Intraseasonal Latent Heat Flux Based on Satellite Observations

SEMYON A. GRODSKY
Department of Atmospheric and Oceanic Science, University of Maryland, College Park, College Park, Maryland

ABDERRAHIM BENTAMY
Institut Francais pour la Recherche et l’Exploitation de la Mer, Plouzane, France

JAMES A. CARTON AND RACHEL T. PINKER
Department of Atmospheric and Oceanic Science, University of Maryland, College Park, College Park, Maryland

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ABSTRACT

Weekly average satellite-based estimates of latent heat flux (LHTFL) are used to characterize spatial patterns and temporal variability in the intraseasonal band (periods shorter than 3 months). As expected, the major portion of intraseasonal variability of LHTFL is due to winds, but spatial variability of humidity and SST are also important. The strongest intraseasonal variability of LHTFL is observed at the midlatitudes. It weakens toward the equator, reflecting weak variance of intraseasonal winds at low latitudes. It also decreases at high latitudes, reflecting the effect of decreased SST and the related decrease of time-mean humidity difference between heights $z = 10$ m and $z = 0$ m. Within the midlatitude belts the intraseasonal variability of LHTFL is locally stronger (up to 50 W m$^{-2}$) in regions of major SST fronts (like the Gulf Stream and Agulhas). Here it is forced by passing storms and is locally amplified by unstable air over warm SSTs. Although weaker in amplitude (but still significant), intraseasonal variability of LHTFL is observed in the tropical Indian and Pacific Oceans due to wind and humidity perturbations produced by the Madden–Julian oscillations. In this tropical region intraseasonal LHTFL and incoming solar radiation vary out of phase so that evaporation increases just below the convective clusters.

Over much of the interior ocean where the surface heat flux dominates the ocean mixed layer heat budget, intraseasonal SST cools in response to anomalously strong upward intraseasonal LHTFL. This response varies geographically, in part because of geographic variations of mixed layer depth and the resulting variations in thermal inertia. In contrast, in the eastern tropical Pacific and Atlantic cold tongue regions intraseasonal SST and LHTFL are positively correlated. This surprising result occurs because in these equatorial upwelling areas SST is controlled by advection rather than by surface fluxes. Here LHTFL responds to rather than drives SST.

1. Introduction

Latent heat flux (LHTFL) links air–sea heat exchange with the hydrological cycle. This evaporative heat loss term balances a significant portion of the surface heat gain due to solar radiation (da Silva et al. 1994). Satellite sensors can measure sea surface temperature (SST), near-surface winds, and humidity and thus provide data for estimating evaporation. Currently, several satellite-based global ocean latent heat flux products are available (e.g., Chou et al. 2003 and references therein). In this study we exploit the availability of a new global 16-yr (1992–2007) record of weekly satellite-based turbulent fluxes of Bentamy et al. (2008) to examine the observed geographic distribution of intraseasonal LHTFL and its role in air–sea interactions.

Most observational examinations of LHTFL focus on its behavior on monthly and longer time scales (e.g., da Silva et al. 1994; Yu et al. 2006). Recent studies of the
midlatitudes (Qiu et al. 2004) and tropics (Zhang and McPhaden 2000) have shown that intraseasonal variations of LHTFL associated with synoptic disturbances can alter SST by up to 1°C. Modeling studies (Maloney and Sobel 2004; Han et al. 2007) suggest that these SST variations may in turn organize intraseasonal atmospheric convection and thus provide an air–sea interaction mechanism for phenomena such as the 30–60-day Madden–Julian oscillations (MJOs; Madden and Julian 1972). Because LHTFL is proportional to evaporation, its intraseasonal variations contribute to variations of surface salinity, thus increasing the impact of LHTFL on surface density.

In the tropics, high temperatures and thus saturated humidity combined with significant synoptic variability such as the MJO lead to significant intraseasonal LHTFL. MJOs may be driven in part by the evaporation–wind feedback (Neelin et al. 1987). The MJO is a feature of all tropical sectors, although it is most pronounced over the eastern Indian Ocean and western Pacific Ocean in boreal winter and is strongly modulated by ENSO; it is characterized by strong 2–4 m s⁻¹ fluctuations of surface winds and precipitation (Araligad and Maloney 2008). As a result, it produces correlated fluctuations of both LHTFL and shortwave radiation (SWR) with amplitudes of 30–50 W m⁻² and is observed to cause 0.5°C fluctuations of SST (Krishnamurti et al. 1988; Shinoda and Hendon 1998; Zhang and McPhaden 2000). Moreover, recent research suggests that these intraseasonal fluctuations may actively interact with lower-frequency climate variations in the tropics, just as in the Pacific, where westerly wind bursts may trigger the evolution of the El Niño–Southern Oscillation (ENSO) cycle (McPhaden 2004).

Some of intraseasonal variability observed at low latitudes and sub-tropics is linked to intraseasonal variability of midlatitude pressure systems. In particular, Foltz and McPhaden (2004) have examined the intraseasonal (30–70 day) oscillations in the tropical and sub-tropical Atlantic and found their link to fluctuations in the strength of the Azores high. The sub-tropics and mid-latitudes are subject to synoptic meteorological forcing originating in the midlatitude storm systems. This additional variability has a strong seasonal component, amplifies in the cold season, and varies from year to year. In the Kuroshio extension region, Bond and Cronin (2008) have found that in late fall through early spring cold air outbreaks associated with synoptic events lead to intense episodes of LHTFL and sensible heat loss. Similar origins of intraseasonal variability are observed by Zolina and Gulev (2003) in the Gulf Stream region.

