JASMINE: The Scientific Basis

P. J. Webster
Program in Atmospheric and Oceanic Sciences:
University of Colorado, Boulder, CO.

CAPSULE: The general failure of climate models to reproduce the mean patterns of monsoon precipitation and their variability, and the inability of numerical predictions of monsoon variability to replicate the skill shown using empirical methods, points to a critical need for more data from the Indian Ocean region and a better understanding of coupling of the atmosphere and ocean. The recent Joint Air-Sea Monsoon Interaction Experiment (JASMINE) aimed to answer these needs.

Abstract

Variability of the Asian-Australian monsoon over a wide range of time scales produces both hardships and benefits on a vast population. Yet, despite its importance, we have only a limited knowledge of the processes that maintain the monsoon and its variability. As a consequence, the World Climate Research Programme's CLIVAR program has emphasized study of the monsoon climate system as a topic of some significance. A recent workshop of the US Global Ocean-Atmosphere-Land System (GOALS) identified as a highest priority the study of the coupled ocean-atmosphere system of the monsoon on intraseasonal time scales with emphasis on the eastern Indian Ocean region. These objectives led to plans for the Joint Air-Sea Monsoon Interaction Experiment (JASMINE).

This paper presents the scientific basis for JASMINE. Using data and model simulations, the climates of the Indian and Pacific oceans are compared. In contrast to the Pacific Ocean, relationships between long-term mean convection and sea-surface temperature (SST) in the Indian Ocean are essentially uncorrelated, suggesting a strong role of organized dynamic phenomena. The heat balance of the Indian Ocean and Pacific Ocean warm pools are also very different. For example, heat balance residual calculations and model results suggest a large (2 PW) oscillation in the annual cycle of the cross-equatorial meridional ocean heat transport. This is far larger than occurs in the Pacific. The character of intraseasonal variability in the two oceans differs, as revealed by composites of SST, measures of convection, and near-surface wind fields. Coherent oscillations appear to exist over the entire tropical Indian Ocean basin, including transitions between "active" and "break" periods of the South Asian monsoon. Analysis suggests that intraseasonal oscillations may be thought of as building blocks of the monsoon annual cycle and its interannual variability.

JASMINE was conducted in the eastern Indian Ocean and Bay of Bengal during the summer of 1999 as a joint venture between the United States and Australia. The principal purpose of the pilot study was to measure the atmospheric, oceanic, interface, and convective characteristics of the intraseasonal oscillations of the monsoon.

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) (Wallace et al. 1998). This progress was under the auspices of the international Tropical Ocean-Global

Atmosphere Programme (TOGA), which focused attention on the tropical Pacific. Now, based on this success, the study of other tropical regions and of higher latitudes is the fundamental objective of the U.S. contribution to the Global Ocean-Atmosphere-Land System program (GOALS) (NRC 1994) of the international CLIVAR (Climate Variability and Predictability Programme: WCRP 1995, 1998). These efforts aim "to understand and quantify" fluctuations in major tropical and subtropical heat sources and sinks and thereby improve predictions of weather and climate globally, not just in the tropics (NRC 1998).

The Indian Ocean-western Pacific region, which encompasses the Asian-Australian monsoon, requires special attention (NRC 1998). The Asian-Australian monsoon system embraces a large part of the globe, spanning the tropics and subtropics of the eastern hemisphere. The circulation embraces a majority of the people on Earth including the most rapidly increasing populations. Thus there are both scientific and societal needs that make the study of the predictability of this monsoon a priority.

Given the scientific advances made toward understanding the coupled ocean-atmosphere system elsewhere in the tropics, it would seem that we are poised to advance significantly understanding and prediction of Asian-Australian monsoon variability. To do so we need more data of fundamental processes that create variability of the monsoon.

At a workshop on the Variability of the Asian-Australasian Monsoon in July 1998 in St. Michaels, Maryland, participants wondered about the role of external forcing in producing monsoon variability (as suggested by Charney and Shukla 1980). Some external forcing may be from remote sources like ENSO, from local SST variations, or from ground moisture fluctuations. They also wondered about the role that higher-frequency (intraseasonal) hydrodynamical instabilities of the monsoon circulation might have in producing interannual variability of the system. Are these medium frequency oscillations controlled and modified by large-scale planetary forcing or do intraseasonal instabilities themselves determine interannual monsoon variability as suggested by Ferranti et al. (1997)?

These questions largely define the limitations to predicting monsoon variability. The monsoon may be slave to other climate systems, like ENSO, or may have its own modes of variability resulting, perhaps, from the coupling of land, atmospheric, and oceanic variability. It may be a hybrid that is forced remotely to some extent yet contains its own local modes of variability (which may have some chaotic element) that are modified by the remote forcing (Palmer 1994, Webster et al. 1998).

Advances are limited by the lack of data in the region and an inability to identify the fundamental processes that create the variability. Little is known about the state of the upper Indian Ocean. Despite a few field campaigns (from INDEX in the mid-1970s to the WOCE Indian Ocean Expedition), there have been relatively few direct measurements in the Indian Ocean basin. Despite climatologies of ship observations (such as Oberhuber 1988), research quality observations (e.g. Hacker et al. 1998) are rare.

More data is needed to establish relationships between the atmosphere and ocean on intraseasonal time scales, which have proven important in other oceans (Lukas and Lindstrom 1991, Vialard and Delecluse 1998, Wang et al. 1999, 2000). For instance, research in the Pacific warm pool had shown the importance of the barrier layer in complex atmosphere-ocean interactions. This salinity stratification retains heat and momentum in the upper layer of the western Pacific (Lukas and Lindstrom (1991). Vialard and Delecluse (1998) showed that westerly wind bursts propel the barrier layer eastwards, moving the impacts of the insulating layer to other regions of the warm pool every month or so. This movement might favor the growth of unstable air-sea interactions in the central Pacific. Thus Vialard and Delecluse argue that the transience of the barrier layer is a key to coupled ocean-atmosphere instabilities and we expect it to be equally significant to intraseasonal oscillations in the summer monsoon. Yet, before JASMINE, the suspected upper ocean barrier layer in the eastern Indian Ocean had not even been documented (Hacker et al. 1998).

