The Joint Air-Sea Monsoon Interaction Experiment (JASMINE): Exploring Intraseasonal Variability in the South Asian Monsoon

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Abstract

The implementation plans for the international Climate Variability and Predictability Programme (CLIVAR) calls for an expansion of domains of investigation from the Pacific Ocean to other regions where predictable elements of the climate system may reside. The monsoon of Asia-Australia are identified as a key phenomena which influences climate in other parts of the globe and whose prediction would bring immeasurable benefits to a large sector of the world's population.

This paper describes an extensive pilot study, the Joint Air-Sea Monsoon Interaction Experiment (JASMINE), held in the Indian Ocean during the summer of 1999 and presents some initial results. The experimental design was based on the precept that the intraseasonal variability of the monsoon, as it sways back and forth from active to inactive (or break) phases, is an important component of the monsoon system and one that is poorly sampled and simulated with great inaccuracy. JASMINE is the first comprehensive study of the coupled ocean-atmosphere system in the eastern Indian Ocean and the southern Bay of Bengal. The scientific objectives of JASMINE are to:

(i) Document of the boundary layer and surface fluxes throughout the transition from the active to the break periods of the monsoon.
(ii) Document components of the upper ocean heat, salt and momentum budgets as well as upper ocean equatorial current structure during active and break periods of the monsoon.
(iii) Acquire statistics of surface turbulent and radiative fluxes to enable a comparison with previous tropical western Pacific field programs.
(iv) Acquire statistics and documentation of convection in order to understand the nature and organization of cloud systems during active and break periods of the monsoon and to compare these statistics with convective organizations in other dynamic regimes, as documented in earlier observational studies in the tropics.

Two research vessels were used to gather data: the NOAA Ship *Ronald H. Brown* and the Australian Research Vessel *Franklin*. There were three components to the pilot study with cruises in April, and May-June aboard the *Ronald H. Brown* and in September aboard the *Franklin*. The cruise paths were made up of a combination of transects and on station periods. Overall, 388 CTD casts, and 272 radiosonde ascents were made. In addition, both ships carried identical flux systems to measure accurately the interaction of the ocean and the atmosphere. The *Ronald H. Brown* was equipped with five separate radar systems and profilers including a cloud radar and a C-band rain radar. All radars operated continually throughout JASMINE. During

JASMINE, METEOSAT-5 satellite was positioned over the central Indian Ocean providing, for the first time, accessible high-frequency geostationary imagery. Observations made during the field phase of JASMINE sampled both prolonged break and active phases of the monsoon resulting in an unprecedented amount of research quality data defining the atmospheric and ocean state during intraseasonal transitions of the monsoon. Although analysis of the data is only in an early stage, some initial results are presented.

1. Introduction:

During the last decade considerable progress has been made towards identifying the coupled ocean-atmosphere modes responsible for the El Niño-Southern Oscillation (ENSO) phenomena initially under the auspices of the international Tropical Ocean-Global Atmosphere Programme (TOGA). Based on this success, the fundamental objective of the U. S. contribution to the Global Ocean-Atmosphere-Land System program (GOALS) (NRC 1998) of the international CLIVAR (Climate Variability and Predictability Programme) is to extend the domains of primary interest from the tropical Pacific Ocean, which was the region of major interest in TOGA, to other regions of the tropics and higher latitudes:

> *"… A compelling reason for this expanded domain is the need to understand and quantify the effect of fluctuations in other major heat sources and sinks of the tropics and subtropics of the global general circulation and thereby improve predictions of weather and climate in both local and remote regions --- including the higher latitudes ..." (NRC,* 1998).

The Asian-Australasian monsoon, and the Indian Ocean-western Pacific region in general, is identified as a major system requiring special attention (NRC, 1998). The Asian-Australasian monsoon region embraces a large sector of the globe, spanning the tropics and subtropics of the eastern hemisphere. The annual variation of rainfall and temperature dominate thriving agrarian economies and social structures. Whereas the summer rains reoccur each year, they do so with sufficient variability to create periods of relative drought and flood throughout the region. Within this region resides the majority of the population of the planet and contain nations with the most rapidly growing populations. Forecasting the vagaries of the monsoon from year-to-year, variations that impact agriculture and water resources, are an extremely high priority for humankind and are a central issue of CLIVAR and GOALS.

Despite the identification of the Indian Ocean and the monsoon as critical areas of research, there is very little data available compared to other regions in the tropics. Except for climatologies sea-surface temperature (SST) and sea-surface salinity, little is known about the state of the upper Indian Ocean. Despite a few field campaigns (e.g., INDEX held during the period 1975--1976, Hastenrath, 1994; MONEX during 1978--1979, Krishnamurti, 1985; the WOCE Indian Ocean Expedition[1]), there have been relatively few direct measurements of the Indian Ocean compared to the other tropical oceans. Important surface climatologies have been

[1] See http://www-ocean.tamu.edu/WOCE/general.html

compiled from ship observations including surface fluxes (e.g., Oberhuber, 1988) but research quality observations are rare. A recent exception is the data from the research cruise in the Bay of Bengal (Hacker et al. 1998).

At the St. Michael's workshop on the Variability of the Asian-Australasian Monsoon[2] held in July 1998, three regional foci for the Asian-Australian monsoon were emphasized. Listed by priority, these regions are (i) The eastern Indian Ocean, the Bay of Bengal and the western Pacific Ocean, (ii) The Indonesian Archipelago, the Indonesian seas and the connection between the Indian and western Pacific Ocean (the throughflow) and (iii) The Arabian Sea and southern Indian Ocean. It was noted that all three regions are important to the overall structure of the monsoon and its variability but priority was determined principally by the lack of research activity in the particular region. Processes occurring in the eastern Indian Ocean, however, appeared to have a direct impact on the variability of the monsoon, especially the intraseasonal aspects. In addition, the coupled ocean-atmosphere structure of the eastern Indian Ocean has received least observational attention of all of the three regions.

There are other scientific reasons for an immediate emphasis on the eastern Indian Ocean. The eastern Indian Ocean, for example, is an important oceanographic area. It contains the entrance of the Indonesian throughflow, which connects the Pacific Ocean and the Indian Ocean. It is a critical region in the salt balance of the entire basin where large scale mixing occurs of the very saline waters between the western Indian Ocean with the fresh water of the eastern basin. It is also a region of considerable dynamic activity where upper ocean currents reverse on seasonal and intraseasonal time scales. The region also contains a pole of the newly discovered Indian Ocean dipole (Webster et al. 1999, Saji et al. 1999, Yu and Rienecker 1999, 2000). There is also some observational evidence of the existence of an oceanic barrier layer (Hacker et al. 1998) as originally documented for the Pacific Ocean by Lukas and Lindstrom (1991). From an atmospheric perspective, it is the region of maximum precipitation with extrema occurring over the northern Bay of Bengal and in the winter hemisphere just to the south of the equator. In fact, MSU satellite precipitation estimates indicate that the northern Bay of Bengal receives the largest mean monsoon precipitation in South Asia. The eastern basin appears to be the location where the monsoonal intraseasonal oscillations (MISOs) of the monsoon reach their maximum amplitude and a source region for active and break periods of the monsoon. The biennial oscillation possesses strong signals in the eastern Indian Ocean (Meehl 1994) and is a location of strong interannual variability of rainfall. Yet, while many of these features are known to be climatological features of the monsoon (see the reviews by Godfrey et al. 1995, Webster et al. 1998), the physical processes that maintain them are not understood.

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[2] Report of the St. Michael's meeting may be found at paos.colorado.edu/~webster

Given the successes of TOGA and the scientific advances in understanding the coupled ocean-atmosphere system, we believe that the field is well-poised to make significant advances in understanding the coupled ocean-atmosphere-land processes that may cause the variation of the monsoon from year-to-year, as well as the intraseasonal variability that exists within a monsoon season. For example, forecasting interannual variability of the monsoon could allow the prediction of strong, weak or normal monsoons, in terms of their integrated rainfall, with sufficient lead-time to allow remedial actions by a user community. Forecasting intraseasonal variability of the monsoon, on the other hand, is important for the following season:

> *"... Intraseasonal variability in monsoon rainfall has strong social and agricultural consequences and is the direct interface of society with the monsoon. Ploughing and planting periods are extremely susceptible to intraseasonal variation in monsoon rains. Once the monsoon has commenced, the timing of the first monsoon break becomes critical. Even if the average seasonal monsoon rains are normal, an ill-timed cessation of rainfall can be devastating to a local economy ..." (Webster et al. 1998)*

Webster et al. (1998) noted that on the 30--60 day time scales, convection first appeared in the equatorial regions of the western Indian Ocean from where it propagated eastward towards Sumatra before bifurcating as distinct propagating troughs, propagating poleward on either side of the equator. When the convection is located at the equator in the eastern Indian Ocean, the monsoon over South Asia is quiescent, or in a break phase. The northward moving element creates the active period of the monsoon over south Asia taking some seven days to propagate from the equator to the northern Bay of Bengal. The southward moving trough produces the near-equatorial rainfall maximum south of the equator eventually leading to enhanced rainfall over southern Australia (Fasullo and Webster 1999, Webster et al. 1998).

It would seem that a perquisite for the systematic forecasting of monsoon variability using numerical techniques is the ability to model the evolving annual cycle of the monsoon in both the atmosphere and the ocean, in addition to understanding the interaction of dry and moist land surfaces with the atmosphere and the interaction of the ocean and atmosphere in the pre- and post-onset periods of the monsoon. In order to gauge the success of models in the monsoon regions, two questions arise: (i) How well do atmospheric models simulate the mean summer monsoon rainfall over South Asia and India? (ii) How well do models simulate the intraseasonal variability of rainfall during a particular monsoon wet season? Good simulations of the monsoon have proven to be elusive and are, in fact, worse than most regions of the tropics. Simulations by different models using the same boundary conditions in the Atmospheric Model Intercomparison

Program (Sperber and Palmer 1996) are very different from one model to another (Figure 1). Furthermore, differences between members of an ensemble of realizations using the same model are also quite different (not shown). Simulations of western Pacific Ocean precipitation, on the other hand, are much more similar from one model to another. Thus, it is probable that either unique model problems or phenomena that are not well modelled, exist in the monsoon regions that create the divergence of model results.

The failure of atmospheric models to simulate the distribution of monsoon precipitation within any one year or from year-to-year, even with stipulated lower boundary conditions, is troublesome and does not bode well for the numerical prediction of monsoon variability. Model problems aside, there are a number of important questions regarding the predictability of the monsoon. Shukla (1987) raised two such questions:

(i) Is the behavior of the seasonal mean monsoon determined primarily by a statistical average of a variety of independent short-period fluctuations that are not related to any seasonal or other low-frequency forcing?

(ii) Are there global- and planetary-scale "forcing functions" (either due to slowly varying boundary conditions at the Earth's surface or to very low frequency changes like the Southern Oscillation and ENSO) that determine the interannual behavior of the seasonal mean monsoon circulation and rainfall, and is the interannual variability of the short period fluctuations controlled by such large-scale low-frequency forcing?

Two more questions may be added:

(iii) Are there low-frequency hydrodynamical instabilities of the monsoon circulation themselves that, on the one hand, exert control on the short term fluctuations (e.g., instabilities associated with the intraseasonal variability of the monsoon) described in question (i), above, and are, in turn, modified by large-scale planetary controls described in question (ii)?

(iv) What is the influence on monsoon variability of local Indian Ocean basin lower boundary circulations such as the Indian Ocean SST variations identified by Clark et al. (2000), and the Indian Ocean "dipole" described by Saji et al. (1999), Webster et al. (1999) and Yu and Rienecker (1999, 2000).