In summer and fall (when the ocean mixed layer shoals), cloud shading effects accompanying synoptic disturbances become important sources of intraseasonal flux variations. Based on experiments with a mixed layer model, Qiu et al. (2004) suggest that these summertime intraseasonal flux variations can induce SST variations with climatologically significant ±1°C amplitudes. This, together with other observational evidence, suggests significant contributions by LHTFL variability in the intraseasonal band to the state of the climate system. In this study we focus on geographical patterns of LHTFL, consistency with SST, interplay with incoming solar radiation, and modulation by longer period processes.

This study is possible because of several improvements to the climate observing system. Beginning in the early 1990s, a succession of three satellite scatterometers provides high-resolution surface winds. Brightness temperature estimates from the Special Sensor Microwave Imager (SSM/I) provide an estimate of relative humidity. When this is combined with estimates of surface temperature, it becomes possible to estimate LHTFL at weekly resolution (Bentamy et al. 2008). Clouds and aerosols, the main factors affecting SWR, are available from a variety of sensors flying in both geostationary and polar orbits (Rosen and Schiffer 1999; Pinker and Laszlo 1992). Finally, an array of more than 90 moorings distributed across all three tropical oceans (McPhaden et al. 1998) provides ground truth at high temporal resolution, which can be used to explore the accuracy of the remotely sensed estimates.

2. Data and method

This research is based on the recent update of weekly satellite-based turbulent fluxes of Bentamy et al. (2003, 2008). The three turbulent fluxes—wind stress (τ), LHTFL (Qₑ), and sensible heat flux (Q_H)—are estimated using the following bulk aerodynamic parameterizations (Liu et al. 1979):

\[
\frac{\tau}{\rho} = C_D \| \mathbf{u}_a - \mathbf{u}_s \| (\mathbf{u}_a - \mathbf{u}_s),
\]

\[
\frac{Q_E}{\rho L} = -C_E \| \mathbf{u}_a - \mathbf{u}_s \| (q_a - q_s),
\]

\[
\frac{Q_H}{\rho C_p} = C_H \| \mathbf{u}_a - \mathbf{u}_s \| (T_a - T_s),
\]

where ρ is the air density, \( L = 2.45 \times 10^6 \) J kg⁻¹ is the latent heat of evaporation, and \( C_p = 1005 \) J kg⁻¹ is the specific heat of air at constant pressure. The turbulent fluxes in (1) are parameterized using wind speed (\( w = \| \mathbf{u}_a - \mathbf{u}_s \| \) relative to the ocean surface current (relative wind speed is close to actual wind speed outside regions of strong currents), the difference of specific air humidity and specific humidity at the air–sea interface (\( q_a - q_s \)), and the difference of air temperature and SST (\( T_a - T_s \)).
The lower indices \(a\) and \(s\) indicate the atmosphere at the reference level (normally 10 m) and at the sea surface, respectively. The bulk transfer coefficients for wind stress \((C_D, \text{drag coefficient})\), latent heat flux \((C_E, \text{Dalton number})\), and sensible heat flux \((C_H, \text{Stanton number})\) are estimated from wind speed, air temperature, and SST using the Fairall et al. (2003) algorithm [the Coupled Ocean–Atmosphere Response Experiment, version 3 (COARE3)]. LHTFL is positive if the ocean loses heat, whereas \(Q_H\) is positive if the ocean gains heat.

The variables needed for the evaluation of (1) are obtained from satellite measurements. Wind speed relative to the ocean surface current \(\mathbf{u}_a - \mathbf{u}_s\) is measured by scatterometers onboard the European Research Satellites ERS-1 (1992–96) and ERS-2 (1996–2001) and by QuikSCAT (1999–2007) (e.g., Liu 2002). The humidity \(q_a\) is derived from the Special Sensor Microwave Imager multichannel brightness temperatures using the Bentamy et al. (2003) method; the specific surface humidity \(q_s\) is estimated from daily averaged SST. This version of LHTFL uses the new Reynolds et al. (2007) daily bulk SST whereas the previous version of LHTFL (Bentamy et al. 2003) is based on the Reynolds and Smith optimum interpolation (OI) SST analysis (OLv2) (Reynolds et al. 2007) weekly bulk SST. In the present version no correction is made for cool skin and diurnal warming. The air temperature is determined from remotely sensed data \((q_a\) and SST) based on the Bowen ratio method suggested by Konda et al. (1996).