The participants at the workshop in 1998 agreed that we need more observations of the Indian Ocean basin. Specifically, there are three foci for study. In order of priority, based on lack of research and observations, they are: the eastern Indian Ocean, Bay of Bengal and western Pacific Ocean; the Indonesian Archipelago, Indonesian seas, and connection between the Indian and Pacific oceans (the throughflow); and the Arabian Sea and southern Indian Ocean. The Joint Air-Sea Monsoon Interaction Experiment (JASMINE), held in the eastern Indian Ocean during the summer of 1999, addressed these data needs with a variety of observational equipment and strategies (see companion paper, JASMINE: The Field Phase). The experiment aimed to provide better understanding of the oceanic and atmospheric fluxes and variability during the critically important monsoon of this region.

2. Importance of the eastern Indian Ocean

There are a additional reasons for immediate emphasis on the eastern Indian Ocean enhanced research and observing. The region has great oceanographic significance. It is the entrance of the Indonesian throughflow that connects the Pacific and the Indian oceans. Here, the very saline waters of the western Indian Ocean and the fresh water of the eastern basin mix, determining salt balance for the entire basin. The eastern Indian Ocean is also a region of considerable dynamic activity. Upper ocean currents reverse on seasonal and intraseasonal time scales, and the region also contains a pole of the newly discovered Indian Zonal Mode, or Indian Ocean "dipole" (Webster et al. 1999, Saji et al. 1999, Yu and Rienecker 1999, 2000; Webster et al. 2001a) Furthermore, there appears to be a substantial oceanic barrier layer similar to that originally documented in the Pacific Ocean by Lukas and Lindstrom (1991). Hacker et al. (1998) observed this in the northeast (winter) monsoon and it is expected to persist through summer.

The eastern Indian Ocean is also critical from an atmospheric perspective, with precipitation extremes falling in the northern Bay of Bengal and just south of the equator. In fact, satellite estimates indicate that the northern Bay of Bengal receives the largest mean precipitation in South Asia during the summer monsoon. Intraseasonal oscillations of the monsoon reach maximum amplitude in the eastern Indian Ocean, too. It is a source for "active" periods of the monsoon (with enhanced precipitation) and break periods (with short-lived droughts or lulls in the precipitation rate) (Webster et al. 1998). Furthermore, the biennial oscillation possesses strong signals in the eastern Indian Ocean (Meehl 1994) and is a center of strong interannual rainfall variance. While many of these features have been known to exist for a very long time (see reviews by Godfrey 1995, Webster et al. 1998), the physical processes that maintain them are not understood.

The potential rewards for increasing our understanding of the monsoon climate are substantial. Forecasting variability of monsoon rainfall from year-to-year with sufficient lead time would allow remedial action by managers of agricultural and water resources. But forecasting the onset of the monsoon and the occurrence, magnitude, and duration of active and break periods during a monsoon season 10-14 days in advance may have a far greater potential impact on an agrarian society than forecasting whether or not the entire monsoon summer rainfall may be anomalous (A. B. Subbiah, Asian Disaster Preparedness Center, Bangkok, Thailand, personal communication). Plowing and planting periods are extremely vulnerable. Once the monsoon begins, the timing of the first break becomes critical. Even if the average seasonal monsoon rains are normal, an ill-timed break in the rain can devastate a local economy (Webster et al. 1998).

To a large degree, interannual variability appears to be associated with the frequency of intraseasonal variability. Hendon et al. (1999) has found that wet Australian monsoon years are associated with fewer break periods and, thus, less intraseasonal variability. Lawrence and Webster (2001a) reached similar conclusions for the Asian summer monsoon. The difference between 1987 and 1988 rains is an example (see Figure 1). Overall, precipitation during 1988 was 27% greater than during 1987. During the summer monsoon, June through September, precipitation was 53% greater. Rainfall started earlier in 1988 but had essentially ceased by the end of September. In 1987, on the other hand, 24% of precipitation occurred after September. The intraseasonal variability was also very different in each year. A break in the monsoon is defined as a period in which three successive pentads (15 days) had less than 50 mm of precipitation. Two breaks occurred in 1987 (in late July and early August, and in late August until the middle of September) while none occurred in 1988.

It has been argued that interannual variability of large-scale climate systems such as the monsoon and ENSO cannot be understood without comprehending the physical processes that control the annual cycle (e.g. Webster and Yang 1992). By extension, it can also be argued that the annual cycle of the monsoon cannot be understood (or perhaps predicted) without a thorough knowledge of the physics of intraseasonal variability. Indeed, intraseasonal variability appears to be a fundamental building block of both the annual cycle and its interannual variability (Ferranti et al. 1997).

If this is true, it is probably not a coincidence that the inability of numerical models to simulate the mean seasonal precipitation patterns of the regional monsoons (Sperber and Palmer 1996, Gadgil and Sanjani 1998) is matched by the model's inability to possess realistic intraseasonal variability, or any variance at all at these time scales (Slingo et al. 1996). In summary, understanding the physical processes that produce intraseasonal variability in the monsoon stands as the fundamental problem in monsoon climate studies. JASMINE addressed this issue of the intraseasonal variability of the monsoon. It is both the major area of uncertainty in understanding and probably the stumbling block for successful modeling and quantitative prediction of the monsoon.

3. Modeling Monsoon Variability

The CLIVAR Implementation Plan calls for determining the inherent predictability of the monsoon system and developing of methods of prediction (WCRP 1998). To date, empirical prediction has proven to be more successful in South Asia than numerical prediction (e.g. Shukla and Paolina 1981, Hastenrath 1994; Webster et al. 1998, Sadhuram 1997, Hazzallah and Sadourny 1997, Clark et al. 2000). Numerical simulation of the mean seasonal monsoon precipitation relative to known lower boundary conditions has proven difficult (Sperber and Palmer 1994). As Figure 2 shows, different models don't agree well with observations or with one another on estimates of seasonal precipitation over India (Gates 1992). Members of an ensemble of realizations using the same model (not shown) are also quite different (Brancovic and Palmer 2000). Model performance for this system is far worse than in most parts of the tropics. Simulations of western Pacific Ocean precipitation, for example, are similar from one model to another and are also closer to observed values.

