Intraseasonal latent heat flux based on satellite observations

Semyon A. Grodsky\textsuperscript{1}, Abderrahim Bentamy\textsuperscript{2}, James A. Carton\textsuperscript{1}, and Rachel T. Pinker\textsuperscript{1}

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\textsuperscript{1}Department of Atmospheric and Oceanic Science, University of Maryland, College Park, MD 20742
\textsuperscript{2}Institut Francais pour la Recherche et l’Exploitation de la Mer, Plouzane, France

Corresponding author:
\texttt{senya@atmos.umd.edu}
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 mid-latitudes. It weakens towards 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 $z=10\text{m}$ and $z=0\text{m}$. Within the midlatitude belts the intraseasonal variability of LHTFL is locally stronger (up to 50 Wm$^{-2}$) 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 due in part to 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 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-year (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 timescales (e.g., da Silva et al., 1994; Yu et al., 2006). Recent studies of mid-latitudes (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, 1994). Since LHTFL is also proportional to evaporation its intraseasonal variations also 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 MJO lead to significant intraseasonal LHTFL. MJO may be driven in part by the evaporation-wind feedback (Neelin et al., 1987). 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. MJO is characterized by strong 2-4 m/s fluctuations of surface winds and precipitation (Araligidad and Maloney, 2008). As a result, it produces correlated fluctuations of both LHTFL and shortwave radiation (SWR) with amplitudes of 30-50 Wm\(^{-2}\) 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 the 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 subtropics 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 subtropical Atlantic and found their link to fluctuations in the strength of the Azores high. The subtropics and mid-latitudes are subject to synoptic meteorological forcing originating in the mid-latitude 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 and other observational evidence suggest 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, as well as modulation by longer period processes.

This study is possible due to 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 provide an estimate of relative humidity. When combined with estimates of surface temperature it is 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 (Rossow 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 ($\tau$), LHTFL ($Q_E$), and sensible heat flux ($Q_H$) are estimated using the following bulk aerodynamic parameterizations (Liu et al., 1979):

$$\frac{\tau}{\rho} = C_D |u_a - u_s| (u_a - u_s)$$

$$\frac{Q_E}{\rho L} = -C_E |u_a - u_s| (q_a - q_s)$$

$$\frac{Q_H}{\rho C_p} = C_H |u_a - u_s| (T_a - T_s)$$

where $\rho$ is the air density, $L = 2.45 \times 10^6$ J/kg is the latent heat of evaporation, $C_p = 1005$ J/kg is the specific heat of air at constant pressure. The turbulent fluxes in (1) are parameterized using wind speed ($w = |u_a - u_s|$) 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 atmosphere at the reference level (normally 10m) and at the sea surface,
respectively. The bulk transfer coefficients for wind stress \( C_D \), drag coefficient, latent heat flux \( C_E \), Dalton number, and sensible heat flux \( C_H \), Stanton number) are estimated from wind speed, air temperature, and SST using the *Fairall et al. (2003)* algorithm (COARE3 version). LHTFL is positive if the ocean loses heat, while \( Q_H \) 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 \( \left| \mathbf{u}_a - \mathbf{u}_s \right| \) is measured by scatterometers onboard the European Research Satellites ERS-1 (1992-1996), ERS-2 (1996-2001), and QuikSCAT (1999-2007) (e.g. Liu, 2002). The humidity \( q_a \) is derived from the Special Sensor Microwave Imager multi channel brightness temperatures using the *Bentamy et al. (2003)* method, while 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 while the previous version of LHTFL (*Bentamy et al., 2003*) is based on the *Reynolds and Smith OIv2* 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 COARE3.0 algorithm from daily averaged values binned onto a 1° global grid over satellite swaths. Due to 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 (PIRATA and TAO/TRITON), the northeastern Atlantic and northwestern Mediterranean (UK Met Office and Météo-France), and the National Data Buoy Center (NDBC) network off the U.S. coast in the Atlantic and Pacific Oceans. 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 from 18 Wm$^{-2}$ and 25 Wm$^{-2}$. The highest bias is found in comparisons with the NDBC buoys in the Gulf Stream region where the time mean satellite LHTFL is 10 Wm$^{-2}$ below in-situ values (or 7% of the NDBC regional LHTFL mean). In the tropics satellite LHTFL overestimates in-situ LHTFL by 8 Wm$^{-2}$. 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 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 °C in the tropics (based on comparisons with the TAO/TRITON and PIRATA buoy data). Standard deviation of daily average satellite and in-situ $T_a$ is around 0.6 °C and but is stronger (0.9 °C) in high gradient SST areas like the Gulf Stream. As expected, weekly averaging decreases slightly (by around 0.1°C) the standard deviation but doesn’t affect the bias much.

