DRAFT

The Monsoon as a Self-regulating Coupled Ocean-Atmosphere System

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Abstract

Observational studies have shown that the Asian-Australasian monsoon system exhibits variability over a wide-range of space and time scales. These variations range from intraseasonal (20–40 days), annual, biennial (about 2 years), longer term interannual (3–5 years) and interdecadal. Despite this range of variability, the South Asian monsoon (at least as described by Indian precipitation) exhibits a smaller range of variability during its summer pluvial phase than variability exhibited in other climate systems of the tropics. For example, drought or flood rarely extend to multiple years, with rainfall oscillating biennially from slightly above average to slightly below average precipitation.

We argue that variability of the monsoon is regulated by negative feedbacks between the ocean and the atmosphere. The annual cycle of the heat balance of the Indian Ocean is such that there is an ocean heat transport from the summer hemisphere from the summer hemisphere resulting principally from wind-driven Ekman transport. Given the configuration of the low-level monsoon winds, the Ekman transport is in the opposite sense to the lower tropospheric divergent wind. The cross-equatorial ocean heat transport is large with amplitudes varying between +2 PW (northward) in winter and –2 PW (southward) in summer. Thus, the wind-induced heat transport works to cool the summer hemisphere upper ocean while warming the winter hemisphere. Similar regulation occurs on interannual time scales. For example, during anomalously strong northern hemisphere monsoon summers (typically a La Niña), strong winds induce a stronger than average southward flux of heat. When the monsoon is weak (typically an El Niño), the wind-driven ocean heat flux is reduced. In this manner, the monsoon regulates itself by reducing summer hemisphere sea-surface temperatures during strong monsoon years and increasing it during weak years. In this manner, the monsoon is self regulating.

It is noted, however, that the ocean heat transport theory of monsoon regulation does not necessarily allow heat anomalies to persist from one year to the next. Furthermore, the theory does not include the Indian Ocean Zonal Mode (IOZM: sometinimes referred to as the Indian Ocean dipole) as a dynamic entity. Finally, we develop a more general theory in which the slow dynamics of the IOZM are integral components of a sequence of processes that regulate the monsoon, thus minimizing radical year to year departures of the monsoon from climatology.

1. Introduction

The majority of the population of the planet reside in the monsoon regions. The livelihood and well being of these monsoon societies depends on the variations of the monsoon and the establishment of a symbiotic relationship between agricultural practices and climate. Whereas the summer rains recur each year, they do so with sufficient variability to create periods of relative drought and flood throughout the region. Hence, forecasting the variations of the monsoon from year-to-year, and their impact on agriculture and water resources are a high priority for humankind and are a central issue in the quest for global sustainable development. With accurate forecasts and sufficient lead time for their utilization, the impact of variability of the monsoon on agricultural practices, water management, etc., could be optimized. Yet, despite the critical need for accurate and timely forecasts, our abilities to predict variability have not changed substantially over the last few decades. For example:

(i) For over 100 years the India Meteorological Department has used models based on empirical relationships between monsoon and worldwide climate predictors with moderate success. These endeavors have been extended for both the South Asian region and in other monsoon regions to predict interannual variability in different monsoon regions (see reviews by Hastenrath 1986a, 1986b, 1994). Many of these schemes use measure of El Niño-Southern Oscillation (ENSO) as a major predictor for monsoon variability (e.g., Shukla and Paolina 1983, Rasmussen and Carpenter 1983, Shukla 1987a, b). Whereas there are periods of extremely high association between ENSO and monsoon variability, there are decades where there appears to be little or no association at all (e.g., Torrence and Webster 1999). The association has been particularly weak during the last 15 years.

(ii) Numerical prediction of monsoon seasonal to interannual variability is in its infancy and severely handicapped by the inability of models to simulate either the mean monsoon structure or its year-to-year variability. Model deficiencies have been clearly demonstrated in the Atmospheric Model Intercomparison Program (AMIP: Sperber and Palmer 1996, Gadgil and Sajani 1998) where the results of 39 atmospheric general circulation models were compared.

In summary, a combination of modeling problems and empirical non-stationarity have plagued monsoon prediction. Empirical forecasts have to contend with the spectre of statistical non-stationarity while numerical models currently lack simulation fidelity.

As noted above, most empirical forecast schemes have concentrated on taking advantage of a ENSO-monsoon rainfall relationship. A recent series of empirical studies have shown that the relationships between Indian Ocean SST and Indian rainfall are stronger than portrayed in earlier studies. For example, Sadhuram (1997), Hazzallah and Sadourny (1997) and Clark et al. (2000a) found that correlations as high as +0.8 occurred between equatorial Indian Ocean SSTs in the winter prior to the monsoon wet season and Indian precipitation. A combined Indian Ocean SST index generated by Clark et al. (2000a) has retained an overall correlation of 0.68 for the period 1945 to 1994, after the removal of ENSO influence. Between 1977 and 1995 the correlation has risen to 0.84 again after the removal of ENSO influence. In addition, the so-called Indian Ocean dipole (Webster et al. 1999, Saji et al. 1999, and Yu and

Rienecker 1999, 2000) appears to have strong coupled ocean-atmosphere signatures across the basin, especially in the equinoctial periods, exerting a very significant influence on the autumnal solstitial rainfall over East Africa (Latif et al. 1999, Clark et al. 2000b).