The failure of atmospheric models to simulate monsoon precipitation on almost any time scale is troublesome and does not appear to bode well for the numerical prediction of monsoon variability. If intraseasonal oscillations are chaotic, as suggested by Palmer (1994), then interannual variability of the monsoon would be essentially unpredictable beyond the intrinsic time scale of fundamental instability. Gross-scale influences of boundary effects (e.g. El Niño or the Indian Ocean Zonal Mode) would nudge the chaotic intraseasonal variability towards an increase or decrease of the probability of occurrence of active or break periods within the monsoon season. This theory, referred to as "nudged chaos" (Webster et al. 1998), concludes that the predictability of seasonal monsoon rainfall is probabilistic. However, the forecasting of individual monsoon intraseasonal oscillations through one cycle may be more or less deterministic. This situation is analogous to the extratropics, where the life cycle of baroclinic waves (the dominant instability) define the limit of prediction.

But forecasting through one cycle of the major instability of the monsoon assumes that current models are even capable of simulating intraseasonal variability of the monsoon. The models may not be capable of this. Slingo et al. (1996) showed that models tend to understate intraseasonal variance. The models may not oscillate between the phases of the intraseasonal variability but remain fixed in one phase or the other. Gadgil and Sajani (1999) provide some support for this idea: they show that models that underestimate seasonal precipitation tend to focus rainfall to the south of India. Those that overestimate rainfall concentrate precipitation over India and underestimate it near the equator. That is, numerical models appear to lock onto a configuration that is either a season-long active phase or break phase of the monsoon. One possibility is that certain physical processes are absent from atmospheric models.

See SIDEBAR, relative to Figure 3, in figure captions.

If the monsoon intraseasonal oscillations are coupled ocean-atmosphere phenomena, then a fixed SST boundary in the models may not permit feedbacks associated with the transition between active and break periods. Furthermore, current horizontal resolution and parameterizations may be insufficient to replicate monsoon variability. The complex geography and tight heating gradients in the Asian monsoon system may tax the current models to their limit. In fact, simulation of monsoon rainfall improves when model resolution is increased substantially (e.g. Shukla and Fennessy 1994).

There is other evidence that atmosphere-ocean coupling is essential to monsoon variability. Figure 4 shows interannual variability of mean net surface flux into the North Indian Ocean. During boreal spring and fall, a time of low winds and high insolations, mean net surface flux into that region is strongly positive. During winter, it is weakly negative. Compensating for these surface fluxes is a strong heat flux in the ocean across the equator: northward in winter and southward in spring through early fall. Values range from +2 to -2 PW. Loschnigg and Webster (2000) noted that the heat flux is generally in the opposite sense to the sign of the divergent component of the lower tropospheric monsoon wind field (Figure 5), which is north in summer and south in winter. They attributed the reversed cycle of heat flux to wind-forced Ekman transports. Other studies (Hsiung et al. 1989, Hastenrath and Greichner 1993, Wacogne and Pacanowski 1996) find similar annual cycles of meridional heat transport with similar amplitude and phase. In climatological averages , the seasonal swings of oceanic cross-equatorial heat transport compensate with a residual southward cross-heat transport (of about -0.15 PW) that matches the net annual mean surface heating of the North Indian Ocean. However, on a year-to-year basis, this compensation is not assured. Considerable interannual variability is not found in each component of the heat balance. For instance, the weak monsoon of 1987 had higher net surface fluxes and weaker southward transports of heat, presumably both associated with clearer skies and generally lighter surface winds. During the strong monsoon of 1988, increased cloudiness and stronger winds decreased the heat transport substantially. Together these years balance out to climatological values. This biennial compensation allowed Loschnigg and Webster (2000) to speculate that the coupled dynamics of the monsoon system act to regulate the monsoon system which is examined more thoroughly by Webster et al. (2001a). For the purposes of this paper, the numerical experiments suggest that the monsoon is a coupled ocean-atmosphere phenomenon with seasonally reversing flow and heat transports in both the ocean and atmosphere.

Whatever the reason for the failure of the models, quality data will be necessary to decipher and model the physical processes that produce monsoon intraseasonal variability. The central

Table 1 Precipitation statistics for 1987 and 1988 at the point (70°E, 15°N) derived from a blended precipitation product using station rain-gauge and satellite microwave sounding data (MSU). Data are plotted as rainfall per pentad (mm/ 5 day).

	Precipitation (mm)				Number of breaks (JJAS)
	Jan-Jun	Jun-Sept	Sept-Dec	Annual total	
1987	65	1363	450	1878	2
1988	246	2065	180	2491	0

focus of JASMINE was to obtain data of coupled processes in the monsoon region.

4. Intraseasonal Variability of the Monsoon

Intraseasonal variability of the monsoon shows up in vast spatial scales. During an active period, above-average and coherent precipitable water content (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. Such large scales of low-frequency variability are apparent in multiple data sets such as precipitation, OLR, and wind fields (e.g. Webster et al. 1998).

Arc or boomerang shaped precipitation maxima appear to emanate slowly from the equator toward both the north and south. These features appear every 20 to 40 days (the scale of monsoon interannual variability) and seem to propagate from the equator to 30°N in about 10 days. The southward leg tends to dissipate relatively rapidly, presumably because it propagates over colder water.

Similar patterns have been noted by a number of earlier studies (e.g. Sikka and Gadgil 1980, Yasunari 1980, Webster 1983, Hartmann and Michelson 1989, Srivasian et al. 1993). Lawrence and Webster (2001b) suggest that the bifurcating cyclonic vortices or troughs represent trailing Rossby waves behind a rapidly propagating Kelvin wave, in agreement with the conclusions of Wang and Rui (1990). The northward propagating convection, associated with the vortex in the eastern Indian Ocean, appears as a consistent precursor of an active phase of the South Asian monsoon.