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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 \((\text{Mestas-Nuñez et al.}, 2006)\). 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. Running standard deviation of the intraseasonal signal is calculated using the same 13 week running window. Comparisons of the satellite intraseasonal LHTFL with \textit{in-situ} data from the TAO/TRITON moorings in the tropical Pacific, the PIRATA moorings in the tropical Atlantic, and the 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 (NCEP/NCAR) reanalysis \((\text{Kalnay et al.}, 1996)\), the Woods Hole Oceanographic Institution objectively analyzed air-sea fluxes (OAFlux) of \(\text{Yu et al.} (2004)\) that combines satellite data with model simulations, and with in-situ ship borne estimates collected by the International Comprehensive Ocean-Atmosphere Data Set (ICOADS) of \(\text{Worley et al.} (2005)\). Mean sea level pressure for this study is provided by the NCEP/NCAR reanalysis. In-situ measurements from the TAO/TRITON moorings in the tropical Pacific Ocean \((\text{McPhaden et al.}, 1998)\), the PIRATA moorings in the tropical Atlantic \((\text{Bourles et al.}, \ldots)\)
For several years now, uniform, long-term data from observations made from numerous satellites relevant for inferring surface shortwave radiation (SWR) 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 (version D1) at a nominal resolution of 2.5° at 3hr 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 NASA/Langley for generating the GEWEX/SRB product\(^1\). Since, several modifications have been introduced to the inference scheme as related to aerosols (e.g. Liu and Pinker, 2008), data merging (Zhang et al., 2007), and elevation correction (Ma and Pinker, 2009) (the SWR data used in this study are derived with Version 3.3.1).

3. Results

Mean LHTFL and seasonal variations

First, presented are 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 intraseasonal signal. Spatial patterns of magnitude and phase of these harmonics are similar to the Mestas-Nuñez et al. (2006) analysis that is based on the three-year record (1996-1998) from the previous release of the LHTFL archive of

\(^1\)http://gewex-srb.larc.nasa.gov
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. 1 a-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 as the Dalton number, $C_e$, has weak dependence on wind speed (for winds ranging from 4 m s$^{-1}$ 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 average wind speed in the 30°S to 30°N belt (Fig. 2a). In the tropical belt the humidity difference, $\Delta q = q_s - q_a$, 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 due to the presence of cold tongues of SST maintained by the equatorial upwelling. Local minimum of evaporation over the cold tongue regions is explained by 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 that 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 $T_a$ 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 $q_s$ 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 – 1d) in spite of rather strong winds in the northern and
especially 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 the ICOADS (Fig. 1d)
and the NCEP/NCAR reanalysis (Fig. 1b), but exceeds the WHOI OAFlux estimates by
20 to 40 Wm$^{-2}$ (Fig. 1c). As a result this study suggests the highest estimate of the
globally averaged LHTFL and evaporation summarized in Table 1.

Table 1. Globally and time averaged LHTFL and evaporation.

<table>
<thead>
<tr>
<th></th>
<th>Global mean</th>
<th>This Study</th>
<th>NCEP/NCAR</th>
<th>WHOI OAFlux</th>
<th>ICOADS 2.4</th>
</tr>
</thead>
<tbody>
<tr>
<td>LHTFL (Wm$^{-2}$)</td>
<td>107</td>
<td>97</td>
<td>92</td>
<td>105</td>
<td></td>
</tr>
<tr>
<td>Evaporation</td>
<td>11.2</td>
<td>10.2</td>
<td>9.6</td>
<td>11.0</td>
<td></td>
</tr>
<tr>
<td>(cm month$^{-1}$)</td>
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Strong time-mean latent heat loss (exceeding 80 Wm$^{-2}$) 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 (Fig. 1 a-d). In both of these regions the LHTFL experiences the strongest annual variation peaking during the winter, when cold dry continental air off-shore of North America and Japan crosses the Gulf Stream north wall in the Atlantic or the Kuroshio SST front in the Pacific, respectively (Fig. 1e, 1f).

Semiannual LHTFL variations are prominent in the Arabian Sea and Bay of Bengal due to annual reversal of winds forced by the South Asian Monsoon (Figs. 1g, 1h). The monsoon flow in the Arabian Sea low-level westerly jet intensifies in boreal summer while 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 (Munoz et al., 2008).

Magnitude of intraseasonal variation

As expected from (1), the intraseasonal variability of LHTFL is defined by intraseasonal variability of winds (Fig. 2c) and sea-air humidity difference (Fig. 2d). But the spatial distribution of the intraseasonal LHTFL (Fig. 2e) bears only partial correspondence to the spatial distribution of intraseasonal winds or sea-air humidity difference. In particular, the decrease in variance of intraseasonal LHTFL towards 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 wind component ($\sim q' \cdot \text{STD}(w')$, Fig. 2f) accounts for a major portion of variability of the intraseasonal LHTFL, neither component dominates globally. In particular, the air humidity variability component ($\sim w' \cdot \text{STD}(q_w')$, 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 ($\sim w' \cdot \text{STD}(q_s')$, 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.