These empirical relationships found between the Indian Ocean SST and monsoon variability are important because they suggest that there may be basin wide relationships that may be, to some extent, are independent of ENSO and, thus, inherent to the Indian Ocean-monsoon system. Some structural independence of basin-wide modes is evident in Figure 1a and b which describe the persistence SST both in space (Figure 1a) and in time along the equator (Figure 1b). The Pacific Ocean possesses a strong persistence minimum in the boreal spring. This is the Pacific Ocean “predictability barrier” of Webster and Yang (1992) and underlines the rapid changes that often occur in the Pacific Ocean during the boreal spring. The pattern in the Indian Ocean is quite different. Strong persistence occurs from the end of the boreal summer until the late spring of the following year. This behavior supports the contention of Meehl (1994a, b, 1997) that there is biennial component in the Indian Ocean SST and monsoon rainfall. Meehl (1994a) also noted that strong Indian monsoons were generally followed by strong North Australian summer monsoon six months later so that the entire Asian-Australian monsoon system followed a biennial pattern. This related behavior can be seen in Figure 2, which plots both the Indian and North Australian rainfall. The South Asian monsoon (at least as described by Indian precipitation) exhibits a smaller range of variability during its summer pluvial phase than variability exhibited by the North Australian monsoon. In fact, the South Asian monsoon is has a variability that is generally smaller than most heavy rainfall regions of the tropics. For example, drought or flood in South Asia rarely extend to multiple years, with rainfall oscillating biennially from slightly above average to slightly below average precipitation.

Given the problems in the predictions of monsoons, and in the light of the recent discoveries regarding the structure of the Indian Ocean and the monsoon, a number of questions need be addressed:

(i) What factors determine the phase and the amplitude of the monsoon annual cycle?

(ii) What factors control the interannual rainfall variability of the South Asian monsoon so that the persistent multiyear anomalies and large excursions from long-term means are rare?

(iii) To what extent is the monsoon a coupled ocean-atmosphere phenomena? For example, do the correlations between Indian Ocean SST and monsoon rainfall indicate the existence of coupled modes?

(iv) What are the relationships, if any, between the bienniality of the monsoon, the Indian Ocean dipole, and longer term interannual variability?

(v) To what degree is variability of the South Asian monsoon independent of outside influences such as ENSO? Is the different character of the persistence of SST shown in Figure 1 indicative of relative independence between the two basins?

The purpose of this study is to make preliminary progress in answering these questions.

The organization of the paper is based on the proposition that in order to understand interannual variability of a phenomenon, it is first necessary to understand its annual cycle. In the next section we will investigate aspects of the annual cycle viewed from a coupled