The onset of the monsoon could be defined as the first major intraseasonal oscillation accompanied by a northward propagation of the precipitation maximum from the equator into the Bay of Bengal region. For example, the monsoon onset in 1987 (Figure 5b, upper panel) follows enhanced precipitation at the equator in early June. In 1988 (Figure 6b, lower panel), the onset defined in this manner occurred about two weeks earlier. This definition is very different from traditional definitions (e.g. Rao 1976), which emphasize the beginning of persistent rainfall at the southern 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 monsoon intraseasonal oscillation matures and moves northward into the northern Bay of Bengal. Viewing the monsoon from a wider perspective allows the possibility of defining a monsoon within a larger geophysical context. Figure 6 gives such a basin-wide view of the local variability seen in Figure 1.

In a composite map of OLR during intraseasonal oscillations of the monsoon (Figure 7a), one can see that organized convection moves to the eastern Indian Ocean and deepens in eight days (to a point at which intense precipitation starts at 90(E and the equator, defined here as day 0). At that stage, the convective mass "bifurcates." The northern element moves slowly up through the Bay of Bengal and eventually over east India, Bangladesh, and Southeast Asia. Convection increases along the west coast of India after day +4 coinciding with the spin up of the active phase of the monsoon. The southern element moves southward and eastward, forming an elongated cloud band across Australia. The southern element, which is not as long-lived or intense as its northern counterpart, is still responsible for the winter hemisphere precipitation maximum just south of the equator. There is also some evidence that these southern branches are responsible for roughly 40% of the wintertime Australian rainfall. Such associations were found by Nicholls (1989) on interannual timescales and Fasullo and Webster (1999) on shorter time scales.

The SST fields (Figure 7b) also show a coherent variation over the period of the composite monsoon intraseasonal oscillation. Anomalous warming along the equator yields to cooling as very deep convection builds up over the eastern Indian Ocean. These changes in SST are directly related to basin-scale variations in the lower tropospheric wind field (Figure 7c). The evolution of the wind field in the composite has three main periods: general anomalous counterclockwise circulation before the growth of convection along the equator, then strong convergence into the western end of the convection as the convection moves east, and finally a strengthening gyre with the bifurcation of the troughs.

Figure 8 shows that the SST has a basin-wide maximum near day -10 and a minimum at day +5. These extremes correspond to a minimum in the strength of the monsoon westerlies and a maximum in OLR (minimum in convection). However, the maximum in convection leads maximum monsoon winds and minimum SST by about 5 days, indicating a coupling between the atmosphere and ocean on intraseasonal time scales. The results are similar to those found by Fasullo and Webster (1999), and intraseasonal oscillations in the TOGA COARE region of the western Pacific had similar phase relationships.

Although the SST variations represent coherent patterns across the entire Indian Ocean, the extrema are fairly small, ranging between ±0.2°C through the phases of the intraseasonal oscillation. However, it is likely that the magnitudes of SST variability displayed in Figures 7b and 8, and computed from the Reynold’s SST fields, are much smaller than those which are actually associated with the monsoon oscillation. SST variations measured during intraseasonal variations occurring in TOGA COARE showed variability between a half and an order of magnitude larger (e.g., Webster 1994, Godfrey et al. 1998). Similar variations were encountered during JASMINE (Webster et al. 2001b). Furthermore, (Sengupta and Ravichandran 2001) has shown using moored buoys that SST in the Bay of Bengal varies about ±1°C through an intraseasonal oscillation. The reason for the underestimation may result from the fact that the Reynold’s SST is a 5–day average and is computed from satellite data that is regressed to match ship observations which are perhaps more representative of subsurface temperatures than surface skin SST.

It is unclear from these SST changes, however, whether or not the wind variations produce a dynamic response in the ocean and whether this response is important to the overall heat balance of the Indian Ocean. In the absence of sufficient data to answer this question, we turn to results from the McCreary et al. (1993) intermediate ocean model. Results indicate a substantial oceanic response on intraseasonal time scales. The heat flux during winter is to the north on the order of 2 PW and is slightly stronger southwards during the boreal summer. Both 1987 and 1988 possess low frequency modulation of north-south heat transports on the 20-40 day time scale. During summer the southward heat transport varies between 0 and -4.5 PW, compared to the climatological maximum of -1.5 to -2 PW. In addition, the intraseasonal variability in each of the two years is very different.

A comparison of Figures 6b and 9 shows the impact of intraseasonal forcing of the atmosphere on the ocean. There is near one-to-one correspondence between the atmospheric intraseasonal convection and the zonally integrated ocean heat flux. For example, the first major southward transport in 1987 in early June (Figure 9) corresponds to the first near-equatorial convection (Figure 6b). Similar correspondences between southward heat flux and convection are evident several times in 1988.

The intraseasonal forcing by the atmosphere can be seen in Figure 7c. Winds along the equator accelerate eastward substantially as the convection grows. At the same time, winds accelerate northward in the western Indian Ocean and increase southward in the eastern basin as convection bifurcates (day +4, Figure 8). The ocean responds strongly to this forcing. The question arises, amid evidence of strong oceanic heat transport variations, as to whether or not the oceanic variability feeds back onto the atmosphere on intraseasonal time scales. This ocean-to-atmosphere feedback has yet to be determined, but it is probably safe to say, at this stage, that the intraseasonal variability of ocean heat transport is an important component of the annual heat balance of the Indian ocean coupled ocean-atmosphere system.

It is also unclear to what degree the intraseasonal variability is responsible for the interannual variability. And it is unclear if year-to-year differences in intraseasonal variability result from a low-frequency interannual variability of the basic state in the region. Modelling studies (Ferranti et al. 1997) show that principal intraseasonal and interannual components map onto each other, suggesting that interannual variability of the monsoon is closely tied to intraseasonal variations. Whether the same is true for the ocean is not known.