The four questions listed above are important as they define the very essence of the limits of predictability of monsoon variability. The degree to which monsoons are predictable depends upon the space scale and timescale for which predictability in question. For example, the day-to-day changes in rainfall at any place are determined by the life cycle of synoptic scale disturbances (lows, depressions, monsoon trough, etc.) at or around the place of consideration. Thus, the problem of predictability of day-to-day weather at a point in the monsoon region is not different from that of any other point on the globe, and is limited to a few days. If it turns out that the monsoon is controlled by these short term weather events then there would be no greater predictability for the monsoons than is found in the extratropics. If the monsoon is controlled solely by large-scale evolving processes such as those listed in (ii) and/or (iv) above, one would expect that the limits of monsoon prediction would be given by the limits of predictability of these planetary-scale forcing functions. If there are large-scale and low-frequency instabilities of the monsoon which might, for example, determine the intraseasonal active and break periods of the monsoon, then one may expect hybrid limitations to predictability existing somewhere between the limits defined by (i), and (ii) and (iv), described above. That is, predictability would be limited to the time scale of the instability.

Relative to these points regarding the predictability of the monsoon, a number of seemingly ambiguous points can be made:

(i) The AMIP integrations show significant simulation degradation in South Asian region at a level higher than elsewhere in the tropics (Figure 1).

(ii) There is a moderately good relationship between the SOI and rainfall over India, but it is highly variable. During certain decades, the correlation between SOI and Indian monsoon rainfall may be as high as --0.8 but falling to insignificant levels (< --0.2) during other decades (Webster et al. 1998, Torrence and Webster 1998). Even within the decades of relatively high correlations there are drought (flood) years in India that appear independent of ENSO variability.

(iii) Sadhuram (1997), Hazzallah and Sadourny (1997) and Clark et al. (2000) found that correlations as high as +0.8 occurred between equatorial Indian Ocean SSTs in the winter prior to the monsoon wet season but these correlations have flagged since 1976, with the ocean region of highest correlation (still statistically significant) shifting from one part of the Indian Ocean to another. Seasonal monsoon predictions by the India Meteorological Department, now made for over 100 years using models based on empirical relationships between monsoon and worldwide climate predictors (including some of the SST precursors used by Sadhuram 1997, Hazzallah and Sadourny 1997, and Clark et al. 2000), are relatively skillful (Gowariker et al., 1989). The success of the empirical forecasts would tend to support the theoretical basis that long-term predictability of the monsoon exists whereby slowly evolving boundary forcing (e.g., the SST in the Pacific Ocean, Eurasian snow cover, Indian Ocean SST) imposes some large-scale control on the monsoon as suggested by Charney and Shukla (1981).

(iv) Anomalously wet or dry monsoon seasons in both the Asian and Australian summer monsoons (see tables 3a, b of Webster et al. 1998) are season-long events. That is, most wet or dry seasons consist of months that are anomalously wet or dry. Thus, whatever the reasons for the anomaly in monsoon rainfall they persist throughout the summer season. Lawrence and Webster (2000) have shown that this persistence of rainfall through a season is associated with the relative absence of intraseasonal variation of the monsoon (anomalously wet monsoon) or their abundance (dry monsoon).

The seeming ambiguities listed above suggest two possible interpretations of the AMIP results shown in Figure 1. One possibility is that the divergence of model simulations in AMIP in the Indian Ocean region is due to the influence of hydrodynamical instabilities of the system such as those that might be associated with MISOs. These low frequency oscillations may be chaotic (Palmer 1994) and if this were the case, the interannual variability of the monsoon would be essentially unpredictable beyond the intraseasonal timescale except for gross-scale influences of boundary effects such as El Niño or the Indian Ocean dipole, described above. Collectively, the predictability of the seasonal monsoon would be a "nudged chaotic system" (Palmer 1994, Webster et al. 1998) whereby the predictability of abundant or deficient seasonal rainfall depends on the large scale forcing to increase the probability of active or break periods within a monsoon season. This theory would conclude that the predictability of seasonal monsoon rainfall is probabilistic. However, the forecasting of individual MISOs (i.e., prediction on the time scale of the inherent instability of the monsoon) may be deterministic.

Another plausible explanation for the poor AMIP simulations of the monsoon is that the complex geography and orography and the tight heating gradients in the monsoon system may tax the current atmospheric models to their limit. Some credence is added to this hypothesis by noting that the difference in estimates of monsoon precipitation shown in Figure 1 appear to be shifted systematically from one another. That is, some models persistently overestimate precipitation in certain monsoon regions while the others tend to provide underestimations from one year to the next. These systematic differences between the models may be related to a fundamental problem of atmospheric (and coupled) models: the inability to simulate intraseasonal variability. For example, some AMIP models remain in a perpetual break monsoon configuration (therefore underestimating rainfall over India and South Asia but overestimating precipitation over the equatorial ocean regions). Others tend to forecast prolonged active phases with rainfall overestimated over South Asia but underestimated nearer to the equator. Almost all AMIP models, and more recent versions of the same models used in the AMIP exercise, do not show evolving MISOs at all.

Despite which of the two classes of possibilities described above is proven to be correct, or in what combination and degree they are mutually correct, the simulation of intraseasonal variability of the summer monsoon emerges as a major issue. This priority arises because of the need to forecast the evolution of intraseasonal variability for practical considerations, its possible role in determining the interannual variability of the monsoon, and from the fundamental failure of models to simulate intraseasonal phenomena.

The fundamental lack of observations in the Indian Ocean sector limit our ability to study processes that determine the extent, duration and intensity of intraseasonal variability and its relationship to processes in the ocean. Even the understanding of some zeroth order processes in the upper ocean is absent. For example, whereas it appears that an upper ocean barrier layer exists in the eastern Indian Ocean (Hacker et al. 1998), at least during the boreal winter, its annual cycle, variability or extent has not been documented. These gaps limit understanding of the Indian Ocean because we may expect the barrier layer to be an important phenomena to a wide range of processes, as may be inferred by clear evidence from the Pacific Ocean warm pool. As described by Vialard and Delecluse (1998):

> "*... The overall effect of the salinity stratification is to retain heat and momentum in the upper layer of the western Pacific by restraining the exchanges with the cooler waters from below and from the central Pacific (by subduction)..."* (Vialard and Delecluse, 1998).

The advection and mixing caused by westerly wind bursts (manifestations of intraseasonal variability in the western Pacific) cause the barrier layer to shift eastward. Thus, the lateral movement of the barrier layer is extremely important in moving the impacts of an insulating layer to other regions of the warm pool. Furthermore:

> "*... these properties of the barrier layer structure might favor the growth of unstable air-sea interactions in the central Pacific Ocean after a westerly wind burst* ..." (Vialard and Delecluse, 1998),

suggesting that changes of the barrier layer are key features in coupled ocean-atmosphere instabilities.

There are four science questions relating to MISOs:

(i) What is the role of intraseasonal variability in the overall annual cycle of the monsoon?

(ii) Does the degree of intraseasonal variability in a particular monsoon season determine whether or not the summer precipitation is above or below average?

(iii) To what degree is the intraseasonal variability of the monsoon a coupled ocean-atmosphere phenomena? Does the MISO modulate the structure of the Indian Ocean barrier layer and is it an important component of the heat balance of the tropical Indian ocean?

(v) Are there consistent and coherent patterns of intraseasonal monsoon variability such that there may be a canonical MISO? What is the relationship of the onset of the South Asian monsoon and intraseasonal variability? Does viewing the monsoon onset as the first summer MISO increase the possibility of the prediction of the onset?

The question should be asked whether or not the points raised above can be inferred from the results of other tropical experiments such as occurred in the Pacific Ocean during TOGA (e.g., TOGA Coupled Ocean-Atmosphere Response Experiment: TOGA COARE: Webster and Lukas 1992). In the TOGA field experiments considerable effort was expended in determining the joint variability of the upper ocean and the atmosphere, the heat balance of the upper ocean, the energy and momentum fluxes between the ocean and the atmosphere, and the relationship between tropical convection, surface conditions and atmospheric dynamics. The question of transferability of these experimental results to the Indian Ocean is unclear. For example it is not known if the Indian Ocean barrier layer performs the same role in the Indian Ocean as in the Pacific warm pool. Nor are there comparative estimates of the mean heat balance of the Indian Ocean and Pacific Ocean warm pools and it is not known if these heat balances are maintained in the same manner in the two basins. Finally, it is not known if the relationships between SST, convection and dynamics are the same in each reason. Such questions can only be answered by significantly extending the observational base in the Indian Ocean.

This document reports on a new and significant step in identifying and measuring fundamental coupled processes that exist in the monsoon system. Specifically, we discuss the Joint Air-Sea Monsoon Interaction Experiment (JASMINE) held in the eastern Indian Ocean during the summer of 1999. For the reasons discussed at some length above, the major concentration of JASMINE was on the intraseasonal variability of the monsoon. To this end, JASMINE measured the heat and momentum fluxes between the atmosphere and the ocean, the state of the upper ocean, and the organization of convection and its interaction with the larger scale environment for a total of 52 days in the southern portions of the Bay of Bengal and the equatorial eastern Indian Ocean. JASMINE followed the Indian Ocean experiment (INDOEX: e.g., Meywerk and Ramanathan 1999, Mitra 1999, Moorthy and Saha 2000) which was designed to study anthropogenic aerosol distributions in the Indian Ocean during the winter and spring (e.g., Krishnamurti et al. 1998). The principal platforms used in JASMINE were the research ship NOAA Ship *Ronald H. Brown* for the spring and early summer and the Australian *R/V Franklin* in September. In July 1999, Indian scientists conducted a national joint meteorology-oceanography experiment (BOBMEX: Bay of Bengal Monsoon Experiment[3]) in roughly the same location as JASMINE. During 1999, the European geostationary satellite METEOSAT-5 was relocated over the central Indian Ocean and provided an unprecedented view of the monsoon through the INDOEX-JASMINE-BOBMEX period. In summary, the summer of 1999 was a unique year of observation in the eastern Indian Ocean.

The paper is organized in the following manner. In the next section a scientific basis for JASMINE is presented. The scientific objectives of JASMINE are described in section 3 including a brief description of the implementation plan. Section 4 presents some of the initial results of the experiment. A description of the available data and where to find it is given in section 5. Finally, some initial conclusions are summarized together with plans for future experiments in the Indian Ocean region.

(2) Science Issues:

In the following paragraphs, we lay out a scientific basis for JASMINE. To accomplish this, observational and modeling results of the large scale South Asian monsoon and its intraseasonal variability are discussed. However, it should be noted that these results are tentative and depend upon data from atmospheric numerical models (e.g., reanalysis products) which contain limited initial data from the Indian Ocean region, ocean models forced by reanalysis products, and interpretations of satellite data (e.g., the Reynold's SST data set) that have not been substantiated to a large degree by in situ data.

a. Differences and similarities between Indian and Pacific Ocean regimes

A first-order comparison between the Pacific and Indian Ocean regimes can be accomplished by comparing the distributions of convection (OLR: W m^{-2}) and SST (˚C) in the two regions (Figure 2). There are fundamental differences. In the Pacific Ocean there is a direct relationship between the SST and OLR with the intensity of convection increasing (note the inverted OLR scale) relatively uniformly with increasing SST. However, the relationship fails where the SST is warmest. This is probably due to intraseasonal oscillations (or Madden-Julian oscillations) in the western Pacific Ocean where prior to the disturbed period of the oscillation,

[3] Professor S. Gadgil, Indian Institute of Science, Bangalore, India: personal communication.

cloudiness is inhibited by dynamic subsidence and the SST is warming (Lau and Sui 1997, Fasullo and Webster 1999). In both the eastern and western Indian Ocean, there is an absence of a relationship between SST and convection over the entire SST range. In fact, the overall relationships appear similar only to that found in the warmest parts of the Pacific Ocean. Consequently, it appears that convection in the Indian Ocean tends to be related to strong dynamical processes rather than to the absolute value of the SST.