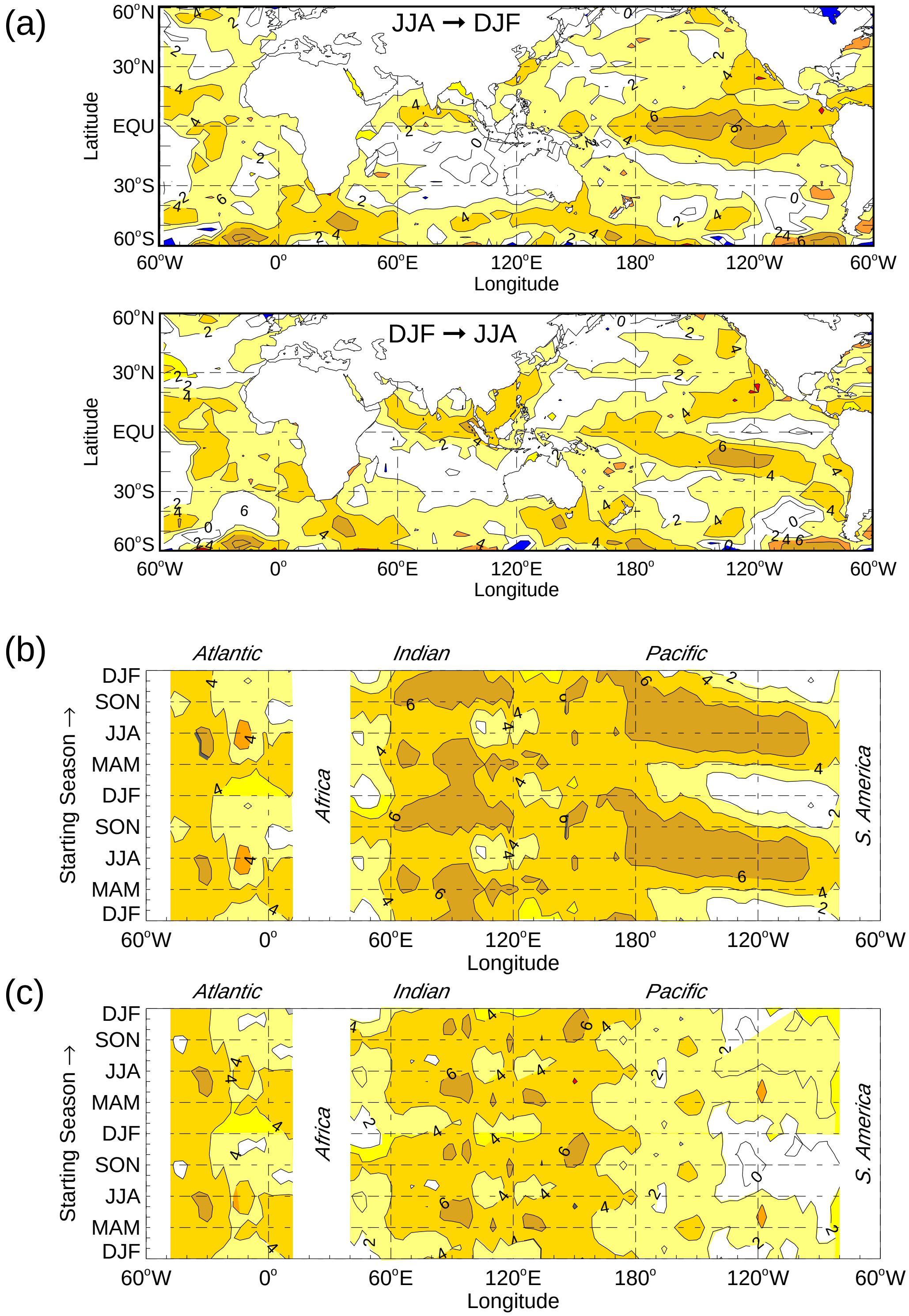

Figure 1 (a) Persistence of the SST over two six month periods: Summer to winter and winter to summer. Extremely strong persistence can be seen in the eastern pacific ocean and the North Indian ocean between summer and winter. However, persistence decreases substantially in the Pacific Ocean between winter and summer. It is maintained, however, in the North Indian ocean. (b) Time sequence of the climatological persistence of SST along the equator for a two year cycle. (c) Persistence with Niño-3 influence removed. (Torrence and Webster 1999).

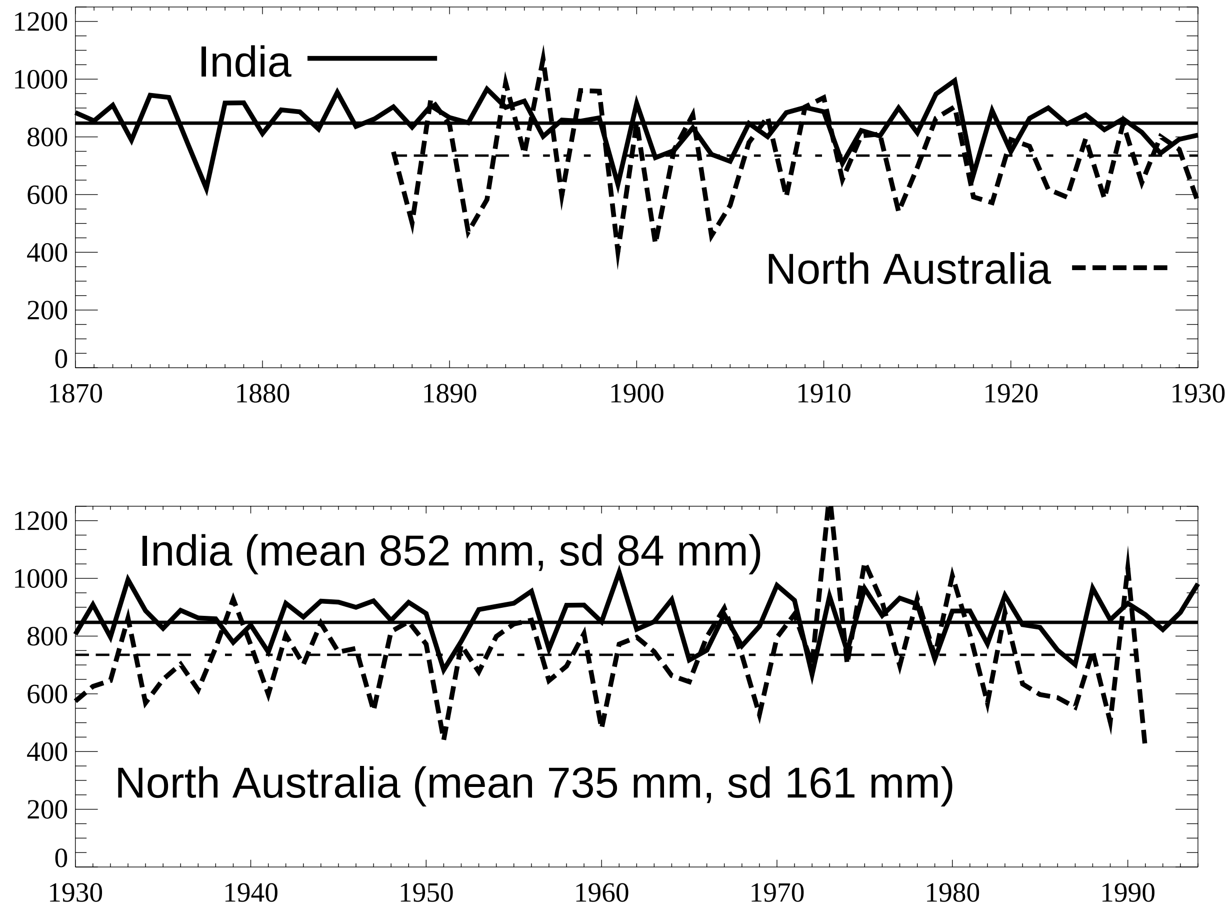

Figure 2 Time-sequence of mean annual precipitation for India and North Australia. All-India rainfall data from Parthasarathy et al. (1992, 1994) and Australian data from Lavery et al. (1997).