**Intraseasonal LHTFL in mid-latitudes**

The strongest variability of the intraseasonal LHTFL occurs in mid-latitudes 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 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). Maximum increase in wind speed is observed over the warm sector of the Gulf
Stream where winds further accelerate due to 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 towards 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 that 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.

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, 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 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 promotes 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 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 the zonal wavenumber of 4.
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. In fact, LHTFL affects SST by affecting the net ocean surface heat balance. But, 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), i.e., 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 Wm$^{-2}$/°C in the regions around 20°S and 20°N, but decreases at high latitudes and in the eastern tropical Pacific and Atlantic where the time average 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 cooling down of SST in response to increasing surface heat lost. But, 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 responses 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, $\partial T / \partial t$, as seen in Fig. 6a. The time correlation of intraseasonal LHTFL and $\partial T / \partial t$ 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 short wave radiation, SWR, (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

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1 LHTFL is positive if the ocean loses heat.
2 The 99% confidence level of zero correlation is 0.1.
3 SWR is inversed to be consistent with the sign of LHTFL.
increases (Fig. 6c) suggesting reasonable correspondence of intraseasonal flux variations with intraseasonal SST.

**Intraseasonal LHTFL and SWR**

Both, the intraseasonal LHTFL and SWR agree reasonably 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 an exception of the tropical Indian Ocean and the western tropical Pacific where the intraseasonal LHTFL 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, 1988). Lagged correlation indicates that LHTFL increases in phase with decrease in SWR, suggesting stronger latent heat loss and evaporation just below 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 Zhang and McPhaden (2000) analysis of the TAO/TRITON surface fluxes who also have found near in-phase relationships among maxima in latent heat flux and minima in solar radiation 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 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. Explanation of the phase relationship between LHTFL and SWR variations on the intraseasonal timescales is not clear. Next presented are 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 dateline at speeds ranging from 4.5 ms\(^{-1}\) to 7.5 ms\(^{-1}\), typical of the MJO. East of the dateline 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 dateline that is the only major source of SWR variability.

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 dateline with a gap in correlation over the maritime subcontinent. As expected, LHTFL strengthens in phase with winds that leads to 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 doesn’t change much because of the thermal inertia of the ocean mixed layer. Difference in responses of $q_a$ and $q_s$ leads to an increase in the vertical gradient of air humidity below convective cloud clusters that, in turn, further enhances anomalous evaporation and LHTFL produced by wind speed anomaly.

Intraseasonal and longer period variability of LHTFL

The interannual evolution of the ocean surface fluxes has been extensively studied. But, it appears that amplitudes of intraseasonal fluxes are not stationary and experience significant modulations by longer period variability. Noting that our dataset is only 16 years long, our study is limited to the tropical Pacific Ocean that 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 warming of the Nino3 SST (Fig. 10a, inlay). Here, the standard deviation of intraseasonal SWR increases by up to 5 Wm$^{-2}$ in response to a 1°C rise of SST in the Nino3 region (Fig. 10a). As such, interannual variation of the amplitude of intraseasonal SWR reaches 15 Wm$^{-2}$ during a mature phase of El Niño when anomalous Nino3 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 doesn’t have a similar significant in-phase variation. Impact of El Niño on the intraseasonal LHTFL differs from its impact on the total anomalous LHTFL that is enhanced in the eastern tropical Pacific, around the Maritime Continent, and the equatorial Indian Ocean (Mestas-Nunez 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 due to moisture transport across the ocean SST fronts by synoptic weather systems. Both the mean LHTFL and its intraseasonal variability weaken over cold SSTs due to 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 Wm$^{-2}$) occurs at mid-latitudes 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-Malvina 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 Wm$^{-2}$) 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 dateline, while east of the dateline 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 while $q_s$ doesn’t change much because of the ocean thermal inertia. 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 Wm$^{-2}$ during mature phase of El Niño following the eastward shift of convection. In contrast to the amplitude of intraseasonal SWR that varies in phase with El Niño, the amplitude of intraseasonal LHTFL doesn’t 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-Nino.

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 that mitigates 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.