The second major difference between the Pacific and Indian Ocean basins is in the manner in which the heat balance is achieved. Figure 3a shows the mean heat balance of the North Indian Ocean as determined by an intermediate ocean model of McCreary et al. (1993) forced by climatological winds and thermal forcing derived from the NCEP/NCAR reanalysis (Loschnigg and Webster 2000). The left panel shows that the annual cycle of the heat balance of the North Indian Ocean is achieved by a strong oceanic northward heat flux across the equator during winter and a southward oceanic flux in the summer. During the boreal winter and early spring, a time of low winds and high insolation, the mean net flux into the North Indian Ocean is strongly positive. During spring, the direction of the ocean heat advection changes dramatically from +2 PW to ---2 PW over a two-month period. Loschnigg and Webster (2000) note that the heat flux is generally in the opposite sense to the sign of the divergent wind component of the monsoon flow which is to the north in summer and to the south in winter. They attributed the reversed seasonal cycle of heat flux to wind-forced Ekman transports. The annual cycle of the climatological heat balance of the Indian Ocean shown here is similar in form to that found by the observational studies of Hsiung et al. (1989) and Hastenrath and Greicher (1993) both of which calculated the heat transport by residual methods, and by Wacogne and Pacanowski (1996) who used an ocean general circulation model.

The right hand panel of Figure 3b shows the annually averaged heat balances of the northern Indian Ocean and cross-equatorial heat transport calculated using the same for a seven year period 1984 through 1990. There is considerable interannual variability in each component of the heat balance. Overall, there is a general balance between the excess surface flux into the North Indian Ocean (about +0.2 PW) and the southward transport across the equator (-0.2 PW). Near balances between the net heat flux and cross-equatorial heat transport can be seen in years 1984, 1985, 1986 1989 and 1990. The years 1987 and 1988, representing successively weak and strong summer monsoon seasons generally attributed to the influence of the El Niño in the first year and the La Nina in the second, are anomalous. The weak monsoon year possessed higher net surface fluxes and weaker southward transports of heat, presumably both associated with decreased cloudiness and generally lighter surface winds. With the reduction of southward cross-equatorial heat transport, heat storage in the upper ocean increased substantially. The following year, 1988, with the stronger winds and increased surface cooling and southward Ekman transport, the heat storage decreased. It is interesting to note however, that if components of the heat budget of the North Indian Ocean for these two anomalous years are averaged together, they balance out to values similar to years shown in the same sequence. This biennial compensations allowed Loschnigg and Webster (2000) to speculate that the coupled dynamics of the monsoon system act to regulate the monsoon system adding dynamical linkages to the arguments of Meehl (1994). However, for the purposes of this paper, the results of the numerical experiments point towards the conclusion that the monsoon system is a coupled ocean-atmosphere phenomena with seasonally reversing monsoonal flow and ocean heat transports in both the ocean and the atmosphere. It appears that the importance of the meridional ocean heat in the heat balance of the Indian Ocean is far greater than the role of heat transports in the western Pacific Ocean.

b. The mean circulation of the South Asian summer monsoon

Figure 4 shows the climatological mean atmospheric summer circulation (June-August) in the lower (850 mb) and upper (200 mb) troposphere using ECMWF data for the period 1979--1993. The mean summer rainfall rate (mm day^{--1}) is shown in the bottom panel. The precipitation fields were compiled using MSU satellite data.

The circulation patterns of the Indian and Pacific Oceans are very different. To a large degree, the mean atmospheric circulation over the Pacific may be thought of as a trade wind system converging into the western Pacific Ocean with return flow aloft in the form of an along-equator longitudinal Walker Cell (Krishnamurti 1971). The monsoon circulation, on the other hand, is strongly latitudinal. The seasonally reversing circulation reacts to the strong cross-equatorial pressure gradient force set up by the annual cycle of land-sea temperature difference resulting in an extremely strong and very complicated circulation system resulting in the strongest across-equatorial near-surface pressure gradient on the planet. The summer circulation shown in Figure 4 possesses the characteristic cross-equatorial flow concentrated in a lower tropospheric jet stream (the Somalia Jet) and diverges over the Arabian Sea. Strong convergence occurs in the eastern Indian Ocean, the Bay of Bengal and southeast Asia in the vicinity of the surface pressure trough (red lines) and the region of precipitation maxima. A secondary maximum of surface convergence occurs along a weak surface trough to the south of the equator. In the upper troposphere, strong anticyclonic outflow occurs over South Asia resulting in an easterly jet stream that extends across equatorial Africa. There is no similar easterly jet stream over the Pacific Ocean. During the winter monsoon, when maximum precipitation lies over northern Australia (not shown), an easterly maximum exists in the upper troposphere but it is much weaker.

Three principal precipitation maxima are apparent in Figure 4c: to the west of India against the Ghat mountains, over the northern Bay of Bengal, and to the south of the equator in the winter hemisphere. Although the largest rainfall occurs over the ocean areas, there is a secondary maximum that extends across southeast Asia linking to a further precipitation maximum over the South China Sea. Although the magnitude of the southern hemisphere precipitation maximum is less than the other maxima in the northern hemisphere, its existence is a vital clue towards understanding of monsoon intraseasonal variability.

c. Intraseasonal Variability of the Monsoon

Within the monsoon rainy season there are great variations of precipitation on time scales of 10 to 30 days. The long periods of rainfall are often referred to as active phases of the monsoon while the dry periods are known as break phases. In a particular region (e.g., India), the active monsoon phase is an envelope of rain-bearing disturbances while the monsoon break signifies long periods of little rain and infrequent disturbances.

Active and break periods of the monsoon are known for their impacts on local climates (e.g., Ramage 1971, Rao, 1976 Das 1986) because of their timing and the intensity of excessive rainfall or drought. However, they exist on very large spatial scales. Figure 5a shows composites of the anomalous precipitable water content[4]. in an atmospheric column for active and break periods of the monsoon. The criterion used to determine an active or a break period is based on precipitation over central peninsular India discussed in Webster et al. (1998). During an active period, above average and coherent PWC anomalies extend from the Arabian Sea to the dateline in a general northwesterly-southeasterly orientation, a distance of 10,000 km. The break situation shows the reverse configuration, with a PWC maximum extending along the equator and a minimum over South Asia. Intraseasonal variability, determined from criteria over a relatively small region of South Asia (in this case India), shows an extremely large-scale signature. Such scales of low-frequency variability are apparent in multiple data sets such as precipitation, OLR, and wind fields (e.g. Lawrence and Webster 2000).

Figure 5b shows time--latitude sections of blended MSU estimates of precipitation for two typical years (1985 and 1995) along 90˚E. This meridian bisects the Bay of Bengal and the southern hemisphere precipitation maxima centered (see Figure 4c). The data set blends surface

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[4] PWC Data provided by Vonder Haar and Randall, Department of Atmospheric Science, Colorado State University.

data and satellite MSU precipitation estimates[5]. A clear feature of the distributions is the arc or boomerang shaped precipitation maxima that appear to emanate from the equator and then propagate both northward and southward. These features appear every 20 to 40 days and have an apparent period of propagation from the equator to 30˚N of about 10 days. The apparent northward propagation of precipitation anomalies in the eastern Indian Ocean has been noted by a number of earlier studies (e.g., Sikka and Gadgil 1980, Yasunari 1980, Hartmann and Michelson 1989). Sikka and Gadgil (1980) and Gadgil and Asha (1992) noted that the migrations were longitudinally coherent from one side of the Indian Ocean to the other. However, all of these analyses concentrated on the northern hemispheric behavior of the MISO and apparently did not notice the southward leg of the bifurcation. Lawrence and Webster (2000) suggest that the bifurcating cyclonic vortices or troughs represent trailing Rossby waves behind a rapidly propagating Kelvin wave, in agreement with the observations of Wang and Rui (1990). The northward propagating convection associated with the vortex in the eastern Indian Ocean appear as consistent precursors of active phases of the South Asian monsoon.

The onset of the monsoon over South Asia may be thought of as the first major MISO or northward "propagation" of the precipitation from the equator. For example, the monsoon onset in 1985 follows enhanced precipitation at the equator that occurred around May 20. In 1995, the onset defined in this manner occurred about a week earlier. It should be noted that the onset of the monsoon, used here, is very different to the traditional definitions of the monsoon employed elsewhere (e.g., Ramage, 1971, Rao 1976, Das 1986). Most definitions refer to rainfall occurring near the southwest tip of India. However, in most years precipitation occurs in the northern Bay of Bengal much earlier than precipitation in the southeast of India. Subsequent discussion will show that precipitation along the east coast of India occurs with the acceleration of the monsoon gyre after the first major MISO matures and moves northward into the northern Bay of Bengal. Thus, viewing the monsoon from a wider basin-wide perspective allows the possibility of defining a monsoon onset in a larger geophysical context.

Some 35 monsoon MISOs were identified in the April-September period between 1979 and 1995. Composites have been computed relative to a "day 0", defined as the start of the intense precipitation at 90˚E at the equator and are shown in Figure 6a together with similarly constructed SST fields (Fig. 6b) and 950 mb NCEP wind fields (Figure 6c) as described by Kalnay et al. (1996). For the precipitation and SST fields, composites are shown 15 days before the maximum of precipitation at 90˚E to 15 days afterwards. For the wind fields, composites are shown for days --5, 0, +5 and +8 days. At --15 days, the Indian Ocean is warmer than average by

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[5] The data have been compiled at the Program in Atmospheric and Oceanic Sciences, University of Colorado. The data set is available on request to pjw@oz.colorado.edu.

about 0.5˚C, consistent with the analysis of Fasullo and Webster (1999). The formation of a precipitation maximum in the eastern-central Indian Ocean occurs along the equator near day --5. At this time, the SST is still above average and the monsoon gyre still anomalously weak and precipitation is relative weak over South Asia. Between day --5 and day 0 the equatorial precipitation maxima intensifies and moves eastward. The SST field at this stage is near average (zero anomaly), having cooled over the previous 10 days, probably as a consequence of the increasing strength of the monsoon wind gyre (Figure 6c). At day +5 the precipitation maxima has clearly bifurcated and large-scale lower tropospheric vortices have formed on either side of the equator. By day +10 the precipitation has increased over the entire North Indian Ocean and the monsoon gyre has increased substantially (i.e., enhanced clockwise circulation). South Asia, previously in monsoon break conditions, is now in an active phase with enhanced westerlies across the North Indian Ocean.

Figure 6c shows a subset of the composites for the low level monsoon wind field for days -5, 0, +5 and +8, constructed using the same precipitation criterion discussed above. Superimposed on the NCEP/NCAR 925 mb anomalous wind fields are anomalous precipitation fields > 8 mm day-1. The panels show a distinct oscillation of the interhemispheric anomalous monsoon wind gyre. The formation of the precipitation maximum in the east-central Indian Ocean (day -5) occurs with a anomalous weak and reversed monsoon suggesting a basin wide period of low wind speed and (from Figure 7b) high insolation in keeping with the earlier results of Webster and Fasullo (1998). As the precipitation maximum grows and moves eastward, strong convergence occurs along the equator (day 0) into the East Indian Ocean coinciding with the precipitation maxima (Figure 6b). Twin cyclonic flows can be seen to form to the north and the south of the eastern Indian precipitation. As the two off-equator cyclonic centers bifurcate the entire monsoon circulation accelerates increasing the cross-equatorial flow (day 5) until the monsoon circulation becomes fully developed (day 8). The acceleration of the circulation coincides with a distinct cooling of the entire Indian Ocean (Figure 7b). Not until the monsoon gyre accelerates does substantial precipitation occur along the east coast of India.

Intraseasonal variability is clearly an important component of any single monsoon season. However, does the intraseasonal variability have any impact on the interannual variability of the monsoon or is it merely just superimposed on a particular monsoon season? As discussed earlier, ensemble runs with numerical prediction models have shown that the principal components of the MISO map onto the principal components of interannual variability (Ferranti et al, 1997). The correspondence of the patterns suggest that interannual variability of the monsoon is closely tied to variations that occur on time scales less than a season thus implying that the MISO is the major component of interannual variability.