The turbulent fluxes are calculated using the COARE3 algorithm from daily averaged values binned onto a 1° global grid over satellite swaths. Because of differences in sampling by different satellite radars and radiometers, the final flux estimate is further averaged weekly and spatially interpolated on a regular 1° grid between 80°S and 80°N using the kriging method as described by Bentamy et al. (1996). The accuracy of the resulting weekly fluxes is assessed by comparisons with in situ measurements from moored buoys in the tropical Atlantic and Pacific [the Prediction and Research Moored Array in the Tropical Atlantic (PIRATA) and the Tropical Atmosphere Ocean–Triangle Trans-Ocean Buoy Network (TAO–TRITON)], the northeastern Atlantic and northwestern Mediterranean (Met Office and Météo-France), and the National Data Buoy Center (NDBC) network off the U.S. coast in the Atlantic and Pacific Oceans.\(^1\) Quite high correlations (ranging from 0.8 to 0.92) are found between satellite and in situ LHTFL, while biases and standard deviations are generally low. Standard deviations of satellite and in situ LHTFL vary between 18 and 25 W m\(^{-2}\). The highest bias is found in comparisons with the NDBC buoys in the Gulf Stream region where the time-mean satellite LHTFL is 10 W m\(^{-2}\) below in situ values (or 7% of the NDBC regional LHTFL mean). In the tropics, satellite LHTFL overestimates in situ LHTFL by 8 W m\(^{-2}\). These comparisons indicate significant improvements of the new LHTFL product over the previous release described in Bentamy et al. (2003).

The same buoy network is used to evaluate the accuracy of \(T_a\) retrieval. The time-mean satellite-derived \(T_a\) is slightly colder than in situ air temperature. The bias is weaker at midlatitudes but magnifies up to \(-0.7^\circ\text{C}\) in the tropics (based on comparisons with the TAO–TRITON and PIRATA buoy data). The standard deviation of daily average satellite and in situ \(T_a\) is around \(0.6^\circ\text{C}\) and but is stronger \((0.9^\circ\text{C})\) in high-gradient SST areas such as the Gulf Stream. As expected, weekly averaging slightly decreases (by around 0.1°C) the standard deviation but does not affect the bias much.

The intraseasonal signal is evaluated in few steps. First, the annual cycle is calculated from the weekly data as a sum of the first three harmonics (Mestas-Nuñez et al. 2006). Next, the anomaly is calculated by subtracting the annual cycle from the original signal. Finally, the intraseasonal signal is calculated as the difference between the anomaly and its 13-week running mean. This procedure retains periods shorter than 3 months that are referred to as intraseasonal in this study. The variability of intraseasonal fluxes is characterized by the running standard deviation that mimics the upper envelope of the intraseasonal signal. The running standard deviation of the intraseasonal signal is calculated using the same 13-week running window. Comparisons of the satellite intraseasonal LHTFL with in situ data from the TAO–TRITON moorings in the tropical Pacific, the PIRATA moorings in the tropical Atlantic and the Research Moored Array for African–Asian–Australian Monsoon Analysis and Prediction (RAMA) moorings in the tropical Indian Ocean are presented in the appendix.

The LHTFL from this study is compared with LHTFL provided by the National Center for Climate Prediction–National Center for Atmospheric Research (NCUR–NCAR) reanalysis (Kalnay et al. 1996), the Woods Hole Oceanographic Institution (WHOI) objectively analyzed air–sea fluxes (OAFlux) of Yu et al. (2004) that combine satellite data with model simulations, and with in situ shipborne estimates collected by the International Comprehensive Ocean–Atmosphere Dataset (ICOADS) of Worley et al. (2005). Mean sea level pressure for this

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study is provided by the NCEP–NCAR reanalysis. In situ measurements from the TAO–TRITON moorings in the tropical Pacific Ocean (McPhaden et al. 1998), the PIRATA moorings in the tropical Atlantic (Bourlès et al. 2008), and the RAMA moorings in the tropical Indian Ocean (McPhaden et al. 2009) are also used for comparisons.

For several years now, uniform, long-term data from observations made from numerous satellites relevant for inferring surface shortwave radiation have been prepared as homogeneous time series. The satellites that are being used for SWR retrieval usually have between two to five channels in spectral intervals that are relevant both for inferring SWR (visible) and for detecting clouds. Cloud data are provided by the International Satellite Cloud Climatology Project (ISCCP; version D1) at a nominal resolution of 2.5° at 3-h time intervals (Rossow and Schiffer 1999). The original version of the SWR retrieval scheme is described in Pinker and Laszlo (1992) and has been used at the National Aeronautics and Space Administration (NASA) Langley Research Center for generating the Global Energy and Water Cycle Experiment (GEWEX) Surface Radiation Budget (SRB) product (available online at http://gewex-srb.larc.nasa.gov). Since then, several modifications have been introduced to the inference scheme with regard to aerosols (e.g., Liu and Pinker 2008), data merging (Zhang et al. 2007), and elevation correction (the SWR data used in this study are derived with version 3.3.1).

3. Results

a. Mean LHTFL and seasonal variations

First, we consider the global patterns of the LHTFL and its annual and semiannual harmonics. These components form the annual cycle that is used as a reference for evaluating anomalies and the intraseasonal signal. Spatial patterns of magnitude and phase of these harmonics are similar to those in the Mestas-Nuñez et al. (2006) analysis, which is based on the 3-yr record (1996–98) from the previous release of the LHTFL archive of Bentamy et al. (2003). Comparison of the time-mean LHTFL from this study with the time-mean LHTFL provided by alternative analyses (NCEP–NCAR Reanalysis, WHOI OAFlux, and ICOADS) indicates reasonable correspondence of spatial patterns (Figs. 1a–d).