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ocean-atmosphere perspective. It will be argued that the ocean and the atmosphere, in concert, act to limit the seasonal extremes of the monsoon. In section 3 the interannual variability of the monsoon system will be considered, again from the view of a coupled system. It will be shown that similar coupled physical processes regulate monsoon interannual variability. In section 4, a more general theory of coupled ocean-atmosphere regulation of the monsoon is proposed. Here, it will be shown that the bienniality of the monsoon and the Indian Ocean dipole are integral parts of a suite of processes and phenomena that regulate the interannual variability of the monsoon.

2. Regulation of the Monsoon Annual Cycle

2.1 The climatological annual cycle

Figure 3 shows the annual cycle of the mean SST, near-surface wind vectors, and outgoing longwave radiation (OLR) for the four seasons. Some pertinent characteristics may be listed briefly as follows:

(i) In general, the warmest SSTs occur in the boreal spring where 29°C surface water covers most of the Indian Ocean north of 10°S. Except in the very far north of the basin, the SST during winter is about 28°C. With the coming of summer and the quickening of the monsoon, the SST of the North Indian Ocean cools, especially in the Arabian Sea. The autumn temperatures are similar to spring but cooler in general by about a degree. In all seasons, the maximum SST gradient occurs south of the equator.

(ii) The lower tropospheric flow of the Asian summer monsoon is much stronger than its wintertime counterpart and possesses a concentrated cross-equatorial flow in the western Indian Ocean compared to the broader, but weaker, reverse flow during the boreal winter.

(iii) During the summer, maxima in convection (low OLR) occurs over south Asia with weaker extensions over equatorial north Africa and in the near-equatorial southern hemisphere. During the boreal winter, an elongated band extends across the Indian Ocean and North Australia culminating in a broad maximum over Indonesia and North Australia. Convection over Asia is located farther poleward than its southern hemisphere counterpart, and the circulations are not symmetrical between the seasons.

Perhaps the most interesting aspect of these climatologies is the lack of variation of the SST between seasons. Most notably, the SST appears to change little during the boreal spring over much of the North Indian Ocean despite weak winds and high insolation. Similarly, the winter SST in the North Indian Ocean is only a degree or so cooler than in the fall or the spring despite significant reductions in surface heating. These aspects of the annual cycle will be addressed in the next section.

2.2 Processes determining the annual cycle of the monsoon

One of the major outstanding problems in climate is how the amplitude and phase of the annual cycle of the tropics, such as the cycle described above for the Indian ocean, adopts the form observed. This problem has been the subject of intensive discussion especially with respect to the western Pacific Ocean following the introduction of the “thermostat hypothesis” by Ramanathan and Collins (1991). In this section, we are particularly interested in how the regulation of the SST distribution in the Indian Ocean because of its close association of the magnitude of the SST with the vigor of the ensuing monsoon (e.g., Clark 2000a).

Determining the surface heat balance of the tropical oceans is difficult, and large differences may occur between estimates. This should not be surprising as the surface heat balances are the relatively small sums of large terms. Furthermore, these large terms are obtained from empirical rules some of which are not known accurately (Godfrey et al. 1995). A result of the TOGA Coupled Ocean-Atmosphere Response Experiment (TOGA COARE: Webster and Lukas, 1992, TOGA COARE, 1994), the surface heat flux into the western Pacific Ocean warm pool is estimated to be between 10 and 20 W m^{-2} (Godfrey et al. 1998).

Unfortunately, the surface heat balance is less well known in the Indian Ocean. Table 1 provides estimates of annual cycle of surface flux for the North Indian Ocean during the boreal spring and early summer using COADS data (Oberhuber 1988). Estimates are listed for the entire North Indian Ocean and for regions north of 10°N and between the equator and 10°N. The daily mean solar radiation into the north Indian Ocean is >200 W m^{-2} during the spring months and 181 W m^{-2} over the entire year compared to an annual average of about 145 W m^{-2} for the western Pacific (Webster et al. 1998). The annual net surface flux averaged over

Flux (W m^{-2})	S	LW	LH	SH	NET	$\frac{dT}{dt}$ (K yr^{-1})
(a) Entire North Indian Ocean						
DJF	125	-57	-103	-4	15	2.2
MAM	223	-49	-82	-2	90	13.4
JJA	190	-39	-117	-0	34	5.1
SON	187	-48	-88	-3	50	7.5
Annual	181	-48	-98	-2	47	7.1
(b) North Indian Ocean equator to 10° N						
DJF	181	-49	-96	-4	24	
MAM	202	-47	-88	-3	65	
JJA	178	-42	-116	-2	17	
SON	186	-46	-87	-4	50	
(c) North Indian Ocean north of 10° N						
DJF	174	-61	-105	-5	3	
MAM	232	-48	-74	-1	112	
JJA	191	-35	-113	0	44	
SON	179	-47	-88	-2	42	

