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Oscillations of Bay of Bengal sea surface temperature during the 1998 summer monsoon

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Abstract. New measurements from moored buoys in the Bay of Bengal, along with satellite cloud data, reveal strong monsoon intraseasonal oscillations (ISO) during the summer of 1998. The active phase of the monsoon is marked by high surface wind and deep atmospheric convection. The buoy data show that sea surface temperature (SST) in the Bay of Bengal warm pool rises and falls with periods of weeks. These intraseasonal oscillations of SST are not adequately captured in a satellite derived weekly SST analysis. They are a direct response to ISO of net surface heat flux into the ocean, which is negative in the active phase of the monsoon and positive in the quiescent phase. Fresh water from rivers and rain appears to control northern Bay of Bengal SST in late summer by allowing sunlight to escape below a shallow mixed layer.

1. Introduction

The slow rhythm of dry and rainy spells over south Asia in summer is a manifestation of the intraseasonal oscillations of the monsoon. Low frequency (10-50 day) ISO of the South Asian summer monsoon are associated with episodes of large scale organised atmospheric convection, often marked by cloud bands moving northward over the warm northeast Indian ocean and over land [*Sikka and Gadgil*, 1980; *Webster et al.*, 1998]. ISO of surface wind and SST in the Bay of Bengal were discovered during the monsoon experiment of 1979. This led to the suggestion that resultant changes in air-sea fluxes influence intraseasonal variability of the monsoon atmosphere [*Krishnamurti et al.*, 1988]. The physical basis for the existence of intraseasonal changes in the upper ocean and their role in monsoon ISO remains a subject of great interest [*Webster et al.*, 1998].

In summer, the Bay of Bengal forms a part of the warm pool covering the tropical east Indian and west Pacific ocean. The warm pool has high climatological SST accompanied by persistent deep convection in the atmosphere. In the equatorial west Pacific, the relation between eastward moving ISO of atmospheric convection and ISO of SST are well documented [*Godfrey et al.*, 1998]. Here we study the influence of monsoon ISO on Bay of Bengal warm pool SST with the help of the first long term *in situ* observations of the near surface ocean and atmosphere combined with satellite estimates of cloud. The *in situ* data from moored surface buoys show energetic ISO of SST in the summer of 1998 that are not adequately captured by a satellite based SST analysis. We find that to the first approximation, these SST changes are a direct response to changes in surface heat flux arising from intraseasonal variations of monsoon clouds and winds, once the effects of surface salinity are taken into account.

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2. Intraseasonal oscillations of SST and surface heat fluxes

Three moored buoys were deployed at 13°N,87°E, 16°N, 82°E and 18°N,88°E in the Bay of Bengal (Figure 1, inset) in September 1997. We refer to these as the southern, central and northern buoy respectively. The buoys measure sea surface temperature, salinity, ocean currents, air temperature and winds every three hours [*Rao and Premkumar*, 1998; *Premkumar et al.*, 1999], giving the first long records with sufficient time resolution to study monsoon ISO in the Indian ocean warm pool. SST fluctuations with daily and intraseasonal periods are prominent at all three buoys (Figure 1). The longer lived ISO events with periods of weeks are coherent across the region, and are the subject of this study. A widely used global weekly analysis from the U.S. National Centers for Environmental Prediction (NCEP) [*Reynolds and Smith*, 1994] that blends satellite SST with ship and buoy reports captures the cooling early in the monsoon season at the northern and southern buoys, and the SST changes in late summer, but not the ISO in July and August. This is because temperature sensors on the satellites do not see through cloud, and in situ data is sparse (R.W. Reynolds, personal communication). Data from the moored buoys does not enter the SST analysis . One of the referees has pointed out that a possible source of cold bias in satellite SST during fair weather is regression to ship intake temperatures in the presence of surface intensified warming of the ocean.