Variations of LHTFL closely follow variations of the product of sea–air humidity difference and wind speed $\Delta qw$ (1). This product accounts for a major portion of LHTFL variability because the Dalton number, $C_D$, has weak dependence on wind speed (for winds ranging from 4 to 14 m s$^{-1}$; Large and Pond 1982). For the four products shown in Fig. 1, the time-mean LHTFL is dominated by evaporation in the trade wind regions and resembles the time-averaged wind speed in the 30°S–30°N belt (Fig. 2a). In the tropical belt the humidity difference, $\Delta q = q_a - q_e$, is high (Fig. 2b) and has relatively weak meridional variations (except in the eastern Pacific and Atlantic). Hence, LHTFL variability is explained by winds.

SST impacts are evident in the equatorial eastern Pacific and Atlantic where the mean LHTFL weakens because of the presence of cold tongues of SST maintained by the equatorial upwelling. A local minimum of evaporation over the cold tongue regions is explained by the direct impact of cool SST on the sea surface and air humidities (Fig. 2b), as well as by indirect impact of SST on the near-surface atmospheric boundary layer, which tends to decelerate over cold water (Wallace et al. 1989). A stronger impact of SST on LHTFL is evident across the subtropical fronts where temperature sharply decreases with latitude. Poleward decrease in SST is accompanied by decrease in $T_d$ because the atmosphere boundary layer stability is close to neutral over much of the interior ocean. Hence, the humidity difference ($\Delta q$) also decreases sharply poleward of 30°S and 30°N (Fig. 2b) because both $q_a$ and $q_e$ decrease with temperature in accordance with the Clausius–Clapeyron law. These meridional changes of $\Delta q$ explain weak LHTFL in the extratropical oceans (Figs. 1a–d) in spite of rather strong winds in the Northern Hemisphere and especially the Southern Hemisphere storm track corridors (Fig. 2a).

Regardless of the good correspondence of the geographical distribution of the time-mean LHTFL, the four analyses are somewhat different in magnitude. In the current analysis, LHTFL (Fig. 1a) has higher values in the trade wind regions than in the other three analyses. This analysis is closer to in situ ship observations from ICOADS (Fig. 1d) and the NCEP–NCAR reanalysis (Fig. 1b) but exceeds the WHOI OAFlux estimates by 20 to 40 W m$^{-2}$ (Fig. 1c). As a result, this study suggests the highest estimate of the globally averaged LHTFL and evaporation summarized in Table 1.

Strong time-mean latent heat loss (exceeding 80 W m$^{-2}$) is drawn from the warm Gulf Stream waters off the east coast of the United States. Similarly, strong time-mean LHTFL is observed near Japan over the warm Kuroshio (Figs. 1a–d). In both of these regions the LHTFL experiences the strongest annual variation peaking during the winter, when cold dry continental air offshore of North America and Japan crosses the Gulf Stream north wall in the Atlantic or the Kuroshio SST front in the Pacific, respectively (Figs. 1e,f). Semiannual LHTFL
variations are prominent in the Arabian Sea and Bay of Bengal because of the annual reversal of winds forced by the South Asian monsoon (Figs. 1g,h). The monsoon flow in the Arabian Sea low-level westerly jet intensifies in boreal summer whereas northeasterly winds spread over the region in boreal winter when the monsoon ceases. Weaker semiannual variability is observed in the Caribbean low-level jet where the easterly winds also intensify twice a year in February and again in July (Muñoz et al. 2008).

Fig. 1. 1992–2007 mean LHTFL (W m$^{-2}$) from (a) this study, (b) NCEP–NCAR reanalysis, (c) WHOI OAFlux, and (d) ICOADS. The means are based on whatever part of this time interval is available. Annual harmonics: (e) magnitude and (f) phase. Semiannual harmonics: (g) magnitude and (h) phase. Phase is in months.
FIG. 2. Time-mean (a) wind speed $w$ and (b) sea–air specific humidity difference, $\Delta q = q_s - q_a$. Standard deviation of intraseasonal (c) wind speed $w$ and (d) humidity difference $\Delta q$. (e) Standard deviation of intraseasonal LHTFL and contribution to it from intraseasonal variation of (f) wind speed, (g) specific humidity, and (h) saturated near-surface humidity. The prime symbol denotes the intraseasonal component.
b. Magnitude of intraseasonal variation

As expected from (1), the intraseasonal variability of LHTFL is defined by the intraseasonal variability of winds (Fig. 2c) and the sea–air humidity difference (Fig. 2d). However, the spatial distribution of the intraseasonal LHTFL (Fig. 2e) bears only partial correspondence to the spatial distribution of intraseasonal winds or the sea–air humidity difference. In particular, the decrease in variance of intraseasonal LHTFL toward the equator reflects relatively weak variability of intraseasonal wind at low latitudes. In contrast to low latitudes, the intraseasonal variability of LHTFL decreases at high latitudes despite stronger wind variability there. This behavior is explained by the spatial distribution of the time-mean humidity difference that is weak over cold SSTs (and cold $T_a$) of each hemisphere (Fig. 2b).