Table 1: The components of the surface heat balance for the North Indian Ocean (a) north of the equator, (b) from the equator to 10° N, and (c) north of 10° N. Unites are W m^{-2}. The net solar radiation, net longwave radiation, latent heat flux, sensible heat flux, and the net flux at the surface are denoted by S, LW, LH, SH, and NET, respectively. Data from COADS (Oberhuber 1987). Heating rates of a 50 m layer for the entire North Indian Ocean are shown in the right hand column (K year^{-1})

the entire North Indian Ocean is about +50 W m^{-2} or at least a factor of two or three larger than in the western Pacific warm pool (Godfrey et al. 1998). Earlier estimates for the Indian Ocean by Hastenrath and Lamb (1978), Hsiung et al. (1989), Oberhuber (1988), and Hastenrath and Greischar (1993) were of order +50 and +70 W m^{-2} on an annual basis. Despite uncertainties, it is clear that over the year there is a much stronger flux of heat into the North Indian Ocean than into the western Pacific Ocean. Furthermore, there is a far larger seasonality in this flux than in the western Pacific Ocean. with maximum values occurring in spring and early summer.

It is a simple matter to estimate the changes in SST resulting just from the observed net fluxes (Table 1). The simplest way is to assume that the net flux is spread through an upper ocean layer of some defined depth for which the heating rate may be written: $dT/dt = -(dF_{net}/dz)/\rho C_p$. Heating rates, assuming a 50 m surface layer, are listed in Table 1. Alternatively, one may use a sophisticated one-dimensional mixed-layer model to make the same computation. Webster et al. (1998) used the model of Kantha and Clayson (1994) finding results quite similar to those listed in Table 1. Both sets of calculations suggest a very different behavior in the evolution

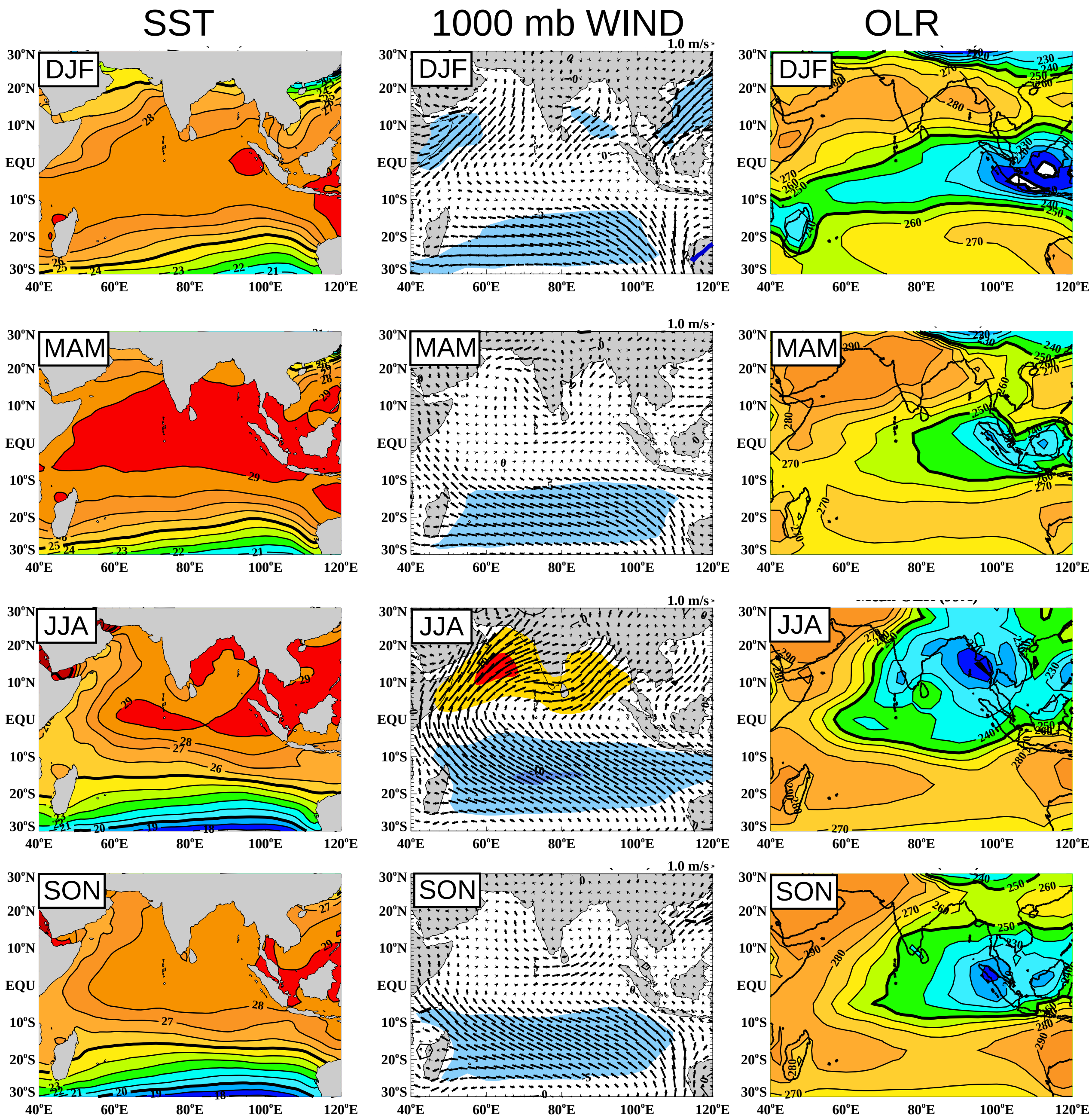

Figure 3 Mean annual climatology of SST (°C), near-surface wind vectors (m s^{-1}), and outgoing longwave radiation (W m^{-2}). Data is plotted for the four seasons March, April and May (MAM), June, July and August (JJA), September, October and November (SON), and December, January and February (DJF).