Intraseasonal changes are seen in several near surface variables measured by the buoys, and in satellite cloud. Fluctuations in surface wind speed and satellite derived outgoing longwave radiation (OLR) [*Gruber and Krueger*, 1984] are modulated on intraseasonal time scales at all three buoys, as illustrated in Figure 2. Over the tropical ocean , OLR is a proxy for convection. High values of OLR indicate clear skies whereas low OLR comes from high clouds with cold tops. Broadly, the low frequency variability of the monsoon atmosphere is an alternation between two phases - a quiescent phase, when it is clear and calm, and a convectively active phase, when it is cloudy and windy.

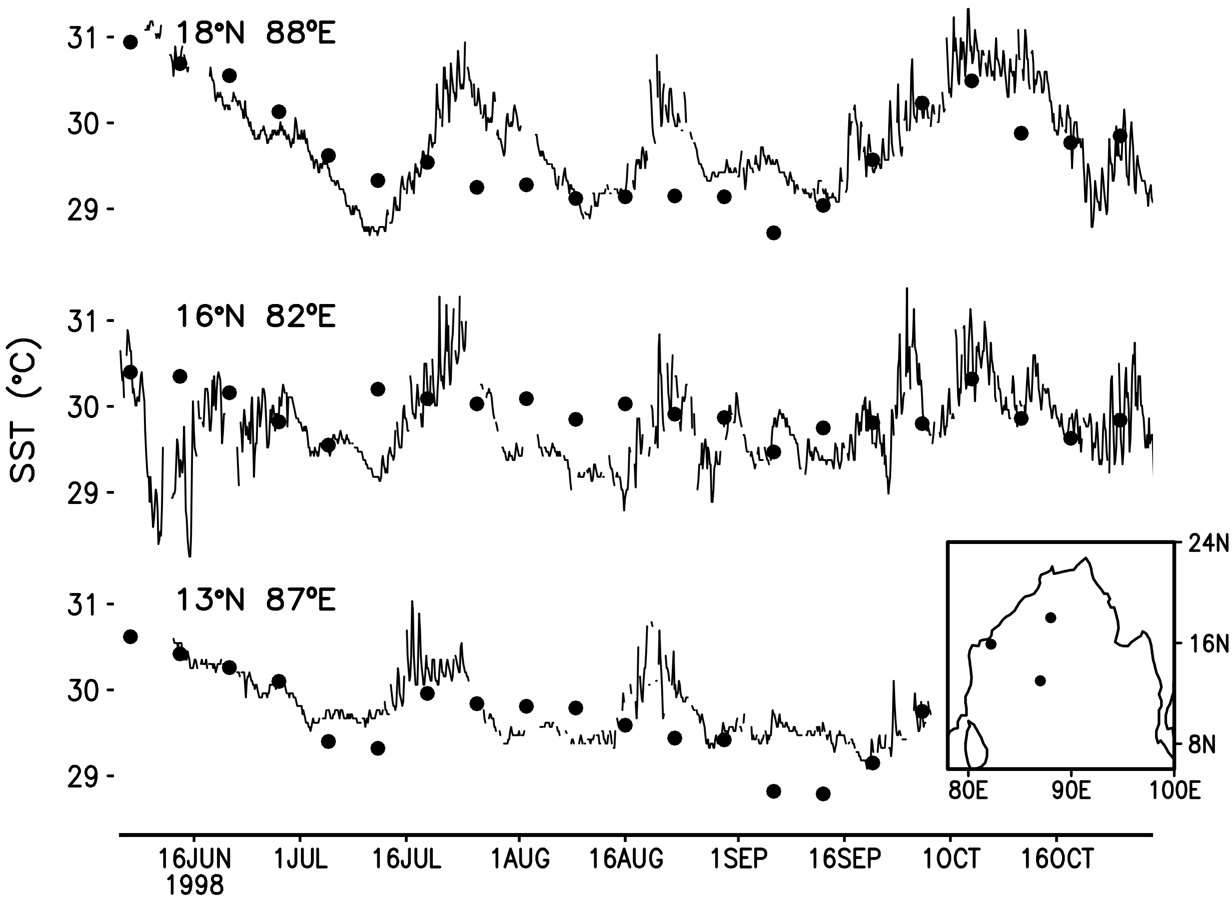

Figure 1. SST evolution at buoy locations in the western Bay of Bengal (inset). The Ganga-Brahmaputra river system discharges into the north Bay between 87°E and 92°E. Three hourly buoy SST (line), and weekly NCEP SST (dots) at grid points closest to the buoy locations; NCEP SST is on a 1°grid. Breaks in the lines indicate missing data. The SST sensor on the southern buoy failed after the third week of September.