The characteristics of the MISOs help to explain the distribution of mean monsoon precipitation patterns shown in Figure 3. The maximum in the northern Bay of Bengal occur as a result of precipitation largely associated with the active phase of the monsoon when the MISOs are located in their most northern location. The precipitation maximum to the south of the equator results from the reoccurring southward migration of the cyclonic center. Its lifetime is relatively short-lived as it eventually propagates over cooler winter hemisphere water. However, their collective effects are sufficient to produce a significant contribution to the mean precipitation pattern of the southern Indian Ocean. There is also some evidence that these southern branches of the MISO are associated with cloud bands that extend across continental Australia and are responsible for roughly 40% of the wintertime Australian rainfall. Such associations were found by Nicholls (1989) on interannual timescales. Fasullo and Webster (1999) suggest that similar manifestations occur on the intraseasonal timescales providing further support for the Ferranti et al. (1997) hypothesis.

d. Intraseasonal variability of Indian Ocean processes

The strong variation in the atmospheric flow during the evolution of an MISO would suggest that should be a dynamic response by the ocean during the course of an MISO. In fact, Figure 6b and c show that there are strong SST and low-level monsoon wind variations associated with the MISO. It is unclear, though, whether or not these wind variations produce a dynamic response of the ocean that is important to the annual cycle of the Indian Ocean heat balance.

Numerical experimentation using the McCreary et al. (1993) intermediate ocean model, forced by the 40--year NCEP/NCAR reanalysis data set using 5--day average forcing indicated a substantial oceanic response on intraseasonal time scales. Figure 7 shows latitude-time north-south ocean heat fluxes averaged across the entire Indian Ocean basin and averaged through the depth of the model. The upper panel shows the mean annual cycle of cross-latitude heat flux from the 40 years of integrations, essentially extending in latitude the cross-equatorial heat transports shown in Figure 3a. During the winter there is a heat flux to the north of order 2 PW and a slightly stronger flux southwards during the boreal summer. Maximum amplitudes of heat transport occur at about 5˚S. The two lower panels show heat transports in the same format but for individual years 1987 and 1988. These were the two years, noted in reference to Figure 3b, that corresponded to weak and strong monsoon years, respectively. Both years possess low frequency modulation of north-south heat transports during both summer and winter, especially during the former season on the 20--40 day or MISO time scale. During the summer the magnitude of the southward heat transport varies between 0 and --4.5 PW, compared to the climatological maximum of 1.5--2 PW. In addition, the intraseasonal variability in each of the two years are very different. Thus, the differences in annually averaged cross-equatorial heat transports noted in Figure 3b between these two years are also apparent in entirely different intraseasonal structure. Whereas it is premature to speculate on a potential feedback from the ocean to the atmosphere related to the differential heat transports, it is probably safe to say that the intraseasonal variability of ocean heat transport is an important component of the annual heat balance of the Indian Ocean.

e. Summary

The preceding discussion pointed out a number of important issues. Foremost among these are that there are clear differences between the Indian and Pacific Oceans as expressed by differences in SST-convective relationships, and the relative magnitudes of the components of the ocean heat balance. In addition, strong intraseasonal variability exists in the atmospheric monsoon flow of both basins but the form of the intraseasonal oscillations is different. In both basins the intraseasonal variability appears to be coupled to upper ocean processes, as seen in the variability of the SST, and in the response of an ocean model. Yet, it is less clear if the role is the same. Given the differences between phenomenology in the Indian and Pacific Ocean regimes, the most important conclusion we can make at this stage is that the results of process experiments in the western Pacific Ocean probably cannot be directly transferable to the Indian Ocean monsoon regime. The main purpose of JASMINE, then, is to provide elemental observations of the coupled ocean-atmosphere system in the Indian Ocean, especially on intraseasonal time scales.

3. The Implementation of JASMINE:

a. The scientific objectives

With the questions raised in the last section in mind, the JASMINE pilot study in the eastern Indian Ocean-Bay of Bengal was designed and implemented. The scientific goals of JASMINE were to:

(i) Document the boundary layer and surface fluxes throughout the transition from the active to the break period of the monsoon.

(ii) Document components of the upper ocean heat, salt and momentum budgets as well as upper ocean equatorial current structure during active and break periods of the monsoon.

(iii) Acquire statistics of surface turbulent and radiative fluxes to enable a comparison with previous tropical western Pacific field programs.

(iv) Acquire statistics and documentation of the convection in order to understand the nature and organization of cloud systems during active and break periods of the monsoon and to compare with convective organizations in other dynamic regimes as documented in earlier observational studies in the tropics.

JASMINE evolved as a bilateral effort involving the United States and Australia. In the United States, JASMINE was funded by the National Science Foundation's Divisions of Atmospheric Sciences and Oceanography and NOAA's Office of Global Programs. The Australian component was supported by the Commonwealth Scientific Industrial Research Organization (CSIRO) Division of Marine Sciences.

b. Participating groups

Four U.S. research groups collaborated in the planning and execution of JASMINE. Each group had specific responsibilities. The University of Colorado (Principal Investigator: P. J. Webster) and the NOAA Environmental Technology (ETL: PI, C. E. Fairall) were responsible for the measurement of the air-seas fluxes during the intraseasonal transitions of the monsoon. The University of Hawaii (PIs: P. Hacker, R. Lukas and E. Firing) measured and documented the upper ocean heat, salt and momentum budgets. The University of Washington group (PIs: R. Houze Jr., S. Yuter and Y. Serra) documented the evolution of convection and its relationship to monsoon transitions. Both the University of Colorado/NOAA-ETL and the University of Washington groups maintained an upper air sounding program throughout JASMINE. These four principal research groups used the NOAA Research Ship *Ronald H. Brown* as the principal observing platform.

The fifth research group participating in JASMINE was the Australian CSIRO Marine Sciences Division (PIs: J. S. Godfrey and F. Bradley) which used the Australian ship *R/V Franklin*.. They were responsible for the measuring the upper ocean heat, salt and momentum budgets, the air sea fluxes and the atmospheric structure during the late summer monsoon.

c. Instrumentation and ancillary data sets

JASMINE featured an ensemble of instruments to make measurements in the ocean and atmosphere using a combination of in situ and remote sensing methods. A summary of the instrumentation on the *Ronald H. Brown* is given in Table 1. Collectively, measurements of the atmosphere and the atmosphere/ocean interface were made by GPS rawinsondes, bulk near-surface meteorological measurements, air-sea turbulent fluxes, radiative fluxes, numerous rain gauges, three profiling Doppler radars, microwave and IR radiometers, a cloud ceilometer, and a scanning C-band Doppler precipitation radar. The state of the upper ocean was measured using CTD (conductivity, temperature and depth) instruments, near-surface thermosalinographs and current profilers. These measurements are addressed in more detail below.

1) OCEAN-ATMOSPHERE INTERFACE MEASUREMENTS

The ETL ship-based air-sea interaction system was used for bulk meteorology, radiative, and turbulent fluxes with additional measurements provided by the ship's operational instruments. The majority of the sensors were mounted on a scaffold tower just aft of the bow (see Figure 8); the turbulence sensors were mounted on a forward-facing boom on the ship's jackstaff at the most forward and best exposed location on the ship. The ETL measurement system is described in detail by Fairall et al. (1997), so only a brief sketch will be given here.

The following ocean-atmosphere interface measurements were made:

(i) Turbulent measurements of stress and buoyancy flux using a sonic anemometer/thermometer.

(ii) Latent heat flux using a fast-response infrared hygrometer was used with the sonic velocity data. Fluxes were computed using covariance, inertial-dissipation, and bulk techniques (Fairall et al. 1996a). Ship motions were corrected by the inertial navigation system (Edson et al. 1997).

(iii) SST derived from bulk water measurements at a depth of 5 cm with a floating thermistor; corrections were applied for the cool skin effect (Fairall et al., 1996b). An IR radiometer system (provided by CSIRO) gave an estimate of true ocean skin temperature to complement the floating sensor (5 cm depth), the ship's thermosalinograph (about 5 m depth), and CTD ocean temperature profiles.

(iv) Mean air temperature and humidity derived from a conventional aspirated temperature/relative humidity sensor. An infrared hygrometer provided redundant information. CSIRO provided a pair of aspirated thermocouple wet/dry bulb psychrometers. The CSIRO psychrometers and rain gauges served as standard for mean air temperature and humidity and for rain rate. CSIRO also provided two STI Mini-ORG rain gauges that have been calibrated in the CSIRO rain tower.

(v) Wind velocity derived from the sonic anemometer after appropriate corrections for ship motion; the winds being referenced to the sea surface using the ship's Doppler speed log or to fixed earth using the ship's pcode GPS. Error estimates are given in Fairall et al. (1997).

(vi) Surface skin temperature using a pair of narrow band IR radiometers facing seaward and skyward (Donlon et al. 1998). The system was provided by CSIRO and an identical set was used on the *R/V Franklin*.

(vii) Atmospheric CO_2 concentration measurements for investigations of air-sea transfer of CO_2. The NOAA/AOML operational bulk CO_2 measurement system was used to provide air and oceanic concentrations of CO_2 and the air-ocean concentration difference (pCO_2). Both open-path and closed-path IR-absorption fast CO_2 concentration instruments were used (Fairall et al. 2000); these sensors provided fast water vapor measurements to complement the standard fast IR hygrometer.

The ETL flux system has been deployed on a number of shipborne experiments such as the Tropical Instability Wave Experiment (TIWE) during November and December, 1991 (Chertock et al. 1993), TOGA COARE, February, 1990, to February 1993 (pilot cruise and three TOGA COARE legs: Young et al. 1995, Fairall et al. 1996a), the Combined Sensor Experiment (CSP), March-April 1996) (Post et al. 1997), among others. In addition, subsets of the system have been deployed on ships during many other campaigns. A complete description of the observing system and the accuracy of individual instruments can be found in Fairall et al (1996b).

The location of instruments on the Australian *R/V Franklin* was generally as illustrated in Godfrey et al. (1999, see their Figure 5), although no Seasoar or buoy were deployed in JASMINE. The ship was equipped with a boom which extended 10 m forward of the bow to carry certain instruments clear of influence of the ship. Others instruments were mounted on an arm near the top of the foremast and some were located at various positions along the yardarm on the main mast. The following measurements were made:

(i) Longwave and shortwave radiation components with a pyrgeometer and pyranometer on the main mast. A broadband net radiometer was mounted at the end of the boom. There are thus two independent estimates of total net radiation for comparison as described by Godfrey et al. (1999).

(ii) Air temperature and humidity using aspirated psychrometers at 6 m above the sea surface on the boom, and at 13 m on the foremast.

(iii) Wind from cup anemometer/wind vane systems located at 13 m on the port side of the foremast and 17 m on the starboard end of the yardarm. Wind speed and direction were calculated from whichever system had the better exposure, and reduced to the 10 m standard height. Correction from relative to true wind was done using the ship's log/gyro system, winds being thus referenced to the surface current.

(iv) SST with a thermistor (Sea-snake) towed from a side boom, outside the ship's bow-wave at approximately 5cm depth, and by the ship's thermosalinograph taking water from an effective depth of about 1 m.

(v) Two optical rain gauges were mounted on the foremast, with optical paths orthogonal to one another, and a siphon rain-gauge was located on the yardarm. Bulk fluxes of momentum and sensible and latent heat were calculated from the above basic variables using version 2.6 of the COARE bulk algorithm (Fairall et al. 1996a).

2) UPPER OCEAN STRUCTURE

A Sea-Bird CTD (conductivity, temperature and depth) instrument was used with duplicate temperature and conductivity sensors. The duplicate temperature and conductivity sensors were calibrated before and after the cruise. Measurements were made to better than 0.01C in temperature and 0.01 psu for salinity below 5 m. In addition, the *Ronald H. Brown* came equipped with a thermosalinograph which provided a continuous, high-resolution depiction of temperature and salinity of the surface layer. Horizontal currents over the depth range of 20--300 m were measured from the shipboard acoustic Doppler current profiler (ADCP) which has a vertical resolution of 8 m.

The *R/V Franklin* oceanographic measurement system mirrored the *Ronald H. Brown*, except that (i) the CTD used was and Neil Brown Mark III; (ii) nitrate and phosphate samples were also taken on each cast during the transect from 5˚S to 12˚N, and salinity was checked from at least 3 bottle samples on each station.