5. Rationale for JASMINE

A number of scientific questions arise related to the issues discussed above. Some questions related to the coupling of the ocean and atmosphere. To what degree is the intraseasonal variability of the monsoon a coupled ocean-atmosphere phenomenon? Does the monsoon intraseasonal oscillation modulate the structure of the Indian Ocean barrier layer and is it an important component of the heat balance of the tropical Indian Ocean? Vialard and Delecluse (1998) found strong connections of this type in the Pacific. Are their conclusions transferable to the Indian Ocean?

Other questions relate to the structure of the variability. Are there consistent and coherent patterns of intraseasonal monsoon variability such that there may be a canonical monsoon intraseasonal oscillation? If so, could this canonical structure be used for forecasting active and break periods of the monsoon? What is the relationship of the onset of the South Asian monsoon and intraseasonal variability? Does viewing the monsoon onset as the first summer monsoon intraseasonal oscillation increase the possibility of its prediction?

With these questions in mind, 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. JASMINE was a bilateral effort of the United States and Australia resulting in intense observations of the monsoon for over 52 days in the summer of 1999. The experiment took place in the southern Bay of Bengal and the equatorial eastern Indian Ocean. The principal platforms used in JASMINE were the NOAA ship Ronald H. Brown in spring and early summer and the Australian R/V Franklin in September.

JASMINE followed the Indian Ocean experiment (INDOEX: Lelieveld 2001) that studied anthropogenic aerosol distributions during winter and spring. Also, in July 1999 Indian scientists conducted a national joint meteorology-oceanography experiment (BOBMEX: Bay of Bengal Monsoon Experiment: Bhat et al. 2001) in roughly the same location as JASMINE. During 1999, the European geostationary satellite METEOSAT-5 was relocated over the central Indian Ocean to provide an unprecedented view of the monsoon. Indeed, the summer of 1999 was a unique season for observation of the coupled ocean-atmosphere system in the eastern Indian Ocean. Initial results of the field phase of JASMINE appear in a companion paper (Webster et al. 2001b).

Acknowledgments: The organization of JASMINE came from recommendations from the St. Michael's Conference on the Variability of the Asian-Australian Monsoon System held in July, 1998. Funding for the conference came from the Climate Dynamics and Large Scale Dynamic Meteorology Programs of the National Science Foundation, the NOAA Office of Global Programs, and NASA. Specific funding for this paper came from Climate Dynamics and Large Scale Dynamic Meteorology Programs of the Atmospheric Sciences Division of the National Science Foundation under NSF Grant ATM-9819618.

References:

Bhat, G. S., S. Gadgil, P. V. Harish Kunmar, S. R. Kalsi, P. Madhusoodanan, V. S. N. Murty, C. V. K. Prasada Rao, V. Ramesh Babu, L. V. G. Rao, R. R. Rao, M. Ravichandran, K. G. Reddy, P. Sanjeeva Rao, D. Sengupta, D. R. Sikka, J. Swain, P. N. Vinayachandran, 2001: BOBMEX-The Bay of Bengal Monsoon Experiment. *Bull. Amer. Meteor. Soc.* **82**, 2217–2245.

Brancovik C. and T. N. Palmer, 2000: Seasonal skill and predictability of ECMWF PROVOST ensembles. *Q. J. R. Meteorol. Soc.*, **126**, 2035–2067.

Charney, J. G. and J. Shukla, 1981: Predictability of monsoons. *Monsoon Dynamics*, edited by J. Lighthill, 99-110, Cambridge University Press.

Clark, C. O., J. E. Cole and P. J. Webster, 2000: SST and Indian summer rainfall: predictive relationships and their decadal variability. *J. Clim.*, **13**, 2503–2519.

Fasullo, J. T. and P. J. Webster, 1999: Warm pool sea surface temperature variability in relation to the surface energy balance. *J. Clim.*,**12**, 1292–1305.

Ferranti, L., J. M. Slingo, T. N. Palmer and B. J. Hoskins, 1997: Relations between interannual and intraseasonal monsoon variability as diagnosed from AMIP integrations, *Q. J. R. Meteorol. Soc.*, **123**, 1323–1357.

Gadgil, S. and S. Sanjani, 1998: Monsoon precipitation in the AMIP runs, *Clim. Dyn.*, **14**, 659-689.

Gates, W. L., 1992: The Atmospheric Model Intercomparison Project, *Bull. Amer. Meteorol. Soc.*, **73**, 1962-1970.

Godfrey, J. S. 1995: The role of the Indian Ocean in the global climate system: Recommendations regarding the global ocean observing system. Report of the Ocean Observing System Development Panel, *Background Report #6*, 89 pp. Texas A&M Univ., College Station, TX.

Godfrey J. S., Houze R. A., Johnson R. H., Lukas R., Redelsperger J. L., Sumi A., Weller R., 1998: Coupled Ocean-Atmosphere Response Experiment (COARE): An interim report. *J. Geophys. Res. (Oceans)*, **103**, 14395-14450.

Hacker, P., E. Firing, J. Hummon, A. L. Gordon, L. Kindle, 1998: Bay of Bengal currents during the northeast monsoon *Geophys. Res. Letters*, **25**, 2769-2772.

Hartmann, D. L., and M. L. Michelsen, 1989: Intraseasonal periodicities in Indian rainfall. *J. Atmos. Sci.*, **46**, 2838-2862.

Hastenrath, S., 1994: *Climate Dynamics of the Tropics: An Updated Edition of Climate and Circulation of the Tropics*, 488pp, Kluwer Academic Publishers, Norwell, Mass..

Hastenrath, S., and L. Greischar, 1993: The monsoonal heat budget of the hydrosphere-atmosphere system in the Indian Ocean sector. *J. Geophys. Res.*, **98**, 6869-6881.

Harzallah A. and Sadourny R. 1997: Observed lead-lag relationships between Indian summer monsoon and some meteorological variables *Clim. Dyn*, **13**, 635-648.

Hendon H. H., Zhang C. D., Glick J. D. 1999: Interannual variation of the Madden-Julian oscillation during austral summer J. Clim. *12*, 2538-2550.

Hsiung, J., R. E. Newell, and T. Houghtby, 1989: The annual cycle of oceanic heat storage and oceanic meridional heat transport. *Quart. J. Roy. Meteor. Soc.*, **115**, 1-28.