Although linear decomposition of the intraseasonal LHTFL suggests that the wind component $[-\Delta q \cdot \text{STD}(w^*)$; Fig. 2f] accounts for a major portion of variability of the intraseasonal LHTFL, neither component dominates globally. In particular, the air humidity variability component $[-w \cdot \text{STD}(q_s^*)$; Fig. 2g] peaks along the major SST fronts and reflects an impact of moisture transport across the ocean SST fronts by synoptic weather systems. SST itself $[-w \cdot \text{STD}(q_s^*)$; Fig. 2h] also impacts the intraseasonal LHTFL along the major western boundary current fronts and in the Agulhas current area. Both the mean LHTFL (Fig. 1a) and its variability (Fig. 2e) weaken over cold SSTs where mean values of LHTFL are also low. This is particularly evident in the cold sector of the Gulf Stream, in the Brazil–Malvina confluence region, in the subpolar North Pacific, and in the Southern Ocean.

c. Intraseasonal LHTFL in midlatitudes

The strongest variability of the intraseasonal LHTFL occurs in the midlatitudes where the regional maxima are linked to areas of major SST fronts (Fig. 2e). In particular, in the Atlantic sector the highest intraseasonal variance is observed along the Gulf Stream front. Similarly high intraseasonal variability is observed in the Agulhas Current and in the Brazil–Malvina confluence region. This suggests that the stratified atmospheric boundary layer plays an important role in amplifying intraseasonal air–sea interactions. The intraseasonal LHTFL variance changes seasonally and peaks in winter (not shown), suggesting an association with midlatitude storms, which also intensify in the cold season.

We next identify weather systems that are responsible for strong intraseasonal variability of LHTFL in these regional maxima areas by projecting the intraseasonal LHTFL time series spatially averaged over a particular index area onto atmospheric parameters elsewhere. This regression analysis reveals correspondence between the strengthening of intraseasonal LHTFL in the Gulf Stream region and midlatitude storm systems in the Atlantic (Fig. 3a). Increase in LHTFL drawn from the Gulf Stream region is associated with the area of mean sea level pressure low and corresponding cyclonic anomalous winds centered east of the region. The air pressure pattern is similar to that deduced by Zolina and Gulev (2003) and by Foltz and McPhaden (2004) in their analyses of the intraseasonal variability of the Atlantic winds. In fact, the anomalous wind in Fig. 3b decelerates the northern flank of the northeasterly trades (where anomalous LHTFL is somewhat weaker) and significantly accelerates off-shore winds over the Gulf Stream (Fig. 3b). A maximum increase in wind speed is observed over the warm sector of the Gulf Stream where winds further accelerate because of the atmospheric boundary layer adjustment (Beal et al. 1997). In addition to intensification of mean winds, the anomalous northwesterly wind outbreaks bring cold and dry continental air toward the sea. Spreading of dry continental air lowers air humidity, thus increasing the air–sea humidity contrast (Fig. 3c). This, in turn, compliments the LHTFL increase due to stronger winds. The ocean responds to continental air outbreaks by cooling SST north of the Gulf Stream northern wall, which is seen in decreasing values of $q_s$ (Fig. 3c). Intraseasonal winds have a weak impact on SST south of the Gulf Stream temperature front where the ocean mixed layer is deep and its thermal inertia is relatively strong.

A similar correspondence between intraseasonal LHTFL and atmospheric synoptic systems has been observed by Bond and Cronin (2008) in the Kuroshio extension region in late fall through early spring, when cold air outbreaks associated with synoptic events lead to intense regional episodes of LHTFL and sensible heat loss. Here we focus on regions of locally strong intraseasonal LHTFL observed in the South Atlantic in the Agulhas Current south of the Cape of Good Hope (Fig. 2e). Similarly to what occurs in the Gulf Stream area, an increase of LHTFL over the warm Agulhas
Current is linked to passing storms (Fig. 4). When the storm center locates to the east of the index area, the anomalous southerly winds bring cold and dry sub-Antarctic air northward. This amplifies the latent heat loss due to increasing wind speed and the increasing air–sea surface humidity difference. Although storm systems are generally strong as they propagate around the globe in the South Atlantic and the Southern Oceans, the intraseasonal LHTFL is stronger in the Agulhas region and in the Brazil–Malvina confluence (Fig. 2e) in comparison to values observed at similar latitude in the ocean interior. Both these areas host sharp SST fronts that promote higher $\Delta q$ and stronger LHTFL. It is interesting to note that the regression analysis in Fig. 4 reveals a sequence of zonally propagating storm systems over the open spaces of the South Atlantic and South Oceans. The mean sea level pressure troughs in the regression pattern are separated by approximately 90° in longitude, suggesting a zonal wavenumber of 4.

**d. Intraseasonal surface fluxes and SST**

The variability of LHTFL and SST is related. Exploring the relationships between the two offers an additional tool to evaluate the consistency of the flux product. LHTFL affects SST by affecting the net ocean surface heat balance; significantly, SST also affects LHTFL directly through $q_s$ and indirectly by affecting near-surface winds that accelerate over warmer SSTs. We next characterize the interplay between intraseasonal variations of LHTFL and SST.

As expected, the LHTFL response to underlying anomalous SST is generally positive (Fig. 5a); that is, LHTFL increases in response to increased SST. This suggests a damping of the underlying SST anomalies, although there are considerable geographical variations (Park et al. 2005). The feedback exceeds 20 W m$^{-2}$ °C$^{-1}$ in the regions around 20°S and 20°N, but it decreases at high latitudes and in the eastern tropical Pacific and Atlantic where the time-averaged LHTFL is also weak.