Net heat flux at the ocean surface Q_{net} is the sum of the fluxes of sensible heat Q_{sen}, latent heat Q_e, net insolation Q_s and net longwave radiation Q_l. The turbulent components Q_{sen} and Q_e are estimated using bulk formulas [*Godfrey et al.*, 1998] based on daily buoy SST, air temperature T_a and wind speed U, and climatological summer relative humidity from the Comprehensive Ocean Atmospheric Data Set [*da Silva et al.*, 1994], with an exchange coefficient of 0.0013. Our estimate of net surface insolation Q_s is based on a simple empirical relation from *Shinoda et al.*[1998] which says that daily mean Q_s is approximately equal to 0.93 times daily mean OLR in Wm^{-2}. The relation was originally developed for the Indo-Pacific warm pool region within ten degrees of the equator. We use this formula at the buoy locations because it gives climatological June to October monthly mean Q_s that lie within 15 Wm^{-2} of Q_s from the International Satellite Cloud Climatology Project [*Bishop and Rossow*, 1996]. Our estimate of Q_l is based on an empirical relation that uses OLR as a proxy for cloudiness [*Shinoda et al.*, 1998]. Using the convention that positive (negative) fluxes are into (out of) the ocean, daily Q_s generally lies between 100 to 250 Wm^{-2}, daily Q_e between -80 to -250 Wm^{-2}, and daily Q_l between -15 to -45 Wm^{-2}. The magnitude of Q_{sen} is generally less than 15 Wm^{-2}. All four components of heat flux are modulated on intraseasonal time scales (not shown). Q_{net} is positive in the quiescent phase of the monsoon because solar heating dominates, and negative in the convectively active phase because the sum of evaporative and longwave cooling dominates, as shown in Figure 3. Even after some smoothing, the peak to peak variation of Q_{net} during an ISO can be larger than 200 Wm^{-2}. It appears that monsoon ISO of winds, surface heat flux and SST in the Bay of Bengal in 1998 have larger amplitudes than characteristic ISO in the west Pacific warm pool [*Godfrey et al.*, 1998]. The empirical estimates of the different components of heat flux can have errors of a few tens of Watts per square metre in weekly mean (e.g. [*Shinoda et al.*, 1998]). Since Q_{net} changes sign on intraseasonal time scales, the summer mean net heat flux cannot be reliably estimated using the present data.

3. Origin of SST intraseasonal oscillations

Let the thickness of the upper mixed layer of the ocean be h. The daily mean temperature of this layer (SST for our purposes) can change in response to entrainment of water from below and advection by prevailing near surface currents, as well as to $Q_{net} - Q_h$, where Q_h is the flux of