3) RADARS

One of the unique characteristics of the *Ronald H. Brown* is the permanently installed, stabilized scanning C-band Doppler radar. The range of the radar is about 240 km. Although, the effective range for Doppler determinations is approximately 100 km. The C-band has a 5 cm wavelength and a 1˚ beam width. Three scan sets were used during JASMINE. The surveillance scan set consisted of 2 low elevation tilts and took about 2 minutes to complete. The two volume scan sets consisted of 21 elevation tilts, slightly staggered to provide better vertical resolution of the precipitating region. Each of these sets took about 8 minutes to complete. The radar alternated between the surveillance and volume scan sets, totaling about 20 minutes for a complete cycle. Volume scans occupied the space between 0.4 and 0.8˚ above the horizon, and

80˚ above the horizon. The latter tilt was undertaken to provide a comparison with the up-looking cloud radar described below[6]. The Single-Doppler ship-based radar reflectivity observations documented the basic three-dimensional structure of the precipitation patterns and the single-Doppler radial velocity observations documented the basics of the convective and mesoscale circulations. From the single-Doppler data it is possible to identify the layers of low-level inflow of warm moist air and the slope and direction of the upward circulation of this air through the storms and so identify the layers of mid-level inflow air and their general paths through the mesoscale systems (Kingsmill and Houze, 1998a, b). Subsequent analysis of this basic documentation of the circulation pattern, will enable the determination of the key levels of the atmosphere where the convective systems are interacting with the large-scale circulation.

A number of profiling Doppler radars and radiometers were used on the *Ronald H. Brown* to define a variety of planetary boundary layers and to provide diagnostics of precipitating systems. The radars utilized on the *Ronald H. Brown* were:

(i) Seagoing versions of the 915 MHz wind profiler (Ecklund et al. 1997) have been used at sea since their development for the TOGA COARE program. The system uses two tilted (15˚) and one vertical beam from a switchable phased array antenna on a mechanical stabilizer to maintain a level orientation. For JASMINE, wind profiles were obtained at 60 m and 100 m resolution; with reliable winds to about 2 km and 5 km, respectively, in clear-air conditions. Reliable winds were usually obtained in precipitation up to about 15 km.

(ii) A 3 GHz (S-band) profiling Doppler radar was used for precipitation measurements (Ecklund et al. 1999). This system uses one vertical beam from a fixed parabolic dish antenna. No stabilization was used. Compared to the 915 MHz wind profiler, the S-band system is much less sensitive to clear-air returns but more sensitive to precipitation. Reliable returns were usually obtained in precipitation up to about 15 km. These two radars have seen extensive use in diagnosing properties of convective systems (Gage et al. 1996; Gage et al. 1999).

(iii) A 35 GHz (K-band) profiling Doppler cloud radar (Moran et al. 1998) was deployed in combination with a dual frequency microwave radiometer (Hogg et al. 1983) for cloud property measurements. The short wavelength (8 mm) make this radar sufficiently sensitive to observe cloud droplets (thus, it is often referred to as a “cloud radar”). Cloud radars have had only limited application to marine clouds (Frisch et al. 1995; White et al.

[6] Details of the volume scan sets given at http://www.atmos.washington.edu/~serra/JASMINE/jasmine_overview.html

1996; Post et al. 1997) and JASMINE is only the second deployment of such a system on a ship.

(iv) The microwave radiometer provided retrievals of total integrated water (vapor and liquid separately) and has been used extensively for marine investigations (Snider and Hazen 1998). The microwave radiometer is an essential complement to the cloud radar for retrievals of cloud droplet size information (Frisch et al. 1995). Additional cloud information was obtained from a commercial lidar ceilometer, which is not 'blinded' by precipitation so it is more capable of detecting cloud base than the cloud radar. The ceilometer possessed a vertical resolution of 35 m and a maximum range of 7.8 km.

No radar system was available on the *R/V Franklin*, except the ship's navigational radar which was not recorded.

4) ATMOSPHERIC SOUNDINGS

The single-Doppler radar observations are useful in analyzing the convective-large-scale interaction occurring in active and break monsoon circulations only to the extent that the ambient large-scale thermodynamic and wind stratification in the vicinity of the mesoscale convective systems is monitored by an extensive time series of soundings in the region of the single-Doppler radar observations. On the *Ronald H. Brown* these soundings were taken at 4--hour intervals during the transects and 3--hour intervals while on station. The GPS tracking system provided great accuracy in wind measurements. In conjunction with the soundings, high-frequency surface meteorology measurements were maintained aboard ship. A similar system was used on the *Franklin* where soundings at 6-hour intervals were during the transect and the subsequent budget study.

5) ANCILLARY DATA

In addition to the shipboard instrumentation, satellite data was available to the investigators in real time. High-resolution satellite data from the orbiting satellites NOAA--12, NOAA--14, NOAA--15 was accessed in high resolution format aboard the *Ronald H. Brown*. In addition, the European geostationary satellite METEOSAT-5 was repositioned over the Indian Ocean for INDOEX and remained operational throughout JASMINE. Three-hourly data for the JASMINE period was made available to JASMINE investigators by the European Space Agency after the experiment. In addition, the European Centre for Medium-Range Weather Forecasts (ECMWF) has made 3--hourly analyses at standard levels and single point data available to the investigators.

d. Experimental design

Initially, JASMINE was planned to be implemented in the eastern Indian Ocean in the period May 12 to June 20 aboard the NOAA Ship *Ronald H. Brown* and in September aboard the Australian *R/V Franklin*. The period was chosen to maximize the chances of encountering a full cycle of an MISO in the eastern Indian Ocean (see Figures 6 and 7). In the final instance, JASMINE acquired an additional two weeks of ship time following INDOEX as the ship relocated from the Maldives to Singapore. The final ship schedule for JASMINE is shown in Table 2.

The experimental design called for a series of transects along 89˚E. The choice of 89˚E ensured that we were in international waters at the northern end of the transect by avoiding India's national waters defined in the east by the Andaman Islands less 200 km. between 5˚S and 10˚N with two stops near (10˚N, 89˚E) for periods of 5 days. At these end points a star pattern was undertaken (see Figure 10) with continual circuits for five day periods. The star pattern was chosen principally for oceanographic reasons. The size of the star needed to be large enough to sample horizontal gradients associated with dominant advective processes yet small enough so that the pattern can be completed in a time short compared to important periods of variability. The circuit time around the star took 1 day which was sufficiently brief compared to the inertial period at 10˚N. The Australian Research Ship *Franklin* chose an on-station triangle pattern (see Figure 10) rather than a star with each triangle taking 8 hours.

The aim of the operations plan was to sample changes in the ocean-atmosphere system through the full life-cycle of an MISO. Figure 9 shows a latitude-time section along 89˚E relative to the composite MISO configured from the data shown in Figure 6. The schematic plan was to be adjusted in the field relative to the phase of the MISO encountered using satellite information aboard the NOAA Ship *Ronald H. Brown*. However, no substantial changes to the plan were necessary as the weather encountered matched the composite weather remarkably well (cf Figures 9 and 11).

The following measurements were taken along each of the transects:

(i) Along each leg CTD measurements were made every 1/3 degrees of latitude. This sampling strategy provided a detailed temperature and salinity structure of the oceanic boundary layer on a resolution sufficient to capture the variability due to the monsoonal forcing.

(ii) Because covariance turbulent flux measurements require correction for ship motion and flow distortion (Edson, et al, 1998), at each CTD station the ship was positioned to optimize the eddy correlation flux estimates. In addition, accurate mean meteorological and SST measurements were gathered from which the bulk algorithm were implemented and compared with the accurate direct covariance flux measurements. Long- and shortwave radiative fluxes were monitored continuously throughout the experiment.

(iii) In transit, bulk estimates of fluxes were made together with continual Doppler radar measurements at 3-GHz and 915-MHz. The continuous radar sampling along the 89E transects resolved the structure of the mesoscale circulations during active and break periods.

(iv) On the *Ronald H. Brown* radiosonde measurements were made 6 times per day and 4 times per day on the *R/V Franklin*. The timing of the soundings matched international agreed release times for upper atmosphere soundings.

During the on-station "star" observation periods:

(i) The number of CTD measurements were increased to give a spacing of xx km at a depth of 300 m which was increased to 1000 m at the apexes of the stars.

(ii) The number of upper air soundings was increased from 6 per day to 8 per day.

(iii) All other measurements continued on the same schedule as during the transects. Six CTD casts were made from the *R/V Franklin* during each triangle. Experience in an earlier cruise in the equatorial East Indian Ocean (Godfrey et al., 1999) had shown this simple design to be effective in upper ocean budget closure experiments.

e. Summary of the data collected

Table 3 shows the total number of CTD casts and radiosonde ascents made in the three phases of JASMINE. In all, there were 388 CTDs and 22 XBTs made during JASMINE. In addition, there were 272 radiosonde ascents. Over 95% of the atmospheric soundings reached the Global Telecommunications System (GTS) and were used as initial data by major numerical centers throughout the world.

(4) Initial Results from the JASMINE Field Phase

We now present a broad overview of the data collected during the field phase of JASMINE, together with some of the results early analyses.

a. Synoptics

To describe the overall meteorological situation during the JASMINE period, and the MISO encountered during the field phase, the brightness temperature (K) averaged in the band 85˚--95˚E is shown as a function of time and latitude using data from the European METEOSAT-5 geostationary satellite in Figures 11a-e. The ship tracks of all three phases are shown in longitude-time space as solid lines. Cold temperatures are indicative of high cloud tops while relatively clear periods appear as warm temperatures representing infrared radiation emitted at the surface or from low clouds near the surface.

During Phase I in April (panel a), a relatively weak disturbance persisted in the Bay of Bengal with the ship passing through the disturbance near Julian day (JD) 103. The first half of Phase II (panel b) was a period of warm emittance indicative of little cloudiness and high insolation. However, during the second half of Phase II the atmosphere was greatly disturbed with low insolation and considerable precipitation. The disturbance encountered during the second half of Phase II was associated with the northward propagation of an MISO. Again, it is worth noting the similarity of the composite oscillation (Figure 9) and the MISO encountered during the JASMINE field phase (Figure 11). It is clear from Figure 11 that the range of weather that occurred during Phase II fulfilled the requirements set out in the planning. Phase III (panel e) was essentially a quiescent period during the late summer monsoon.

One of the striking features of the brightness temperature distribution in the Bay of Bengal is the extremely strong diurnal variability occurring during the disturbed period of Phase II. This variability can be identified as southward propagating deep convection with nocturnal maxima (brightness temperature < 200 K). Figures 11c and d, show brightness temperature sections for June and July and indicate that the southward propagating diurnal disturbances are present during each disturbed phase of an MISO in the Bay of Bengal.

Figure 12 plots the variation of wind in the upper and lower troposphere during Phase II of JASMINE. Winds are plotted as a function of latitude and show horizontal wind vectors along 89˚E using ECMWF data. Wind speeds are color coded. During the first northward transect there were moderate winds which lessened to values < 6 m s^{-1} during Star 1. Moderate upper tropospheric easterlies changed to almost calm conditions during the same period. However, during the second northerly transect, the winds steadily increased in strength in both the upper and lower troposphere and reaching gale force during Star 2. Strong winds persisted for over 10 days as the intraseasonal event moved northward through the Bay of Bengal. As the convection increased in the northern Bay of Bengal the upper tropospheric winds exceeded 20 m s^{-1} during

Star 2 intensifying to 30 m s^{-1} during the last southward transect. The acceleration of the upper tropospheric easterlies heralded the arrival of the monsoonal easterly jet stream for the 1999 summer monsoon. Note, though, that there appears to be an intraseasonal component to the strength of the easterly jet stream. This is not surprising given that convection essentially drives the easterly jet (e.g., Webster 1972, Gill 1980) and there is a strong intraseasonal component to convection. Overall, the wind patterns during the latter half of Phase II is quite similar to the climatology (although stronger) displayed in Figure 3.