In contrast to the LHTFL response to underlying SST that is positive, the SST response to intraseasonal variation of LHTFL is negative over much of the ocean (Fig. 5b), suggesting a cooling down of SST in response to increasing surface heat lost. However, in several regions SST warms up in response to LHTFL increase. In particular, this behavior occurs in the cold tongue regions of the eastern tropical Pacific and Atlantic Oceans. The relationship between intraseasonal LHTFL and SST depends on the relative role the LHTFL plays in the mixed layer heat balance. If this balance is local and governed by the surface flux, the SST cools down in response to increasing latent heat loss (negative correlation when LHTFL leads). This negative relationship dominates away from the cold tongue regions and strong currents. In contrast, in the cold tongue regions the mixed layer temperature balance is governed primarily by the vertical (upwelling) or horizontal (tropical instability waves; e.g., Grodsky et al. 2005) heat transports.
Here the positive correlation between LHTFL and SST is explained by the stratified atmospheric boundary layer adjustment and associated wind acceleration over warm SSTs. Therefore, in the cold tongue regions the LHTFL increases in response to increasing winds and SST rather than SST responding to change in LHTFL. Over the regions where the surface heat flux dominates the mixed layer heat budget, the variations of LHTFL force variations of the mixed layer temperature and thus should be negatively correlated with the SST rate of change, $\frac{\partial T}{\partial t}$, as seen in Fig. 6a. The time correlation of intraseasonal LHTFL and $\frac{\partial T}{\partial t}$ is statistically significant over much of the ocean. It decreases at high latitudes where the upper ocean stratification is weak, the mixed layer is deep, and SST response is weak. The time correlation is also weak in the tropical Pacific and Atlantic Oceans in the regions where vertical and horizontal ocean heat transports (rather than surface flux) dominate the mixed layer heat budget. Similar but weaker correlation is found for the shortwave radiation (Fig. 6b).

In fact, the SST rate of change is driven by the net heat flux across the air–sea interface, for which the LHTFL and SWR are the major components. If these two components of the surface flux are combined to better represent the ocean heat lost (LHTFL-SWR), the correlation increases (Fig. 6c), suggesting reasonable correspondence of intraseasonal flux variations with intraseasonal SST.

e. Intraseasonal LHTFL and SWR

Both the intraseasonal LHTFL and SWR agree reasonably well with independent measurements of the rate of change of intraseasonal SST. We next explore the global correspondence between intraseasonal variations of the two surface flux components. They are weakly correlated over much of the global ocean, with the exception of the tropical Indian Ocean and the western tropical Pacific, where the intraseasonal LHFL and SWR are negatively correlated (Fig. 7a). Coherent variations of clouds and winds are evident in the Indo-Pacific warm pool where the eastward-propagating Madden–Julian oscillations are the most pronounced (Shinoda and Hendon 1998). Lagged correlation indicates that LHTFL increases in phase with decrease in SWR, suggesting stronger latent heat loss and evaporation just below the convective systems. Negative zero-lag correlation of SWR and LHTFL is observed with both satellite flux data and in situ TAO–TRITON mooring data (Fig. 7a, inlay). It is consistent with the analysis of the TAO/TRITON surface fluxes by Zhang and McPhaden (2000), who also have found nearly in-phase relationships among latent heat flux maxima and solar radiation minima during passage of the MJO events. This phase relationship remains in place if higher temporal resolution (i.e., daily) LHTFL (WHOI OAFlux by Yu et al. 2004) and daily SWR are used instead of weekly averaged data (Fig. 7b). It may be noted that lagged correlations in Figs. 7a and 7b have positive peaks at approximately 2 to 3 week lags (also seen in Fig. 9) that reflect an oscillatory behavior of lagged correlation for periodic processes like the MJO.

On the one hand, the out-of-phase variations of intraseasonal LHTFL and SWR support a hypothesis that the evaporation affects the humidity and therefore the cloudiness and thus solar radiation at the sea surface. But theoretical considerations [see Zhang and McPhaden (2000) for a summary of existing approaches] suggest a lagged relationship between intraseasonal LHTFL and SWR forced by MJO. In particular, in the Neelin et al. (1987) model the maximum LHTFL is shifted to the east of the convective center if the mean wind is easterly. The explanation of the phase relationship between LHTFL

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2 LHTFL is positive if the ocean loses heat.
3 The 99% confidence level of zero correlation is 0.1.
4 SWR is inversed to be consistent with the sign of LHTFL.
and SWR variations on intraseasonal time scales is not clear.

Next we present observed relationships between intraseasonal variations of SWR, LHTFL, and parameters affecting LHTFL. Coherent variations of the intraseasonal LHTFL lows and SWR highs (and vice versa) are apparent in the time–longitude diagrams in Fig. 8. These accorded intraseasonal variations propagate eastward between 60°E and the date line at speeds ranging from 4.5 to 7.5 m s⁻¹, typical of the MJO. East of the date line the correlation between LHTFL and SWR is weak (Fig. 8). This zonal change of correlation is explained by the lack of cloudiness east of the date line, which is the only major source of SWR variability.