Appendix

Comparisons of in-situ LHTFL with satellite-derived LHTFL in the intraseasonal band is shown in Fig. 11. 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 data set has 30592 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 while it is slightly higher $corr_{99%} = 0.14$ for the Indian Ocean due to shorter time series. Time series of intraseasonal LHTFL at each buoy location also indicate significant correlation (Figs. 11a, 11c). Time correlation (TCORR) exceeds 0.6 over much of the tropical Pacific where average length of the LHTFL time series at particular buoy is around 450 weeks ($corr_{99%} = 0.12$). TCORR increases towards the west following the westward increase of the mean LHTFL in the tropical Pacific (Fig. 1a). In contrast, somewhat weaker TCORR is observed along $5^\circ$ 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. 11c). These comparisons suggest that the LHTFL retrieval should be rectified in the ITCZ area. The air relative humidity has regional maximum in the ITCZ area. Therefore, the meridional displacement of the ITCZ could produce variations of the relative humidity strong enough that need to be accounted for 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. 11b, 11d, 11f). 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% to 20% weaker than those from buoys, while 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.

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Table 1. Globally and time averaged LHTFL and evaporation.

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<th>Global mean</th>
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<th>WHOI OAFlux</th>
<th>ICOADS 2.4</th>
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<td>9.6</td>
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Figure legends.

Figure 1. 1992-2007 mean LHTFL (Wm$^{-2}$) 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 month.

Figure 2. Time mean (a) wind speed, $w$, (b) sea-air specific humidity difference, $\Delta q = q_s - q_a$. Standard deviation of intraseasonal (c) wind speed, $w$, (d) humidity difference, $\Delta q$. (e) Standard deviation of intraseasonal LHTFL and contribution to it from intraseasonal variation of (f) wind speed, (e) specific humidity, and (h) saturated near surface humidity. The prime symbol denotes the intraseasonal component.

Figure 3. Time regression of intraseasonal LHTFL index averaged over the Gulf Stream area onto (a) mean sea level pressure and winds, (b) LHTFL elsewhere and wind speed, (c) saturated near surface humidity and humidity. The index area is defined as the area where STD of intraseasonal LHTFL exceeds 40 Wm$^{-2}$ (see Fig. 2e) and is dotted in panel (a). The index is defined as the index area average intraseasonal LHTFL normalized by its standard deviation (25 Wm$^{-2}$). Only wind arrows exceeding 0.4 ms$^{-1}$ are shown in panel (a).

Figure 4. Time regression of intraseasonal LHTFL index averaged over the Agulhas Current area onto intraseasonal mean sea level pressure (contours) and winds (arrows).
The index is defined as an area average intraseasonal LHTFL normalized by its standard deviation (37 Wm\(^{-2}\)). The index area is shown by the shaded rectangle. Only wind arrows exceeding 0.4 ms\(^{-1}\) are shown.

Figure 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 panel (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.

Figure 6. Time correlation of the rate of change of intraseasonal SST (\(\partial T / \partial t\)) with (a) intraseasonal LHTFL, (b) inversed intraseasonal short wave radiation (-SWR), and (c) sum of the two (LHTFL-SWR). Correlation exceeding 0.1 is significant at the 99% confidence level.

Figure 7. Time correlation of (a) weekly average and (b) daily average intraseasonal LHTFL and SWR. Inlay in each panel shows lagged correlation of LHTFL and SWR (solid) averaged over the equatorial Indian Ocean area marked by the rectangle and (dashed) from the TAO/TRITON mooring at 0°N, 165°E (shown by the cross). Lags are in weeks. Positive lags imply that SWR leads LHTFL. Weekly average LHTFL is from this study, daily average LHTFL is the Woods Hole Institution OAFlux by Yu et al. (2004).
Figure 8. Time-longitude diagrams of intraseasonal (a) LHTFL and (b) SWR averaged 5°S to 5°N in the Indian and western Pacific Oceans. Slope lines mark the same events and correspond to eastward propagation at 4.5 to 7.5 ms⁻¹.

Figure 9. Lagged correlation along the equator of intraseasonal (a) wind speed (W) and SWR, (b) LHTFL and SWR, and (c) specific air humidity (Qa) and SWR. All variables are averaged over the 5°S to 5°N belt. Land points are vertical boxes shaded in gray.

Figure 10. Time regression of anomalous Nino3 SST (210°E-270°E, 5°S-5°N) with running standard deviation, σ, of intraseasonal (a) SWR and (b) LHTFL. Correlation for SWR is instantaneous, while LHTFL is correlated with Nino3 SST that lags it by 35 weeks. Inlays show lagged correlation of Nino3 SST with the intraseasonal variance of flux spatially averaged over the rectangle shown in each panel. Positive lags imply flux leading Nino3 SST. Areas where the time regression is significant at the 99% level are cross-hatched.

Figure 11. Comparison of intraseasonal buoy and satellite-derived LHTFL. Spatial maps of time correlation (left) and scatter diagrams (right) for (a,b) tropical Pacific, (c,d) tropical Atlantic, and (e,f) tropical Indian Ocean. Gray shading in (b,d) shows the standard deviation of satellite intraseasonal LHTFL in 2 Wm⁻² intervals of in-situ data. TCORR is the time correlation evaluated from the aggregate satellite/buoy comparisons.
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