Figures 13 a, b show the synoptic situations during the two stars of Phase II. Both 24--hour average mean sea level pressure (from ECMWF) and brightness temperature fields (from METEOSAT) are plotted. The white dot denotes the location of the *Ronald H. Brown*. During the period 12--15 May 1999 (Star 1), the surface pressure gradients were slack prior to the onset of the southwest monsoon and the Bay of Bengal was virtually convection free. During the second period (22--25 May: Star 2), the southwest monsoon had commenced with the northward movement of the MISO (see Figure 11). The surface pressure gradient became very tight and strong cyclonic surface wind shear existed through the whole Bay of Bengal. The flow was essentially southwesterly and possessed a strong ageostrophic component towards the low pressure trough located in the northern Bay. Strong convection covered most of the Bay with 24--hour brightness temperatures < 220 K near the ship.

b. Structure of the upper ocean

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Need lots of help here to describe (a) the transects (T, S and u), and (b) the T and S variations during the two Star patterns. I.e., Figures 14 and 15.

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A major goal of JASMINE is to determine the variability in the upper ocean that occurred during the intraseasonal transitions of the monsoon. The concentrated upper ocean observations made during JASMINE allow such a documentation. As will be shown below, the ocean structure shows marked intraseasonal variability.

Figure 14 shows latitude-depth sections of the temperature, salinity and zonal current structure of the upper ocean along the north-south transects during Phase I and II of JASMINE. The sections are arranged in the following manner. In each figure, five sections are shown, each representing a north-south transect of the *Ronald H. Brown* as described in Table 2. Arrows indicate the direction of the transects. Star patterns indicate the time of the on-station periods at the northern end of the transects. The temperature and salinity structures during the stars will be discussed subsequently, relative to Figure 15.

The overall temperature structure of the Bay of Bengal during the JASMINE period shows a well-developed mixed layer from the equator to the north. Warmest temperatures occur generally in the northern regions. Over the two-month period, there is considerable variation of the depth of the thermocline. In the four weeks between Phase I and II, there are large changes in the depth of the thermocline. For example, in the vicinity of 4˚N the thermocline deepens from about 100 m to 150 m. This thickening appears to move northwards during the remainder of Phase I finishing during the last transect at about 6˚N. The salinity sections (Figure 14b) show that the freshest upper ocean water is constrained to the north of the transects. The salinity features also show variability during the Phase I and II of JASMINE. For example, in the latitude band between 3˚N--6˚N the water changes from being relatively fresh during Phase I, increasing salinity during the first three transects of Phase II before freshening considerably during the last transect. Ideally, the temperature and salinity should be considered concurrently. In fact, collectively the two distributions suggest a well-developed barrier layer in the Bay of Bengal. Details of the barrier layer structure and its transitions during JASMINE are given by Hacker et al. (2000).

The zonal velocity field for Phase I and II (Figure 14c) also show considerable variability. During Phase I, the near-equatorial upper ocean zonal currents are strongly to the east with speeds or order 1 m s^{-1}. Four weeks later the currents have reversed sign and are about the same strength to the west. By the time of the last transect, the currents have reversed once again. The oscillation of the zonal currents is also discussed by Hacker and Lukas (2000). The oscillation is most certainly the result of the intraseasonal oscillation of the lower tropospheric wind field. Han and Webster (2000) have managed to reproduce many aspects of the near-equatorial current field observed in JASMINE using ECMWF wind forcing. The model results simulate the observed current structures rather well.

c. Atmospheric Structure

Figure 16 provides a more detailed view of the convective and dynamic structure of the atmosphere compared to the "broad-brush" description shown in Figures 11 and 12. The figure shows two sets of height-time sections of the horizontal wind and humidity, and radar reflectivity as measured by the C-band radar. Sections are shown for both Star 1 and 2. Two curves are shown in these sections representing the percent area within 100 km of the *Ronald H. Brown* with reflectivity in the range 15 dbz to 35 dbz (blue curve) and > 35 dbz (red curve). Deep convection is an order of magnitude more abundant during the night than during the day during the disturbed period.

During Star 1 the winds were very light throughout the column. At this time, the middle and upper troposphere were very dry with relative humidities less than 20--25%. Moist air was confined to the boundary layer where light southwesterlies persisted during the entire period. There is very little convection as suggested by the very weak C-band radar echoes although persistent thin cirrus cloud existed during the entire period. The cirrus, although only barely discernible for the ship, occasionally was 2--5 km thick. During Star 2 the winds had freshened with strong north-westerlies up to 400 mb and very strong north-easterlies in the upper levels producing strong vertical shear throughout the troposphere. Individual winds in the upper troposphere exceed 30 m s^{-1}. Lower tropospheric winds exceed 20 m s^{-1} for considerable periods. Compared to the Star 1 structure, the entire troposphere had moistened remarkably during the intervening week probably because of the intense convection that is evident in both the cloud radar and C-band reflectivity plots. Very strong diurnal variability in the radar reflectively indicating a distinct nocturnal maximum in convection.

d. Surface flux variability

One of the major scientific objectives of JASMINE was to measure the interaction of the ocean and atmosphere through a detailed documentation of the surface fluxes. Surface fluxes were measured for the duration of the Phase II and III cruises using equipment and techniques described in detail in section 3. In the following paragraphs we discuss the variability of radiative and turbulent fluxes between undisturbed and disturbed periods of the monsoon. Also, estimates of the total heat flux into the upper ocean during Phases II and III of JASMINE will be discussed.

Figure 17 shows the components of the surface energy balance for Star 1 and 2. Two very different distributions are apparent. For the relatively undisturbed period during Star 1, which corresponds to the period before the onset of the southwest monsoon, or a break period in the monsoon, the net heat flux into the ocean is +92 W m^{-2}. During the active monsoon encountered during JASMINE Phase II (Star 2), the net flux reversed and showed a loss of 89 W m^{-2}. A summary of the fluxes and a comparison with data collected in the Pacific Ocean is given in Table 4.

During Star 1, the 24--hour solar fluxes averaged +260 W m $^{-2}$ compared to and average of +162 W m $^{-2}$ obtained the disturbed Star 2 period. Peak solar irradiance was in excess of +900 W m $^{-2}$ each day during Star 1 compared to a depleted maximum in the range of +500 to +800 W m $^{-2}$ during the disturbed period. The net longwave radiation at the surface was --49 W m $^{-2}$ during Star 1 but decreased by nearly 20 W m $^{-2}$ on average during Star 2. The reason for this decrease in longwave loss was due to the increase of downwelling radiation associated with the extensive cloud systems and the increase of moisture through the entire troposphere (Figure 16b). The major ocean surface cooling agent in all phases of JASMINE was due to the latent heat flux. During Star 1 the latent heat flux was --115 W m $^{-2}$ compared with --162 W m $^{-2}$ during Star 2. The increase of this flux corresponds to the much stronger winds during the disturbed period. The variability of the latent heat flux was also very large ranging between --50 W m $^{-2}$ to --300 W m $^{-2}$. The larger values were associated with the nocturnal gust fronts moving through the Bay of Bengal.

The change from a net heating of the upper ocean to a net cooling (i.e., +92 W m $^{-2}$ to --89 W m $^{-2}$) is a combined result of decreased solar radiation and increased evaporation, a cooling slightly offset by the decrease in longwave cooling, described above. The sensible heat loss by the ocean due to turbulent transfer and rain cooling was not important during the undisturbed period having values of --5 and essentially zero, respectively, during Star 1. However, these flux components become moderately important averaged over the 5 days of Star 2 with mean values of --17 and --7 W m $^{-2}$, respectively. Again, the increase in turbulent sensible heat flux is determined by the strength of the wind. The sensible heat of rainfall (see Gosnell et al. 1995) was greater on average during the disturbed period than the turbulent sensible heat during the undisturbed period. In fact, instantaneous values of the sensible heat loss due to rain cooling of the surface was occasionally greater than --200 W m-2.

Figure 18 shows the daily averaged net heat flux into the ocean for Phases II and III of JASMINE. The two star periods are shaded. The mean net heat flux into the ocean for Phase II is +27 W m $^{-2}$. A very strong MISO is apparent in the plot with values ranging from +125 W m $^{-2}$ to --150 W m $^{-2}$. During Phase III of JASMINE, the mean fluxes were generally strong and positive with a mean value of +96 W m $^{-2}$. Much of Phase III appears to have been in a break phase of the monsoon or after the final retreat of the monsoon. Figure 11e suggests that there may have been one more active phase of the monsoon although the location of the disturbance is located further equatorward than earlier active events.

The strong intraseasonal variability of the fluxes is generally larger than that encountered during TOGA COARE in the western Pacific Ocean. However, the amplitude of the flux variability suggests that the period of observations needed to calculate the heat balance of the Indian Ocean warm pool must encompass multiple MISOs or active/break periods. Shorter periods of observations may be biased by over sampling either active or break phases of the monsoon. There are periods even in the western Pacific Ocean where the net heating of the ocean is as large as observed in Star 1. For example, early in the TOGA COARE field phase the middle and troposphere were uncharacteristically dry following subductions of higher latitude air (Parsons et al. 2000). They suggest that the exaggerated atmospheric stability at the time precluded the formation of clouds and the 24--hour net solar radiation reaching the surface was +222 W m^{-2} or higher by nearly 40 W m^{-2} than the average for the remainder of TOGA COARE. These values are listed as "TC-period 1" in Table 4. A more detailed comparison of the Pacific Ocean fluxes with those measured during JASMINE will be given in section 6.

e. Diurnal variability

One of the characteristics of the tropical atmosphere is the strong diurnal variability in convection, SST and surface fluxes as were noted in the western Pacific Ocean TOGA COARE period (Godfrey et al. 1998). The largest diurnal variability is normally thought to occur mainly during undisturbed periods when insolation is strong and the winds are light. In JASMINE we had the opportunity to study the diurnal variability during both undisturbed and disturbed conditions. Comparing the two Star periods shows that, contrary to expectation, a surprising amount of variability exists during disturbed periods.

1) ENVIRONMENT:

Figure 19 shows the mean diurnal variability of the atmosphere and the ocean-atmosphere interface, plotted as a function of local time of day for Star 1 (dashed curves) and Star 2 (solid curves). Figure 19a shows time sections of the SST and the 10 m near surface air temperature. During Star 1, the SST shows a diurnal variability with an SST and air temperature maximum in late afternoon. However, during this period in the Bay of Bengal, the amplitude was less than that observed in similar periods in the western Pacific Ocean during TOGA COARE (e.g., Webster et al. 1996). The air temperature tends to follow the SST and possesses a nocturnal minimum. During the disturbed Star 2 period, the SST shows little or no diurnal variability. However, the air temperature showed an extremely large variation with an amplitude of >2˚C. The nocturnal cooling of the boundary layer was caused by the convective downbursts associated with the nocturnal squall lines that propagated southward each night during Star 2.

Figure 19b shows the mean diurnal variability of the surface wind speed. In the undisturbed case the winds are generally light with a tendency for a nocturnal maximum. In the disturbed case there are universally stronger surface winds with larger variability at night (not shown) but associated with the reoccurring surges that were apparent in the latent heat fluxes in Figure 17b. However, most evident is the very strong nighttime maximum in rainfall rate. Again, the nocturnal maximum is associated with the propagating disturbances.

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Comparison with the diurnal cycle of precipitation in the WPO and other monsoon regions: Bob Houze and Yolande Serra?

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Figure 20, mean time section diurnal variability of T(z) as a function of time for Star 1 and 2 as a function of local time.

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2) SURFACE FLUXES:

The mean diurnal variability of the turbulent fluxes and the net surface longwave radiation, as a function of local time are plotted in Figure 21a for Star 1 and 2. The latent heating distributions largely follow the mean diurnal variation of the winds (Fig. 19b) as does the variation of the sensible heat. However, the latter flux is exaggerated in Star 2 by the strong decrease in 10 m air temperature during the night. Figure 21b shows the variability of the radiative fluxes. The solar radiation is severely reduced during the disturbed period with an average peak nearly +300 W m^{-2} less than in Star 1. The downwelling infrared is about 30 W m^{-2} higher in Star 2 compared to Star 1. The surface radiative forcing for both shortwave and longwave radiation is shown in Figure 21c. Very strong negative surface shortwave forcing is apparent in the Star 2 period during the day time because of the generally overcast conditions that attended the disturbance (Figure 11). The solar radiative forcing is much smaller in the Star 1 period. These small negative values occur because of persistent high cirrus cloud.