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of SST than is observed in nature. Figure 3 shows a rather gradual change in SST from one season to another whereas the calculations listed in Table 1 suggest that the Indian Ocean would be continually warming at a large annual rate > 7°C year^{-1}. If this were the case, the observed cyclic equilibrium noted in Figure 3 would not be achieved. On the other hand, using flux values for the Pacific Ocean also suggest a net warming but at a rate of order 1–2 °C year^{-1}.

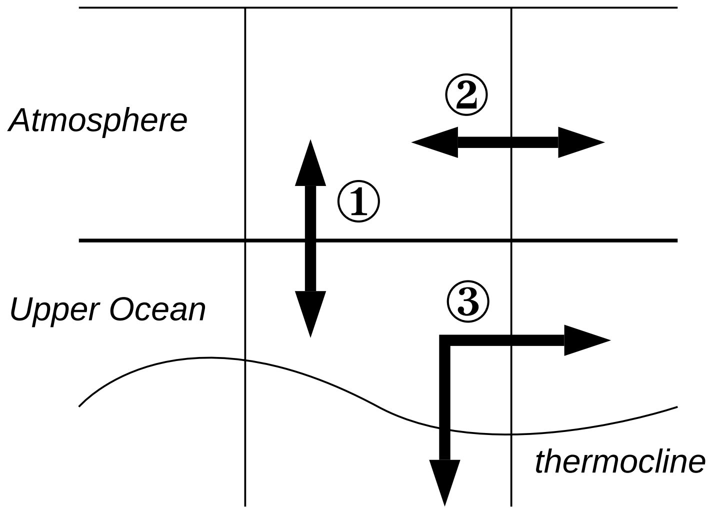

Figure 4 Schematic of the basic combinations of processes considered in SST regulation. Process 1 describes a local heat balance through thermodynamic factors. Process 2 uses heat transports by the atmosphere. Process 3 utilizes ocean transports of heat (Webster et al. 1998).

Figure 4 shows a schematic diagram of the basic combination of processes used in studies of SST regulation. The first combination of processes (Set 1 in Figure 4) is *regulation through local thermodynamic adjustment*. Such a state was defined by Ramanathan and Collins (1991) as the "natural thermostat" in which an SST warming produces cloud, which, in turn, reduces surface solar radiation flux and thus regulates SST. The second combination of processes (Sets 1 and 2) describes *regulation by a combination of atmospheric dynamics and transports in addition to negative radiational feedbacks*. Here an invigorated atmospheric circulations assumed to respond to increased SST gradients which enhance evaporative cooling of the ocean surface (Fu et al., 1992, Stephens and Slingo, 1992, Wallace 1992, Hartmann and Michelson, 1993). The third combination of processes (Sets 1, 2 and 3) is *horizontal or vertical advection of heat or changes in the storage of heat by oceanic dynamics in addition to local thermodynamic adjustment and cooling by an invigorated atmosphere.* Within this third combination, the transport of heat by the ocean is important.

It is clear that the Ramanathan and Collins (1991) mechanism cannot regulate SST in the Indian Ocean. The warmest SSTs on the planet during spring and early summer are not associated with convection (Figure 3) so that there is no cloud shielding to mitigate the solar radiation flux at the surface. Furthermore, the second class of theories for regulation of the SST in the Pacific Ocean (Fu et al., 1992, Stephens and Slingo, 1992, Wallace, 1992, Hartmann and Michelson, 1993) also cannot solve the problem; winds remain light during the boreal spring and early summer, and evaporation is relatively low (Figure 3).

It is clear that the ocean must play an extremely important dynamic role in achieving the cyclic equilibrium of the mean annual cycle. As summarized by Godfrey et al. (1995, p. 12):

> *"... on an annual average there is positive heat flux into the Indian Ocean, nearly everywhere north of 15°S. The integral of the net heat influx into the Indian Ocean over*

the area north of 15° S ranges between 0.5–1.0 x 10^{15} W, depending on the climatology. Thus, on the annual mean, there must be a net inflow of cold water (into the North Indian Ocean), and a corresponding removal of warmed water, to carry this heat influx southward, out of the tropical Indian Ocean. ...”

Oceanic heat transports of heat represent the only means that allows the large net annual surface heating of the North Indian Ocean to be removed without raising the SST substantially. It should be noted that Godfrey et al. (1995) were referring to annually averaged ocean heat transports. In the next section we will show that the annual average is made up of seasonal swings with amplitudes that vary between ±2 PW.