Fig. 5. Lagged regression of intraseasonal LHTFL and SST: (a) SST leads LHTFL by 1 week; (b) LHTFL leads SST by 1 week. Areas where time correlation exceeds the 99% confidence level of zero correlation are dotted. Inlay in (b) shows lagged correlations of LHTFL and SST time series spatially averaged over the equatorial east Pacific (black) and the midlatitude Pacific (red). Negative lags are LHTFL lead time in weeks.
Intraseasonal LHTFL variations are mostly driven by the intraseasonal variations of wind speed (Araligidad and Maloney 2008). Wind strengthens just below convective clusters where SWR is low (Fig. 9a). Coherent variations of intraseasonal winds and SWR occur mostly west of the date line, with a gap in correlation over the maritime subcontinent. As expected, LHTFL strengthens in phase with winds, leading to an out-of-phase relationship.
between intraseasonal LHTFL and SWR (Fig. 9b). Evaluation of $q_a$ indicates that intraseasonal wind fluctuations are not the only forcing of intraseasonal LHTFL. In fact, $q_a$ also varies in accord with intraseasonal SWR (Fig. 9c). It may be suggested that specific air humidity decreases below convective systems in response to cooling of the near-surface atmosphere while the sea surface saturated humidity does not change much because of the thermal inertia of the ocean mixed layer. The difference in responses of $q_a$ and $q_s$ leads to an increase in the vertical gradient of air humidity below convective cloud clusters, which in turn further enhances anomalous evaporation and LHTFL produced by wind speed anomaly.

f. Intraseasonal and longer period variability of LHTFL

The interannual evolution of the ocean surface fluxes has been extensively studied. However, it appears that amplitudes of intraseasonal fluxes are not stationary and that they experience significant modulations by longer period variability. Noting that our dataset is only 16 yr long, our study is limited to the tropical Pacific Ocean,
which hosts the ENSO and thus displays significant interannual variability that can be resolved by relatively short records. Interannual SWR anomaly is modulated by ENSO through zonal displacements of convection. These interannual displacements of convection between the western tropical Pacific and the central tropical Pacific produce SWR anomalies that are well detected by satellite techniques (Rodriguez-Puebla et al. 2008). Because clouds are the only physical mechanism driving the intraseasonal SWR, the amplitude of intraseasonal SWR also shifts zonally following anomalously low SWR. In the central equatorial Pacific, the magnitude of intraseasonal SWR increases in phase with the warming of the Niño-3 SST (Fig. 10a, inlay). Here, the standard deviation of intraseasonal SWR increases by up to 5 W m\(^{-2}\) in response to a 1°C rise of SST in the Niño-3 region (Fig. 10a). As such, interannual variation of the amplitude of intraseasonal SWR reaches 15 W m\(^{-2}\).
during a mature phase of El Niño when anomalous Niño-3 SST warms up by 3°C. This interannual modulation of amplitude of the intraseasonal SWR is comparable to the characteristic amplitude of SWR variation by the MJO (Shinoda and Hendon 1998).

In contrast to the amplitude of intraseasonal SWR that varies in phase with El Niño, the magnitude of intraseasonal LHTFL does not have a similar significant in-phase variation. The impact of El Niño on the intraseasonal LHTFL differs from its impact on the total anomalous LHTFL, which is enhanced in the eastern tropical Pacific, around the Maritime Continent, and the in the equatorial Indian Ocean (Mestas-Nuñez et al. 2006). In contrast, the magnitude of intraseasonal LHTFL amplifies over the western tropical Pacific approximately 8 months in advance of the mature phase of El Niño (Fig. 10b and inlay). This amplification reflects impacts of the westerly wind bursts that often precede the onset of El Niño, which were evident in advance of the 2002/03 El Niño and particularly noticeable in advance of the 1997/98 event (McPhaden 2004).

4. Conclusions

Although the major portion of the intraseasonal variability of LHTFL is accounted for by winds, no one
component (wind, air humidity, or sea surface humidity) dominates the variability globally. In particular, contributions of $q_a$ and $q_s$ are significant along major SST fronts because of moisture transport across the ocean SST fronts by synoptic weather systems. Both the mean LHTFL and its intraseasonal variability weaken over cold SSTs because of the low air–sea humidity difference. In contrast, the strongest intraseasonal LHTFL is observed over the warm sectors of SST fronts.

The strongest variability of the intraseasonal LHTFL (in excess of 50 W m$^{-2}$) occurs at midlatitudes where the regional maxima are linked to areas of major SST fronts. In particular, in the Atlantic sector the highest intraseasonal variance is observed along the Gulf Stream. Similarly high variability is observed in the Agulhas Current and in the Brazil–Malvinas confluence. Coincidence of the regional maxima of intraseasonal LHTFL with SST fronts suggests the important role the stratified atmospheric boundary layer plays in amplifying intraseasonal air–sea interactions. Temporal variations of the intraseasonal LHTFL in these regional maxima are linked to passing midlatitude storms. The intraseasonal variability of LHTFL forced by these passing storms is locally amplified by unstable atmospheric stratification over warm SSTs.