3) LARGE-SCALE CONVECTION:

Figure 11 showed sections of satellite derived brightness temperature as a function of latitude along 90E for most of the summer monsoon of 1999. It was pointed out that very strong diurnal variability existed during the disturbed period of Star 2 with extremely strong convection appearing to move from north to south through the Bay of Bengal. This variability appears as a common feature of disturbed periods in the monsoon and is probably an integral part of the MISO in the Bay of Bengal. The ubiquity of the nocturnal intense convection and the rapid southward advection, can be seen in the latitude-time plots of brightness temperature for June and July as well. In June and July, the diurnal variability of convection is also marked as southward propagating disturbances which form in the north of the Bay of Bengal. The August brightness section is not shown because of satellite problems.

The southward propagation speed of the nocturnal disturbances is universally about 60 km $hour^{-1}$ or 15--20 m s^{-1} which can be seen from the 60 km $hour^{-1}$ phase lines drawn Figure 11c. From the flux sections shown in Figure 17 it appears that these propagating disturbances severely perturbed the surface fluxes as they propagated southward past the ship. Noting that the direction of the lower tropospheric wind is southwesterly and the upper tropospheric winds are northeasterly (Figs. 12 and 16), these diurnal events propagate orthogonally to the winds at all levels. Furthermore, there is evidence that the propagation continues into the southern hemisphere as far south as 10˚S. Along a track, the amplitude of the convection appears to decrease through the next local daytime but growing once again further to the south during the following night. In summary, during each of the intraseasonal events, propagations occur along trajectories that may be 2--3000 km in length and appear to remain as discrete entities for about two days.

The source region and the horizontal scale of the propagating disturbances have been determined from the study of the three hourly satellite brightness temperature distributions (not shown). The disturbances appear to form in the early afternoon in the northwestern reaches of the Bay of Bengal in the vicinity of the surface trough running along the western Indian coast from Bangladesh to the south. The disturbances adopt a southwest-northeast orientation roughly parallel to the surface isobars (Figure 13b) and extend in that direction for scales of about 300--500 km. The propagating lines are marked by deep convection that extends through the troposphere. In the direction of propagation, the disturbances are rather narrow (>50 km).

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A search of the literature and prominent texts on the monsoon (e.g., Ramage, 1971, Rao 1976, Das 1986) failed to find any reference to these events. Basically, the cause of the persistent nocturnal disturbances is not known. Nor is it understood why they propagate orthogonally to the wind field or persist as definable entities over such great distances. Given the time and location of their formation, it is highly probable that they are related in some manner to land-ocean processes that occur in the afternoon at the head of the Bay of Bengal. Whether these squall lines have similar formation mechanisms and maintenance controls as the West African squall lines (e.g., Payne and McGarry 1977, McGarry and Reid 1978, Dudhia 1987), the Amazonian squall lines (e.g., Garstang et al. 1994, Greco et al. 1994) or the systems documented off Kalimantan during Winter MONEX (e.g., Houze 1981), is not known.

d. Clouds and Convection

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Bob Houze and Yolande Serra help?

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Figure 22 shows snapshots of the convection on the 22 May, 1999, during the Star 2 period. Figure 22a shows the cloud radar image as a function of time and height. The propagating squall line is most evident with very intense echoes occurring after 1500 UTC (2100 local time). Figure 21b shows the C-band reflectivity for 1704 UTC on the same day. The image shows the part of the propagating disturbance closest to the ship (center of figure). A cross-section of reflectivity at 1708 UTC in a northwesterly direction is shown in Figure 22c. The position of the ship is at the left hand axis. The panel provides an indication of the depth of convection associated with the nocturnal storms.

5. Data

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Should we have a small section that talks about the availability of data? I.e., a start of a data catalogue? Should we include this in an appendix or should we not bother at all?

6. Some Tentative Conclusions and Future Plans

During the summer of 1999, oceanic, atmospheric and ocean-atmosphere interface data was collected in JASMINE for a period of 52 days in the eastern Indian Ocean and the southern Bay of Bengal. Initial results, discussed in detail in section 4, show a clear distinctions between active and break periods of the southwest monsoon. A summary of initial results is given below:

(i) The atmospheric structure throughout the entire troposphere changed significantly through the life cycle of the MISO encountered in the field phase. In the quiescent phases of the MISO, the troposphere was remarkable dry above the boundary layer with weak upper level easterly winds surmounting weak surface southwesterlies. During the disturbed phase, the entire troposphere becomes very moist and the weak wind structure is replaced by strong lower level southwesterlies and upper level northeasterlies. The dry middle and upper troposphere is similar to the early period in TOGA COARE where there was a strong incursion of higher latitude dry midtropospheric air over the western Pacific (Parsons et al. 2000). Whereas the Pacific Ocean remained moist throughout the troposphere during subsequent intraseasonal transitions in TOGA COARE, suggesting that the early period of COARE was anomalous, there is some evidence (e.g., Figure 5) that the drying of the troposphere during break periods throughout the summer monsoon may be common. If this is the case, the pre-onset environment encountered during the first half of Phase II of JASMINE may be indicative of break periods of the monsoon later in the season.

(ii) The surface heat flux at the surface changed dramatically between the pre-onset and the active phase of the monsoon. Mean surface fluxes between Star 1 (undisturbed) and Star 2 (disturbed) changed from +92 W m^{-2} to --89 W m^{-2}. The differences were caused by a severe reduction in net surface solar radiation, and increases in the turbulent fluxes, offset slightly by a decrease in net longwave radiation. The changes in turbulent fluxes can be accounted for, to a large degree, by the dramatic increases in surface winds and the decrease of surface radiation accompanying the disturbance. The representativeness of these fluxes for subsequent active and break periods needs to be assessed. If the surface heating were as low in break periods as in the undisturbed periods in the western Pacific (see Table 4) then SST variability in the eastern Indian Ocean and the Bay of Bengal would be different from that observed. This conclusion is supported by Indian moored buoy data in the Bay of Bengal (J. Sangupta, Indian Institute of Science, Bangalore, India, personal communication). The buoys were deployed during the summers of 1998 and 1999 at (13˚N, 87˚E), (16˚N, 82˚E) and (18˚N, 88˚E). Time sections of SST showed oscillations between about 29˚C during the disturbed period of the MISO to 30.5˚C in the undisturbed periods. The period of the oscillations is between 20--40 days.

(iii) The Indian Ocean responds to the intraseasonal forcing of the monsoon winds with reversals of the equatorial currents in the eastern Indian Ocean on the same time scales (Figs. 14). These oscillations are discussed at length by Hacker and Lukas (2000) and shown to be wind forced by numerical experiments (Han and Webster 2000). Whether or not these intraseasonal changes in ocean currents are responsible for the intraseasonal oscillations in cross-equatorial heat transports found in models (Figure 8) cannot be determined from the JASMINE data.

(iv) A barrier layer structure exists in the eastern Indian Ocean and the Bay of Bengal. As in the Pacific Ocean, the barrier layer is modulated by intraseasonal atmospheric forcing (Hacker and Lukas 2000).

(v) Strong diurnal variability is evident in the eastern Indian Ocean in both the undisturbed and disturbed periods. The diurnal variation during the undisturbed periods is similar to that found in the western Pacific Ocean with an afternoon maximum of the upper layer temperature. There was a strong diurnal variation of convection during the active periods especially in the timing of maximum precipitation and convection. The nocturnal convection appeared to have a form that was unique to the region with strong propagating convective bands that moved southward from the head of the Bay of Bengal in directions orthogonal to the wind direction at all levels. There is some evidence that these features remain as coherent entities into the southern hemisphere persisting through the daytime and regrowing within the southern hemisphere part of the MISO during the following night.

(vi) The flux data collected during JASMINE allows an initial comparison of net flux and variability between the Indian and Pacific oceans. Table 4 lists data from a number of western Pacific Ocean experiments and from JASMINE. The average fluxes for JASMINE Phase II are very similar to the average fluxes for the TOGA COARE period and the recent Nauru-99 experiment. However, this similarity may be deceptive and covers up the large intraseasonal variability in the Indian Ocean. The extent of the intraseasonal variability during TOGA COARE can be seen by comparing the three periods of the experiment. The first period (TC-period 1) was a relative quiescent period with a net flux of +65 W m . The second period (TC-period 2) contained one of the largest westerly wind bursts recorded in the western Pacific Ocean. The net flux for the period was --12 W m . The next period possessed above average winds and showed a net flux of +13 W m . While the Pacific Ocean fluxes vary between +65 and --12 W m , the Indian Ocean fluxes oscillated between 90 W m throughout the lifetime of an MISO. The major differences between the disturbed periods of the Pacific Ocean and the Indian Ocean appears to be strength of the winds and reduced insolation during disturbed periods of the monsoon. From this comparison, it is clear that the results of western Pacific warm pool experiments cannot be simply transferred to the Indian Ocean.

The results of the pilot study raise a number of questions that require investigation:

(i) What is the wider scale impact of the intraseasonal wind forcing on the Indian Ocean? We are aware that there is a strong local response both in the current field and in the thermodynamics of the upper ocean in the East Indian Ocean and the Bay of Bengal but what occurs in the wider field? What is the role of the ocean response in the longer term heat balance of the Indian Ocean and, consequently, on the variability of the monsoon itself?

(ii) Are the surface flux variations through the complete cycle of the MISO (i.e., undisturbed and disturbed periods) observed in JASMINE representative of the variability of the maritime monsoon climate? Data from the Indian surface moorings in the Bay of Bengal, discussed above, suggests that they may be representative otherwise it might not be possible to explain the 1.5˚C differential in SST between the break and active periods.

(iii) Can the results of the JASMINE pilot study be used by the modeling community to improve the ability of models to simulate MISOs and possibly precipitation distribution in a given monsoon season? What aspects of the monsoon are omitted from current models? Does the fault lie simply in model resolution? Is it necessary to include diurnal variability in SST, for example, or more keenly represent land surface processes to improve simulation? Are the surface flux, convective and radiative parameterizations used in models suitable for the monsoon climate or have they been developed with other radiative-convective-dynamic synergies in mind?

(iv) Will the increased knowledge of the structure of the MISO gained in JASMINE lead to medium term prediction of active and break periods of the monsoon? Do these results suggest that coupled ocean-atmosphere models are required for such forecasts?

(v) It has been presumed that the TOGA COARE period I studied by Parsons et al. (2000) was an anomalous situation. This presumption needs to be checked possibly by the examination of long-term precipitable water data from satellite because it resulted in a significant perturbation of the Pacific warm pool region. The structure encountered during TC-period 1 appears very similar to that found during the undisturbed period of the MISO in JASMINE (see Figure 16). If the two panels Figure 16 represents the variability of upper atmosphere between break and active periods, it is important that the source of the dry middle and upper tropospheric air be determined.

Because of recent studies on a new type of interannual variability in the Indian Ocean, we add another question:

(vi) What is the nature of the Indian Ocean dipole discussed by Yu and Rienecker (1999, 2000), Webster et al. (1999) and Saji et al. (1999). The Indian Ocean dipole is an east-west oscillation of SST between the eastern and western Indian Ocean with an amplitude of variability of 1--3˚C. The dipole appears to exist often in the absence of El Niño and sometimes in tandem and is strongly correlated with east African rainfall. Is the dipole oscillation a coupled ocean-atmosphere phenomena, as described by Webster et al. (1999), or is it a resonant response of the Indian Ocean to a variety of remote forcing? What is its role of the dipole in the interannual variability of the monsoon? How is it related to the biennial variability noted in the Indian Ocean SST and Indian rainfall?