Kalnay, E., M. Kanamitsu, R. Kistler, W. Collins, D. Deaven, L. Gandin, M. Iredell, S. Saha, G. White, J. Woollen, Y. Zhu, A. Leetmaa, R. Reynolds, M. Chelliah, W. Ebisuzaki, W.Higgins, J. Janowiak, K. C. Mo, C. Ropelewski, J. Wang, R. Jenne, D. Joseph, 1996: The NCEP/NCAR 40-Year Reanalysis Project. *Bull. Amer. Meteor. Soc.* **77**, 437–471.

Lau K. M. and C. H. Sui, 1997: Mechanisms of short-term sea surface temperature regulation: Observations during TOGA COARE. *J Clim.* **10**, 465-472.

Lawrence, D, and P. J. Webster, 2001a: Interannual variability of intraseasonal convection and the Asian monsoon. *J. Clim.*, **14**, 2910–2922.

Lawrence, D, and P. J. Webster, 2001b: Intraseasonal oscillations in the South Asian monsoon. In press *J. Atmos. Sci.*.

Lelieveld, J., Crutzen P. J., Ramanathan, V., Andreae, M. O., Brenninkmeijer, C. A. M., Campos, T., Cass, G. R., Dickerson, R. R., Fischer, H., de Gouw, J. A., Hansel, A., Jefferson, A., Kley, D., de Laat, A. T. J., Lal ,S., Lawrence, M. G. , Lobert, J. M. , Mayol-Bracero, O. L., Mitra, A. P., Novakov, T., Oltmans, S. J., Prather, K. A., Reiner, T., Rodhe, H., Scheeren, H. A., Sikka, D., Williams, J., 2001: The Indian Ocean Experiment: Widespread air pollution from South and Southeast Asia, 2001: *Science*, 1031-1036 .

Loschnigg and Webster, 2000: A coupled ocean-atmosphere system of SST regulation for the Indian Ocean. *J. Clim.*, **13**, 3342–3360.

Lukas, R., and E. Lindstrom, 1991: The mixed layer of the western equatorial Pacific Ocean, *J. Geophys. Res.*, **96**, 3343-3357.

McCreary, J. P., P. K. Kundu and R. L. Molinari, 1993: A numerical investigation of the dynamics, thermodynamics and mixed-layer processes in the Indian Ocean, *Progress in Oceanogr.*, **31**, 181-244.

Meehl, G. A., 1994: Coupled ocean-atmosphere-land processes and south Asian Monsoon variability, *Science*, **265**, 263-267.

National Research Council (NRC), 1994: *GOALS; Global Ocean-Atmosphere-Land System for Predicting Seasonal-to-Interannual Climate*, 103 pp., National Academic Press, Washington, D. C.

National Research Council, 1998: A scientific strategy for U.S. participation in the GOALS (Global Ocean-Atmosphere-Land System) Component of CLIVAR. 69 pp, National Academic Press, Washington DC.

Nicholls, N., 1989: Sea surface temperatures and Australian winter rainfall *J. Clim.,* **2**: 965-973.

Oberhuber, J. M., 1988: An atlas based on the COADS data set: The budgets of heat, buoyancy and turbulent kinetic energy at the surface of the global ocean, Rep. 15, 20 pp., 160 figs, Max-Planck Institut, Hamburg, Germany.

Palmer, T. N., 1994: Chaos and the predictability in forecasting the monsoons, *Proc. Indian. Natn. Sci. Acad.,* Part A., **60**, 57–66.

Rao, Y. P., 1976: Southwest Monsoon, Meteorological Monograph Synoptic Meteorology No. 1/1976, Indian Meteorological Department, 367pp.

Reynolds, R. W., 1988: A real-time global sea surface temperature analysis. *J. Climate*, **1**, 75-86.

Sadhuram, Y. 1997: Predicting monsoon rainfall and pressure indices from sea surface temperature *Curr. Sci. India*, **72**, 166-168.

Saji, N. N., B. N. Goswami, P. N. Vinayachandran, and T. Yamagata, 1999: A diploe mode in the tropical Indian Ocean. *Nature,* **401**, 360–363.

Sengupta, D., Ravichandran, M. 2001: Oscillations of Bay of Bengal sea surface temperature during the 1998 summer monsoon. *Geophys. Res. Lettr.* **28**, 2033-2036.

Shukla, J. and D. A. Paolina, 1983: The southern oscillation and long range forecasting of the summer monsoon rainfall over India, *Mon. Weather Rev.*, **111**, 1830-1837.

Shukla, J., and M. J. Fennessy, 1994: Simulation and predictability of monsoons, in *Proceedings of the International Conference on Monsoon Variability and Prediction,* Tech. Rep. WCRP–84, 567–575, World Climate Research Programme, Geneva, Switzerland.

Sikka, D. and S. Gadgil, 1980: On the maximum cloud zone and the ITCZ over Indian longitudes during the southwest monsoon. *Mon Weather Rev.* **108**, 1840–1853.

Slingo, J. M., Sperber, K. S., Boyle, J. S., Ceren, J. P., Dix, M., Dugas, B., Ebisuzaki, W., Fyfe, J., Gregory, D., Gueremy, J. F., Hack, J., Harzallah, A., Innes, P., Kitoh, A., Lau, W. K. M., McAveney, B., Madden, R., Matthews, A., Palmer, T. N., Park, C. K., Randall, D., Renno, N., 1996: Intraseasonal oscillations in 15 atmospheric general circulations models: Results from the AMIP diagnostic subproject. *Clim. Dyn.*, **12**, 325–357.

Sperber, K. R., and T. N. Palmer, 1996: Interannual tropical rainfall variability in general circulation model simulations associated with the Atmospheric Model Intercomparison Project, *J. Clim.*, **9**, 2727–2750.