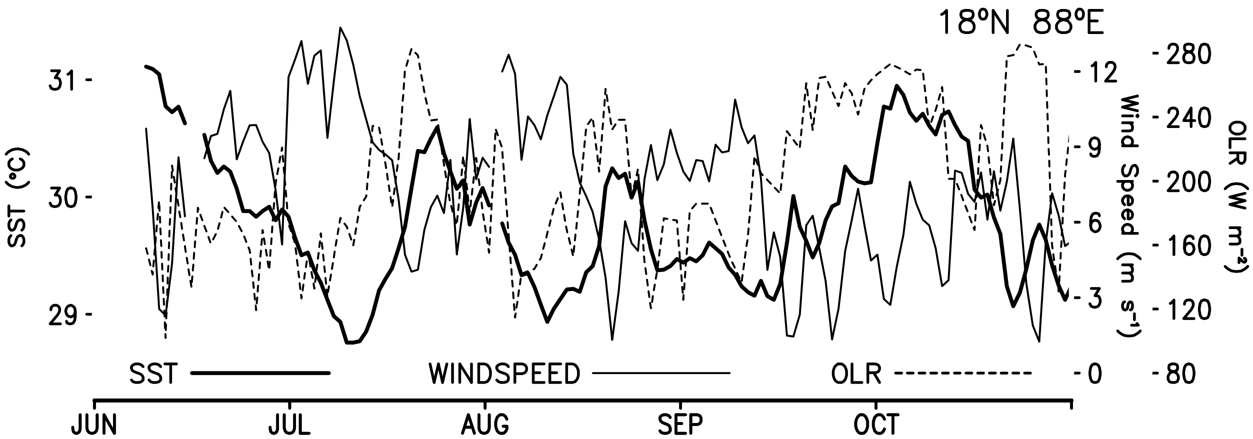

Figure 2. Monsoon fluctuations at the northern buoy. Daily mean buoy SST (bold), buoy wind speed (thin), and OLR at 17.5°N,87.5°E (dashed) the OLR data is on a 2.5°grid. Meteorological and oceanographic sensors on the buoys are located 3.2m above and 3m below the sea surface; winds were extrapolated to standard height of 10m.

solar radiation that penetrates below h and does not participate in warming the mixed layer [*Godfrey et al.*, 1998]. The simplest possible thermodynamics of the upper ocean is a balance such that the rate of change of SST is proportional to net surface heat flux. Such a balance can be expressed as $hC\rho\frac{\partial}{\partial t}SST = Q_{net}$ with constant h, where ρ is density and C the specific heat of seawater. Figure 3 suggests that this simple balance is valid to the first approximation in the Bay of Bengal during most of summer 1998. In the absence of subsurface profiles of temperature or salinity at the buoy locations, we do not know the time evolution of h. The scatter in the data is too large to reliably estimate h from a regression of Q_{net} versus $\frac{\partial}{\partial t}SST$, underscoring the errors in heat flux estimates and the possible contribution of other physical processes to SST evolution. Changes in entrainment rate might be expected to lead to changes in h as the wind speed fluctuates on intraseasonal time scales. Although signs of possible entrainment cooling can be discerned in Figure 3, it does not dominate SST evolution during every episode of strengthening winds. This is consistent with the presence of a barrier layer in the Bay of Bengal in summer [*Sprintall and Tomczak*, 1992] with relatively fresh water in the mixed layer and a deeper, more saline isothermal layer.

Note that the simple balance discussed above is not even approximately valid at the northern buoy from mid September to the end of October, when the mean Q_{net} is more than 80 Wm^{-2} into the ocean. In July-August, SST at the northern buoy rises by more than 0.1°C per day when Q_{net} is 80 Wm^{-2}. Therefore one might expect SST to rise by several degrees in the last six weeks of the record, but it does not (Figure 1). Salinity data from the buoy suggests a simple resolution of this discrepancy.

During summer, the north Bay of Bengal receives a large amount of fresh water from rivers, mainly the Ganga and Brahmaputra, and from the excess of rainfall over evaporation . The lighter, relatively fresh water forms a shallow pool in the north Bay. Surface ocean current from numerical ocean models (e.g. [*McCreary et al.*, 1993]) and ship drift show generally eastward Ekman flow in the north Bay during the monsoon. Surface currents at the northern buoy are towards the east in the monsoon months (Figure 4). This surface flow confines the fresh pool to lie to the east of the northern buoy until the end of August. As the monsoon winds weaken in September, the fresh pool engulfs the buoy, giving rise to very low surface salinity (Figure 4). The following arguments, however, are independant of the exact route taken by the fresh water. Low surface salinity gives rise to extremely stable density stratification that inhibits vertical mixing, leading to a shallow mixed layer in the north