Based upon what was found in the JASMINE Pilot Study and on the newly emerging ideas regarding phenomenology in the Indian Ocean-monsoon regime, the following research is planned:

(i) Conduct a more expansive survey of the coupled ocean-atmosphere interaction on intraseasonal time scales in the central and eastern Indian Ocean during the 2003--4 framework. The aim of the experiment is to obtain detailed flux, and evolving ocean and atmospheric structures for a longer period during the established monsoon circulation. Major emphases will be the determination of the wider field responses of the upper ocean to intraseasonal atmospheric forcing, and the gauging of the spatial and temporal patterns of interface fluxes. The timing of a JASMINE II is crucial and tied to the maximum coverage of satellites which will occur in the 2002--2005 framework. The process study is planned to take place at the same time as the GEWEX Coordinated Enhanced Observing Period (CEOPS).

(ii) Work with the international community to develop an Indian Ocean Monsoon Observing System for the study and monitoring of the Indian Ocean dipole. Such an effort requires the maintenance and expansion of XBT lines existing in the Indian Ocean and the deployment of moored buoys in the eastern and western Indian Ocean.

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Figure Captions:

Figure 1: Results of the Atmospheric Model Intercomparison Experiment (AMIP: Sperber and Palmer 1996) showing comparisons of the average boreal summer rainfall over India (left panel) and the boreal winter rainfall over the western Pacific warm pool (right panel) for the 9--year period 1979--1987. The models shown represent the seven "best" of the AMIP models. Each was forced with identical observed SST distributions. The observed precipitation is denoted by the dashed gray contour.

Figure 2: Comparison of convection in the Indian and Pacific Ocean basins in the form of OLR (W m-2) and SST (C) relationships for the tropical Pacific Ocean (panel a), the western tropical Indian Ocean (panel b) and the eastern tropical Indian Ocean (panel c). The locations of the three regions are shown in the maps below the figures. Note that the OLR scale is inverted with increased convection being associated with decreasing OLR.

Figure 3: The heat balance of the North Indian Ocean showing (a) the climatological mean annual cycle of the net surface heat flux into the Indian Ocean and heat storage north of the equator, and the northward heat flux across the equator, and (b) the interannual variation of the same three components for the years 1984 to 1990. Units PW (1015 W)

Figure 4: The mean near-surface circulation over the monsoon regions for June-August period using ECMWF data from 1985--1995. Upper panel shows the mean 925 mb wind field. The red dashed lines show the positions of the surface pressure troughs. Middle panel shows the 200 mb wind field. The long term average of summer precipitation (mm month-1) as determined from MSU precipitation and land surface data is shown in the bottom panel.

Figure 5a: Mean precipitable water content (PWC: mm) composited for active and break conditions over a 15--year period after Webster et al. (1998). The mean precipitation patterns shown in Figure 4 are made up of strong monsoon intraseasonal oscillations (MISOs) of 30--50 day time scales. These oscillations constitute "active" and "break" periods of the monsoon. Active periods of the monsoon are associated with substantial rainfall over Indian, the Bay of Bengal and South Asia and minima over the equatorial regions. Break periods are associated with minima in rainfall over the land areas but substantial rainfall in the equatorial regions and the foothills of the Himalayas. Data from Randall and Vonder Haar (Colorado State University).

Figure 5b: Intraseasonal variability of the monsoon depicted by the daily MSU precipitation along 90E plotted against latitude as a function of time for the boreal summers of 1985 and 1995. Beginning in early summer, precipitating events begin near the equator and extend polewards in each hemisphere. The northward extension of the equatorial precipitation becomes the active period of the monsoon while the southward extension produces precipitation to the south of the equator and, eventually, an enhancement of winter precipitation over southern Australia. Thirty-nine such events were identified in the years 1984 to 1995. These events were used to create composites of monsoon intraseasonal events. Day "0" is defined as the occurrence of precipitation occurring at 90E with intensity greater than 10 mm/day and persisting for four days.

Figure 6a: The composite evolution of the MSU precipitation (mm/day), shown as a function of latitude and longitude, relative to the composite MISO configured from 39 events in the period 1985-1995. Composite distributions were computed for 15 days on either side of day 0. Organized convection occurs first in the western Indian Ocean (day --8) , moves along the equator to the eastern Indian Ocean before bifurcating northward and southward from the equator (day 0). Note that the geographical area of the figure is larger than in subsequent Figures 6a and b.

Figure 6b: Same as Figure 6a except sea-surface temperature (SST). Prior to the formation of the intraseasonal event the SST is anomalously warm over the entire basin. After the formation of the MISO the basin SST cools. These variations are associated with a waning (prior to the event) and a waxing (after the event) of the entire cross-equatorial monsoon circulation (see Figure 6c). That is, the warm SSTs are associated with an anomalously weak monsoon gyre and the cooler SSTs with an anomalously strong gyre. Variations in the basin-wide SST range between -1C to +0.6C and are consistent with Fasullo and Webster (1999).

Figure 6c: Same as Figure 6a except for anomalous wind variability (m s-1). but for days --5, 0, +5 and +8,of the composite. MSU precipitation rate >8 mm day-1 has been superimposed. A comparison of the day --5 and +8 shows the differences in the sign of the anomalous monsoon gyre. The bifurcation of the precipitation events (days 2 to 6, Figure 6b) are accompanied by basin-scale troughs that migrate poleward from the equator in both hemispheres. Vector magnitude and precipitation rates are shown below the panels.

Figure 7: Time-latitude sections of zonally integrated ocean heat flux averaged across the Indian Ocean after Loschnigg and Webster (2000). The model used is an intermediate ocean model of McCreary et al. (1993) forced by 5--day average winds and insolation from the NCEP/NCAR reanalyses. Upper panel shows the mean

annual cycle from 40 years of forcing. The bottom two panels show the heat transport for the individual years of 1987 and 1988 which correspond to weak and a strong summer monsoon years, respectively. Units PW.

Figure 8: The principal platform used in the first two phases of JASMINE was the NOAA Research Ship *Ronald H. Brown*. Numbers on the figure refer to the location of many of the instruments used during JASMINE and noted in Table 1. During Phase III of JASMINE, the Australian Research Ship *Franklin* was used.

Figure 9: The composite precipitation pattern associated with MISOs as a function of time and latitude. The pattern was used for experimental design with the proposed cruise path shown as a black line. Plans called for sampling the atmosphere and ocean in an "undisturbed" state before the advent of an MISO, and in the "disturbed" state after the MISO had formed. Two sampling techniques were to be used: north-south transects along 89E and two "on-station" periods near 5N. As it turned out, the sequence of meteorological events encountered in the eastern Indian Ocean followed the composite almost exactly (see Figure 11).

Figure 10: The three phases of the field phase of JASMINE showing the cruise paths in Julian days. Phase I (panel a) was made up of transects along 89E. In Phase II (panel b), transects along 89E were accompanied by "on-station" periods, each five days in duration, where star patterns were executed (Star 1 and 2: see inset). Each star pattern was planned to take 18 hours so that about 6 circuits were completed during each 5--day period. During Phase III (panel c), the Australians executed triangle maneuvers (see inset) each taking 8 hours. During the cruise, CTD casts and upper-atmosphere radiosonde measurements were made. The number and locations of these observations are recorded in Table 2. All other instruments described in Table 1, and shown in Figure 11, were run in continual mode. Way-points in Julian days are noted in the figure.

Figure 11: Time-latitude sections of brightness temperature (see color-coded scale) from the European Space Agency METEOSAT-5 geostationary satellite (a) April, 2000, (b) May, 2000, (c) June, 2000, (d) July, 2000, and (e) September, 2000. All sections are averaged between 85E and 90E. Ship tracks for Phase I, II and III are shown in panels (a) (b) and (e), respectively.

Figure 12: Daily averaged (a) 925 mb and (b) 200 mb vector winds from ECMWF along 89E for Phase II of JASMINE. Color coding denotes speed relative to the scale to the right of the panel.

Figure 13: (a) The synoptic situation during Star 1 of JASMINE. Both 24--hour average brightness temperature and surface pressure fields are shown for the period 12--15 May, 1999. The period is highlighted by weak surface winds and strong insolation.

(b) Same as (a) except for Star 2 during 22--25 May, 1999. The white dot denotes the location of the *Ronald H. Brown*.

Figure 14: (a) The temperature structure (C) of the upper ocean for the five transects of Phase I and II of JASMINE plotted as a function of depth and latitude. Vertical lines are one day apart. Sections are arranged from north to south and the arrows indicate the direction of the transects. (b) Same as (a) except for salinity (psu). Note the large difference in the salinity between the northern (right) and southern ends of the Bay of Bengal. (c) Same as (a) except for the zonal currents (m s-1). Note the reversal of zonal current with latitude. Near the equator, the zonal current reverses direction twice, going through a complete oscillation in a 50--day period. There is also some evidence of vertical wave propagation in the structure. Overall, the ocean currents show a strong intraseasonal oscillation.

Figure 15: (a) Temperature (upper panel) and salinity (lower panel) plotted as a function of depth for Star 1.(b) Same as (a) except for Star 2. Note cooling and deepening and slight freshening of upper 30 m. (c) Same as (a) except for Phase III from the CSIRO Research Ship *Franklin*. Time in Julian days. Vertical line denotes 00 UTC of each day.

Figure 16: Time-height sections of horizontal wind (vectors: m s-1) and relative humidity (%) (upper panels) and radar reflectively (lower panels percent with values 15 dbZ < Z <3 5 dbZ in blue and Z > 35 dbz in red) for (a) Star 1 and (b) Star 2. During the undisturbed period (Star 1) the middle and upper troposphere is extremely dry. Little convection beyond trade cumuli existed at this time. During the disturbed period (Star 2) the middle and upper troposphere has markedly moistened, stronger winds prevail throughout the column and convection is deeper and stronger. Note the marked nocturnal maximum in the convection. Time in Julian days.

Figure 17: Components of the surface energy balance (W m-2) for (a) Star 1, and (b) Star 2. Five surface fluxes are shown: the solar radiative flux, sensible turbulent heat flux, latent turbulent heat flux, net longwave flux (outgoing surface minus incoming atmospheric), and the sensible heat flux of rainfall. Units W m-2.

Figure 18: Daily averaged net heat flux into the ocean during (a) Phase II and (b) Phase III. The shaded areas indicate the on-station star patterns (Phase II) or the triangle patterns (Phase III). Units W m-2.

Figure 19: The average diurnal variability of the (a) surface and 10 m air temperature (SST, TA: C), (b) surface wind speed (WS: m s-1), and (c) rain rate (R: mm hr-1) for the Star 1 (dashed lines) and Star 2 (solid lines).

Figure 20: Mean diurnal variability of the upper ocean temperature anomaly (C) for (a) Star 1, and (b) Star 2, as a function of depth.

Figure 21: The diurnal variation of the surface fluxes between the atmosphere and the ocean for the periods of Star 1 (dashed) and Star 2 (solid): (a) the surface turbulent fluxes and the net longwave radiation, (b) the solar and downwelling longwave radiation, and (c) calculations of the diurnal variability of the shortwave and longwave surface cloud forcing. Units W m-2.

Figure 22: (STILL TO BE DEFINED: HOUZE AND SERRA) Radar images of the diurnal event. (a) Cloud radar images between 12--24 UTC on 22 May, 1999. The radar shows extremely strong reflectivity up to 14 km as the line passes the ship between 15--18 UTC. (b) Horizontal scan of the of the Doppler radar at 1704 UTC, 22 May, 1999. (c) Vertical scan of the Doppler radar at 1708 UTC showing a complex convective structure of the squall line.

Table Captions:

Table 1: The observation system on the NOAA Research vessel *Ronald H. Brown.* Left hand column indicates the instrument number which is used to show the location of the instrument in Figure 8. The second column lists the instrument and the third column shows its utility.

Table 2: The three phases of JASMINE showing the dates of the north-south transects and their way-points. During Phase II and III there were periods were the ship remained “on station” executing maneuvers around a specific point for a number of days. These are referred to as Stars or triangles, depending on the shape of the maneuver. The NOAA Research Ship *Ronald H. Brown* was the principal research platform used during Phase I and II. The Australian CSIRO Research Ship *Franklin* was used during Phase III.

Table 3: The number of upper atmosphere soundings and CTDs by Phase and leg.

Table 4: A comparison of Indian Ocean fluxes obtained during JASMINE with those obtained in process experiments in the western Pacific Ocean. Units W m^{-2}.