Although weaker in amplitude (but still significant), intraseasonal variability of LHTFL (standard deviation of 20 to 30 W m$^{-2}$) is observed in the tropical Indian and Pacific Oceans. This variability is linked to the eastward-propagating Madden–Julian oscillations. In this tropical region the intraseasonal LHTFL and incoming solar radiation vary out of phase (i.e., evaporation enhances just below the convective clusters). The out-of-phase relationship between the intraseasonal LHTFL and SWR is observed west of the date line, whereas east of the date line both intraseasonal LHTFL and SWR are weak and their relationship is not significant. Both intraseasonal variations of wind speed and $q_a$ contribute to this out-of-phase relationship. Intraseasonal wind speed amplifies below convective clusters where SWR is low. Specific air humidity decreases below convective clusters following cooling of the near-surface atmosphere whereas $q_s$ does not change much because of the ocean thermal inertia. The difference in responses of $q_a$ and $q_s$ increases the vertical gradient of air humidity below convective cloud clusters and thus enhances anomalous evaporation and LHTFL produced by anomalous wind speed.

Amplitudes of intraseasonal LHTFL and SWR display significant interannual variations in the tropical Pacific Ocean. Amplitudes of intraseasonal SWR increase in the central equatorial Pacific by 15 W m$^{-2}$ during the mature phase of El Niño following the eastward shift of convection. In contrast to the amplitude of intraseasonal SWR, which varies in phase with El Niño, the amplitude of intraseasonal LHTFL does not exhibit similar significant in-phase variation. In contrast, the intraseasonal LHTFL amplifies over the western tropical Pacific approximately 8 months in advance of the mature phase of El Niño. This amplification reflects impacts of the westerly wind bursts that often precede the onset of El Niño.

Over much of the interior ocean where the ocean–atmosphere heat exchange drives the ocean mixed layer balance, SST cools down in response to anomalously strong LHTFL. There are considerable geographical variations in magnitude of this response that are related in part to the spatial variations of oceanic mixed layer depth and its thermal inertia, which mitigate the impact of surface fluxes. Moreover, in the eastern tropical Pacific and Atlantic cold tongues the SST warms up in response to LHTFL strengthening. In these equatorial upwelling areas the SST is strongly affected by the ocean advection and LHTFL responds to this rather than driving SST.

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APPENDIX

Validation of Satellite-Derived LHTFL in the Intraseasonal Band

Comparisons of in situ LHTFL with satellite-derived LHTFL in the intraseasonal band are shown in Fig. A1. This comparison is based on in situ buoy measurements in the tropics including 68 TAO–TRITON buoys in the Pacific, 21 PIRATA buoys in the Atlantic, and 10 RAMA buoys in the Indian Ocean. During 1992–2007 the dataset has 30 592 concurrent buoy–satellite weekly measurements in the Pacific, 3044 weeks of data in the Atlantic, and 318 weeks of concurrent buoy and satellite data in the Indian Ocean. The aggregate time series of buoy and satellite LHTFL have statistically significant correlation around 0.6. The 99% confidence level of zero correlation is corr$_{99\%} < 0.1$ for the Pacific and Atlantic whereas it is slightly higher (corr$_{99\%} = 0.14$) for the Indian Ocean because of the shorter time series. Time series of intraseasonal LHTFL at each buoy
location also indicate significant correlation (Figs. A1a,c). Time correlation (TCORR) exceeds 0.6 over much of the tropical Pacific where the average length of the LHTFL time series at particular buoy is around 450 weeks (corr$_{99\%}$ = 0.12). TCORR increases toward the west following the westward increase of the mean LHTFL in the tropical Pacific (Fig. 1a). In contrast, somewhat weaker TCORR is observed along 5\degree N where LHTFL is weaker due to weaker winds and higher specific humidity in the ITCZ. The impact of the ITCZ is better seen in the Atlantic where TCORR decreases below 0.5 in the ITCZ area (Fig. A1c). These comparisons

Fig. A1. Comparison of intraseasonal buoy and satellite-derived LHTFL. Spatial maps of (left) time correlation and (right) scatter diagrams for the (a),(b) tropical Pacific, (c),(d) tropical Atlantic, and (e),(f) tropical Indian Ocean. Gray shading in (b) and (d) shows the standard deviation of satellite intraseasonal LHTFL in 2 W m$^{-2}$ intervals of in situ data. TCORR is the time correlation evaluated from the aggregate satellite/buoy comparisons for each basin; NUM is the length of the aggregate record (in weeks). Buoy locations are shown by closed circles in (a),(c), and (e).
suggest that the LHTFL retrieval should be rectified in the ITCZ area. The air relative humidity has a regional maximum in the ITCZ area. Therefore, the meridional displacement of the ITCZ could produce variations of the relative humidity strong enough to require being taken into account in the Konda et al. (1996) Bowen ratio approach.

Satellite intraseasonal LHTFL compares well with in situ LHTFL to within the scatter of the data (Figs. A1b,d,f). However, the magnitude of satellite intraseasonal LHTFL is weaker than in situ data. This bias is more evident in the Pacific where the intraseasonal variations of satellite LHTFL are 15%–20% weaker than those from buoys, whereas this bias is less than 10% in the Atlantic and is not evident in the Indian Ocean. We attribute this bias to the spatial and temporal smoothing of satellite data that inevitably results in losing of a portion of variance observed at fixed location and high temporal resolution.

REFERENCES