2.3 Role of ocean dynamics in the annual heat balance of the Indian Ocean

To understand the role of dynamical transports by the ocean, Loschnigg and Webster (1996) used the McCreary et al. (1993) dynamic upper ocean model, which incorporates full upper ocean dynamics and thermodynamics. Although there is little subsurface data in the Indian Ocean, the model has been shown to replicate the surface structure of the Indian Ocean, as well as its annual cycle, when the ocean model is run in stand-alone mode with prescribed atmospheric forcing (Loschnigg and Webster 2000). The instantaneous northward heat flux between two positions along a line of constant latitude is defined as

$$F_Q = \rho_w c_w \int_{\phi} \int_{z} H_i v_i T_i dx dz \qquad (1)$$

where H_i, T_i , and v_i are the depth, temperature, and meridional velocity component of the i^{th} layer of the model. The storage in a volume is defined by

$$S_Q = \rho_w C_w \int_{x} \int_{y} \int_{z} H_i T_i dx dx dz \qquad (2)$$

Figure 5a shows the annual cycle of meridional heat transport and heat storage changes in the Indian Ocean forced by the annual cycle of climatological winds and heating. The net surface flux into the North Indian Ocean appears to have a semiannual variation. This is due to the combination of net solar heating and cooling by evaporation. Evaporative cooling is largest in the summer associated with the strong monsoon winds. There is also a second maximum in winter associated with the winter monsoon. Solar heating also has two minima: in summer where cloudiness has increased with the onset of the monsoon and in winter. Together these two complicated heating mechanisms provide the double maximum evident in Figure 5a. The major components of heat flow are between the transport and storage terms. A very strong southward flux of heat is evident during the spring and early summer, and a reverse flux is evident during the winter, when the north Indian Ocean is losing considerable amounts of heat by both evaporation, and vertical turbulence mixing which entrains colder water from below the thermocline. The net heat flux across the equator is made up of opposing flows in the upper and lower layer of the model and may be thought of as a seasonally reversing meridional ocean circulation (McCreary et al., 1993). The magnitude of the net southward flux across the equator during the spring and summer months more than makes up for the excess surface flux into the

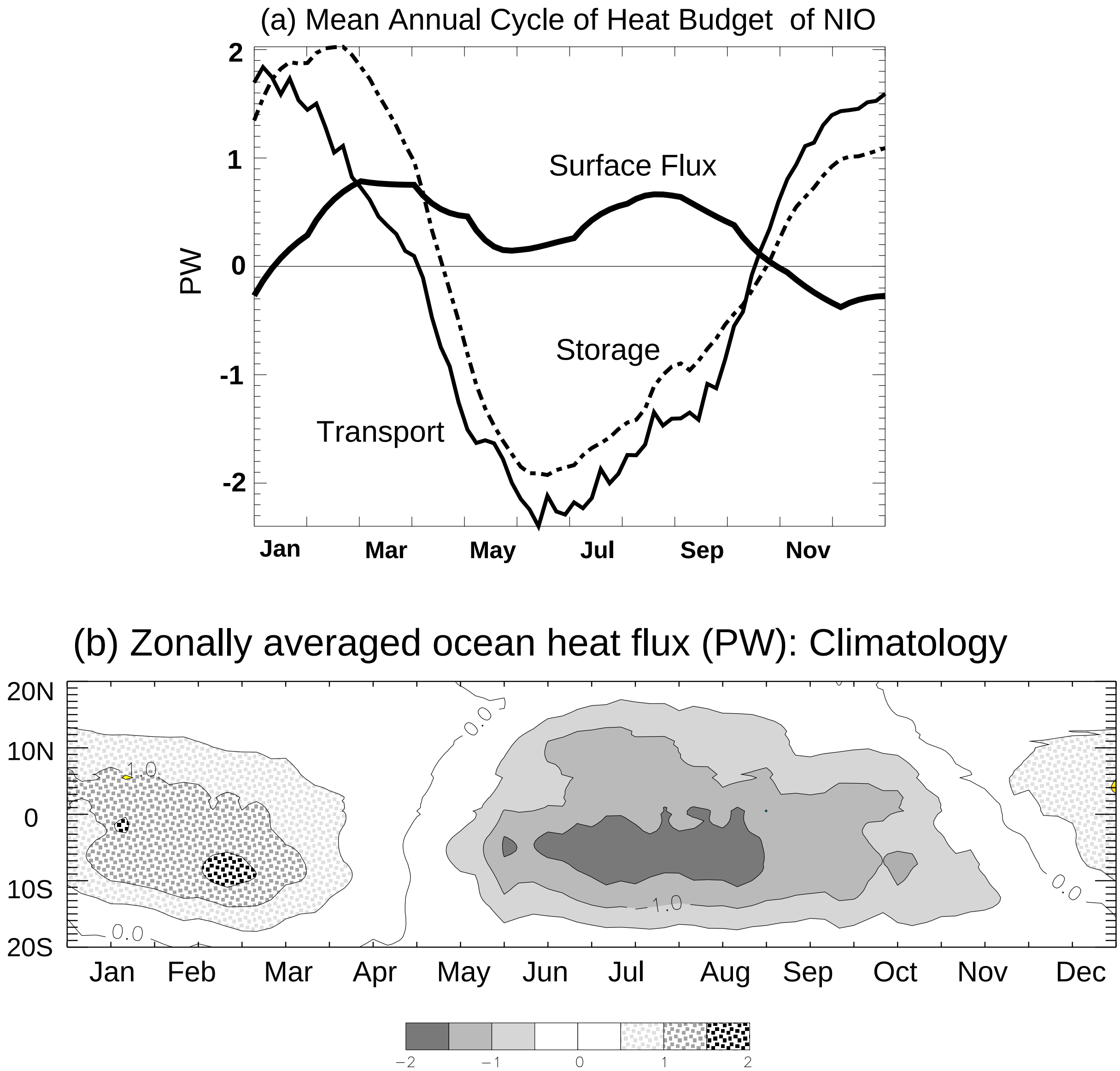

Figure 5 (a) Annual cycle of meridional heat transport across the equator, changes in heat storage, and net heat flux into the North Indian Ocean. Positive surface heat flux indicates a heating of the ocean. Positive heat transport is to the north. (b) Annual cycle of meridional ocean heat flux as a function of latitude and longitude. Units are PW. (Loschnigg and Webster 2000).