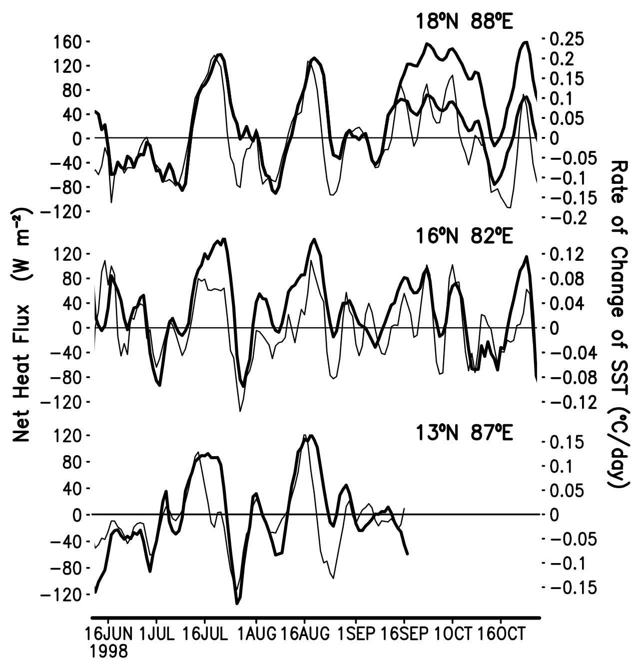

Figure 3. Heat balance at buoy locations. Evolution of Q_{net} (bold) and $\frac{\partial}{\partial t}SST$ (thin); both quantities have been slightly smoothed using a five day running mean; note the different scales for the latter. Two heat flux time series are shown at the northern buoy after 15 September - the upper curve is Q_{net}, while the lower curve represents Q_{net} "corrected"for solar penetration by multipying Q_s by a factor 0.65 (see text). The correlation between smoothed Q_{net} and $\frac{\partial}{\partial t}SST$ is about 0.8 at the northern buoy and 0.7 at the other two locations.

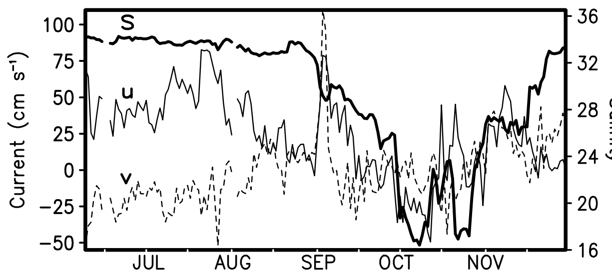

Figure 4. Salinity evolution at the northern buoy. Surface salinity (S; bold), and eastward (u; thin) and northward (v; dashed) surface ocean current at 18°N, 88°E from 10 June to 30 November 1998.

Bay [*Shetye and Gouveia*, 1998]. In clear water, about 35 percent of the net surface insolation penetrates below five metres [*Dickey and Simpson*, 1983]. It can be inferred from the OLR (Figure 2) that the average Q_s over the last six weeks of our record at 18°N, 88°E is about 220 Wm^{-2}. We propose that at this time h is about five metres, and assume that the water in the mixed layer is optically clear. Q_h is then about 75 Wm^{-2}; there is no net SST change between mid-September and the end of October because $Q_{net} - Q_h$ is close to zero. Agreement between $\frac{\partial}{\partial t}SST$ and Q_{net} "corrected" for solar penetration (Figure 3) offers support to our hypothesis. Note that enhanced entrainment is unlikely during this period; nor is sustained advective cooling because the surface current changes direction (Figure 4). However, the relative importance of these physical processes [*Moshonkin and Harenduprakash*, 1991] cannot be quantified with the present data.

The present observations show that it is important to monitor the ocean and atmosphere in the monsoon region using *in situ* measurements as well as satellites. Finally, the large intraseasonal signals in convection, surface heat flux and Indian ocean warm pool SST together present a clear target for models of the monsoon coupled air-sea system.

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