Srinivasan, J, S. Gadgil and P. J. Webster, 1993: Meridional Propagation of Large-Scale Monsoon Convective Zones. *Meteor. Atmos. Phys.*, **52**, No: 1–2, 15–35

Vialard, J. and P. Delecluse, 1998: An OGCM study for the TOGA decade. Part II: Barrier layer formation and variability. *J. Phys. Ocean.*, **28**, 1089–1106.

Wacongne, S., and R. C. Pacanowski, 1996: Seasonal heat transport in the tropical Indian Ocean. *J. Phys. Ocean.*, **26**, 2666-2699.

Wallace, J. M., Rasmusson, E. M., Mitchell, T. P., Kousky, V.E., Sarachik, E. S., von Storch, H. 1998: The structure and evolution of ENSO-related climate variability in the tropical Pacific: Lessons from TOGA. *J. Geophys. Res. (Oceans)*, **103**, 14241-14259.

Wang, B. and H. Rui, 1990: Synoptic climatology of transient tropical intraseasonal convection anomalies: 1975–1985: *Meteor. Atmos. Phys.*, **44**, 43–61.

Webster, P. J., 1983: "Mechanisms of Monsoon Transition: Surface Hydrology Effects". *J. Atmos. Sci.*, **40**, 2110–2124.

Webster, P. J. and S. Yang, 1992: "Monsoon and ENSO: Selectively Interactive Systems". *Quart. J. Roy. Meteor. Soc.*, **118**, 877–926.

Webster, P. J., and R. Lukas, 1992: TOGA COARE: The Coupled Ocean-Atmosphere Response Experiment, *Bull. Am. Meter. Soc.*, **73**, 1377-14162.

Webster, P. J. and R. A. Tomas, 1997: An atlas of precipitation and boundary layer winds during intraseasonal oscillations events in the Indian Ocean (1985–1995 extended summer season). Technical Report No. 2, Program in Atmospheric and Oceanic Sciences, University of Colorado, Boulder.

Webster, P. J., T. Palmer, M. Yanai, V. Magana, J. Shukla and A. Yasunari, 1998: Monsoons: Processes, Predictability and the prospects for prediction. *J. Geophys. Res. (TOGA special issue)*, **103**, 14,451–14,510.

Webster, P. J. A. M. Moore, J. P. Loschnigg, and R. R. Leben, 1999: Coupled ocean-atmosphere dynamics in the Indian Ocean during 1997–98. September 23, *Nature*, **401**, 356–360.

Webster, P. J., C. Clark, G. Cherikova, J. Fasullo, W. Han, J. Loschnigg, and K. Sahami, 2001a: The Monsoon as a self-regulating coupled ocean-atmosphere system. In *Meteorology at the Millennium.* Academic Press. (in press)

Webster, P. J., E. F. Bradley, C. E. Fairall, J. S. Godfrey, P. Hacker, R. A. Houze Jr., R. Lukas, Y. Serra, J. M. Humman, T. D. M. Lawrence, C. A. Russell, M. N. Ryan, K. Sahami, P. Zuidema 2001b: JASMINE: The Field Phase. Submitted to *Bull. Amer. Meteorol. Soc.*.

World Climate Research Programme (WCRP), 1990: Scientific plan for the Global Energy and Water Cycle Experiment (GEWEX), WMO/TD 376, World Climate Research Programme, Geneva, Switzerland, 137pp.

World Climate Research Programme (WCRP), 1995: CLIVAR: A study of climate variability and predictability, WMO/TD 690, Geneva, Switzerland, 157pp.

World Climate Research Programme (WCRP), 1998: CLIVAR Initial Implementation Plan, WCRP Report 103, WMO/TD 869, Geneva, Switzerland, 195 pp.

Yasunari, T., 1980: A quasi-stationary appearance of the 30–40 day period in the cloudiness fluctuations during the summer monsoon over India, *J. Meteorol. Soc. Jpn*, **58**, 225–229.

Yu, L. and M. Rienecker, 1999: Mechanisms for the Indian Ocean warming during the 1997–98 El Niño. *Geophys. Res. Lett.*, **26**, 735–738.

Yu, L. and M. Rienecker, 2000: Indian Ocean Warming of 1997–1998. *J. Geophys. Oceans*, **105**, 16923–16939.

Figure and Table Captions

Figure 1: Histograms of 5-day average rainfall at (70°E, 15°N) along the east coast of India for 1987 and 1988 representing, respectively, weak (1987) and strong (1988) summer monsoon precipitation seasons presumed to be associated with the 1987 El Niño and 1988 La Niña. Data is a blend of MSU satellite precipitation and land-based gauge data. Both interannual and intraseasonal variability are apparent. Table 1 summarizes the seasonal distribution of precipitation. Data available through pjw@oz.colorado.edu.

Figure 2: 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 from the seven “best” models of the Atmospheric Model Intercomparison Experiment (AMIP: Gates 1992). Each model was forced with identical observed SST distributions. Observed precipitation shown as the dashed gray contour.

Figure 3: Comparison of convection-SST relationships in the Indian and Pacific Ocean basins. OLR (W m^{-2}) is plotted against SST (°C) 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. OLR scale is inverted with increased convection being associated with decreasing OLR.

SIDEBAR for Figure 3: A Comparison of Variability in the Indian and Pacific Oceans

During the last decade there have been a large number of experiments in the Pacific Ocean aimed at understanding the relationship between the ocean and the atmosphere. Most notable of these is the TOGA Coupled Ocean-Atmosphere Response Experiment (TOGA COARE: Webster and Lukas 1992). A principal question arises. Is it possible to transfer the results of the many Pacific Ocean experiments directly to the Indian Ocean? Many similar phenomena exist such as large amplitude intraseasonal oscillations. However, specific comparisons suggest that many results don’t transfer well.

An immediate comparison of the Pacific and Indian Ocean regimes can be made by considering the outgoing longwave radiation (OLR, in Wm^{-2}: a surrogate for convection) and sea-surface temperature (SST, in (°C) in the two regions (Figure 3). There are fundamental differences. In the Pacific Ocean, when the SST $<$ 29°C, there is a relatively direct relationship with the intensity of convection growing fairly uniformly with SST. For SST > 29°C, however, the relationship disappears. The reason that there are long periods of either strong convection and an absence of convection at the warmest SST is relatively straightforward. In the warm pools of the tropics, SSTs are generally warmest during periods of low wind and high insolation, which occur during the undisturbed period of the intraseasonal oscillation (e.g., Lau and Sui 1997, Fasullo and Webster 1999). The period of the oscillation (30–60 days) is sufficiently long to produce mean monthly values of low convective activity when the SST is warmest.

