both near 0.2 mm h⁻¹. However, the analysis increments that enforce corrections to the state variables are significant relative to precipitation and other heat and moisture budget terms and provide evidence of systematic deficiencies in GEOS-5 moist physics.

Radiative forcing and the role of clouds are reasonably well simulated from the standpoint that they replicate the results of Spencer et al. (2007). The evolution of cloud cover in association with the enhanced convection is configured so that high, middle, and low clouds act to cool the planet. Low clouds and their appearance in the eastern Pacific play an important contributing role though in MERRA there may be some problems in representing these. MERRA total fractional variability is less than that detected by MODIS and the SW CRE less than that seen by the SRB data. In addition, evidence suggests that MERRA convectively produced clouds may deepen too soon as convection increases early in the phase when atmospheric moisture is recovering in the Indian Ocean. Cloud tops deeper than 700 hPa appear nearly 10 days ahead of similar cloud types seen by MODIS. One might expect that with recovering low-level moisture over the Indian Ocean collocated with cold tropospheric temperatures early in the recharge phase, parameterized convective updrafts in MERRA are most likely to be diagnosed as minimally entraining parcels, a situation likely to provide unrealistically deep clouds. The earlier phasing of the dominant LW warming effects combine with the weaker SW effects of high clouds to allow the net reduction in TOA net radiation after convection and precipitation to have peaked. It is worth noting that no cloud information is assimilated by MERRA and that its clouds are all produced by parameterized prognostic condensate at the grid scale and in dynamically inert anvils created by detrainment from the convective parameterization. Clouds in climate models are tuned, sometimes strongly, to yield climatological net balance at TOA; however, they are not tuned to replicate behavior on subseasonal time scales.

Increments of water vapor and heat that the assimilation uses to force the trajectory of state variables close to the observations assume an important role in MERRA (as similar forcing does in other reanalyses). Just prior to the maximum precipitation date (~lag = 7 days) the increments switch from net moistening and cooling to drying and heating over the tropics, integrated through the column. In general, regions of precipitation are always being cooled by the increments, though the association of precipitation and moistening is not as clear. Divergence of momentum forcing, though indirect in nature, is very important at the lowest model levels and is anticorrelated with precipitation. Results presented here are consistent with those of Mapes and Bacmeister (2012) who also find excessive heating and drying by RAS and evidence that the interaction of the RAS convective parameterization with moisture is suspect in terms of enabling lower-tropospheric moisture buildup characteristic of observed D/R behavior.

There is important communication between the ocean and atmosphere with energy being extracted from the ocean, moistening of the PBL, increased CWV supporting the buildup of convection, and eventual rejection of this heat deposited in the atmosphere by convection to space via TOA fluxes. As convection subsides and the atmosphere cools, the reduced LHF and the reduction of cloudiness enable recharge of the upper-ocean mixed layer. This sequence of events is consistent with D/R mechanism articulated by Blade and Hartmann (1993). Our analysis here links in the behavior of TOA fluxes, the importance of CRE in particular, and views the full sequence of events from a tropical–areal-mean perspective. We interpret this storage–release cycle simply as the way radiative–convective adjustment behaves in the intraseasonal band, owing to important constraints imposed by the existence of zonal SST asymmetries and the Indo-Pacific warm pool, which in have their origin in the configuration of landmasses and their influence on oceanic heat transports.

Finally, we note that these results seem largely consistent with those of Spencer et al. (2007) and that the differences are explainable. The smaller tropospheric temperature response found here (about half as large)
results primarily from our composite including a larger number of smaller events, but is also the result of filtering the AMSU-A and MERRA temperature for intraseasonal signals only. This filtering picks out dates of the intraseasonal extremes that may differ from dates where the synoptic signal remains embedded. These differences in methodology likely result in a somewhat longer lag between tropospheric temperature, CWV, and precipitation. While Spencer et al. (2007) have interpreted the TOA radiative flux response to $T_{CH5}$ during these events as a potential climate feedback diagnostically, our view is more circumspect. The D/R mechanism allows “capacitance” in the form of $\hat{q}$ and CAPE buildup as an integral part of the adjustment process. In contrast, convective adjustment in the face of external radiative forcing acts to equilibrate the ocean–atmosphere system and restore TOA balance. Furthermore, since the intraseasonal atmospheric temperature perturbation is an integral part of the adjustment process and contains no part due to any external forcing, the relationship of TOA radiative response to atmospheric temperature change likely says more about how the present climate oscillates about its preferred equilibrium than about the cloud radiative response to future greenhouse gas forcing.

In summary, MERRA has produced a very credible picture of intraseasonal variability as judged by comparisons with radiative fluxes and precipitation data that are largely independent of the assimilation. The MERRA project set out to improve quantitative depictions of weather to climate-scale hydrologic processes. Fields of precipitation and radiative fluxes analyzed here show that this has been achieved on intraseasonal scales. An important caveat though is the importance of the assimilation increments and their testimony to systematic shortcomings in modeled convective processes. Further efforts are needed to explore the processes highlighted here in greater detail. Stratification by season should help with the understanding of MJO versus monsoon-related variations as well as provide a clearer picture of how the seasonality affects the required increments. It may also offer clues as to the way forward in improving model physics. Finally, we also note that in support of the Year of Tropical Convection initiative (YOTC; Waliser and Moncrieff 2008), the NASA Global Modeling and Assimilation Office (GMAO) is producing 0.25° resolution integrations with a GEOS-5 version similar to that used for MERRA. Plans are also in place to repeat this assimilation with an improved coupled atmosphere–ocean system. Exploring intraseasonal signals in these assimilations in a case study framework should offer an important, additional perspective on issues raised here.

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