By contrast, the northern Indian Ocean shows a conspicuous lack of SST-OLR correlation over the entire SST range. Between 25°C and 29°C, the near linear increase of convection with OLR observed in the Pacific is not apparent and the distribution is more similar to the convection-SST relationships occurring at the warmest temperatures in the Pacific Ocean. One interpretation of this is a greater dominance of intraseasonal variability in the Indian Ocean than in the Pacific.

There also appear to be large differences in the manner that a heat balance is achieved in the upper layers of the two oceans. Godfrey et al. (1998) showed that in the mean the tropical Pacific atmosphere and ocean were closely in balance with a net heat flux into the ocean of order 15-20 Wm^{-2}. Were it not for oceanic circulation, this heating would increase SSTs about 2-3°C/year averaged over a 50-m layer. The North Indian Ocean, on the other hand, is much further away from local thermodynamical equilibrium. Net average heating of the ocean is about 100 Wm^{-2} during the height of the monsoon in July. Thus there must be a substantial southward oceanic cross-equatorial flux of heat during spring and early summer for the SST in the North Indian Ocean to remain within the narrow range actually observed. In fact, the meridional ocean heat transport must be nearly an order of magnitude larger in the Indian Ocean than in the western Pacific.

There are also differences in large-scale atmospheric circulation (see Figure 5). To a large degree, the mean atmospheric circulation over the Pacific is considered a trade wind system converging into the warm western part of the basin with return flow aloft. This forms an along-equator longitudinal Walker Cell. The atmospheric flow over the Indian Ocean, on the other hand, is strongly latitudinal. The Pacific atmosphere responds to a zonal pressure gradient; the Indian Ocean atmosphere to a meridional pressure gradient set up by land-sea temperature differences. In summer, the characteristic cross-equatorial flow concentrates in a lower tropospheric jet stream (the Somalia Jet) in the western Indian Ocean and diverges over the Arabian Sea. Strong convergence occurs in the eastern Indian Ocean, the Bay of Bengal and southeast Asia. A secondary maximum of surface convergence occurs south of the equator. In the upper troposphere, an easterly jet stream extends across equatorial Africa. There is no similar easterly jetstream over the Pacific Ocean. In the boreal winter, an upper tropospheric easterly jet exists but is much weaker than its northern hemisphere counterpart.

Several principal precipitation maxima are apparent in Figure 5c, mostly over water: to the west of India against the Ghat Mountains, over the northern Bay of Bengal, and south of the equator in the winter hemisphere. As we shall see next, the existence of precipitation maxima in the southern hemisphere is a vital clue towards understanding monsoon intraseasonal variability.

In summary, it seems unlikely that all questions about the Indian Ocean and the coupled ocean-atmosphere monsoon circulation monsoon can be answered from the results of other tropical experiments, such as TOGA COARE.

Figure 4: Heat balance of the North Indian Ocean from the intermediate ocean model of McCreary et al. (1993) forced by NCEP/NCAR reanalysis output (Kalnay et al. 1996) showing (a) climatological mean annual cycle of net surface heat flux into the ocean and heat storage north of the equator, and the northward heat flux across the equator, and (b) the interannual variation of the same components for 1984-1990. Units PW (10^{15} W).

Figure 5: Mean near-surface circulation over the monsoon regions for June-August for the 1985-95. Panel (a) shows the mean 850 mb wind field. Broad dashed lines show positions of surface pressure troughs. Panel (b) shows 200 mb wind field. The long-term average of summer precipitation rate (mm month^{-1}) as determined from a combination of MSU precipitation and land surface data is shown in panel (c). ECMWF reanalysis wind data were used

Figure 6: (a) mean precipitable water content (PWC: mm) composited for active and break periods of the monsoon over 15-year period. The criterion used to determine an active and break period is described by Webster et al. (1998). (b) Intraseasonal variability of the monsoon depicted by daily MSU precipitation along 90°E plotted against latitude as a function of time for the boreal summers of 1987 and 1988. 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 south of the equator and,

eventually, enhances winter precipitation over southern Australia. Thirty-nine such events were identified in the 1984-95 period. These events were used to create composites of monsoon intraseasonal events. Data provided by T. Vonder Haar and D. Randel, Colorado State University.

Figure 7: Spatial plots of the composite evolution of anomalous (a) OLR (Wm^{-2}), (b) SST (°C, Reynolds 1988), and (c) 925 mb wind variability (with MSU precipitation rate >8 mm day^{-1} superimposed), relative to the composite monsoon intraseasonal oscillation configured from 39 events in the 1985-95 period. Composite distributions were computed for 15 days on either side of day 0. Annual cycle was removed from all fields. Day "0" is defined as the occurrence of >10 mm/day precipitation (persisting four days) at 90°E. Panels (a)-(c) show composite OLR, SST, and NCEP/NCAR reanalysis 925 mb wind field relative to day 0. Each member of the composites is described in detail in Webster and Tomas (1998). Data files are available from P. J. Webster (pjw@oz.colorado.edu) and R. A. Tomas (tomas@monsoon.colorado.edu).

Figure 8: OLR ($W\ m^{-2}$), SST anomaly (°C) and zonal wind speed (ms^{-1}) averaged over the region 12.5°N-12.5°S and 50°E-100°E for 15 days about day 0 in the composites shown in Fig. 7.

Figure 9: Zonally integrated 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 1987 and 1988, which weak and strong summer monsoons, respectively.

Table 1: Precipitation statistics for 1987 and 1988 at the point (70°E, 15°N) derived from a blended precipitation product using station rain-gauge and satellite microwave sounding data (MSU). Data are plotted as rainfall per pentad (mm/5 day).