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north Indian Ocean, with peak summer magnitudes of meridional heat transport reaching values of 2 PW. Whereas the changes in heat storage had been anticipated by earlier studies (e.g., Vonder Haar and Oort, 1973), the strong role of advection was not considered.

Figure 5b displays a latitude-time section of the annual cycle of the climatological meridional oceanic heat flux averaged across the basin. The year is divided into a period of northward heat flux in winter and spring and a southward, slightly stronger, heat flux between late spring and early fall. Maximum transport occurs at all seasons close to 10°S neat the zone of maximum SST gradient.

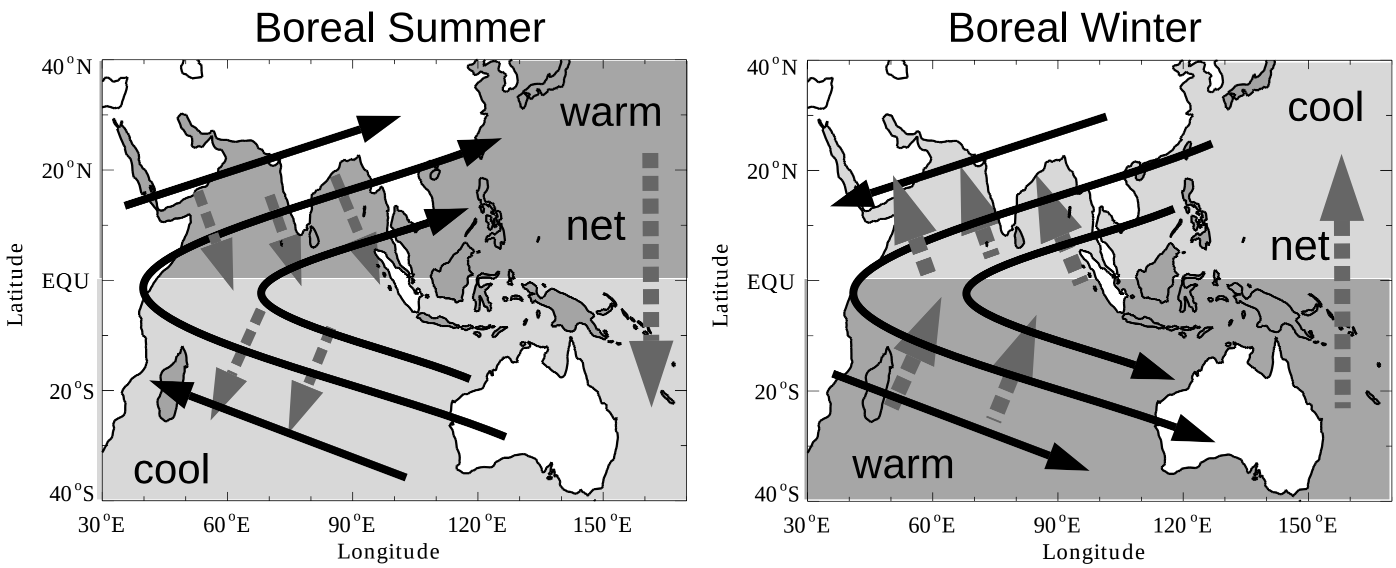


Figure 6 A regulatory model of the annual cycle of the Indian Ocean-monsoon system depicted for (a) summer (June-September), and (b) winter (December-February). Curved black arrows denote the wind forced by the large-scale differential heating denoted by "warm" and "cool". The small grey arrows are the Ekman transports forced by the winds. The large vertical black arrow to the right of the panels show the net ocean heat transport which reverses between boreal summer and winter. The net effect of the combined wind forced ocean circulation is to transport heat to the winter hemisphere thus modulating the SST differences between the hemispheres.

2.4 Regulation of the annual cycle of the monsoon: an ocean-atmosphere feedback system

From the calculations above, it is clear that without ocean transport across the equator and changes in the heat storage of the north Indian Ocean, the cross-equatorial buoyancy gradient during early summer would be very large. Yet the processes that accomplish the cross-equatorial transport of heat in the ocean are essentially wind driven (Mohanty et al., 1996; McCreary et al., 1993; Godfrey et al., 1995). In turn, the atmospheric circulation is driven by surface fluxes and heating gradients associated with the buoyancy gradient and atmosphere-land interaction. Thus, the annual cycle in the Indian Ocean is a coupled phenomenon resulting from ocean-land-atmosphere interactions and balanced, to a large extent, by cross-equatorial oceanic transports.

The form of interaction between the atmospheric monsoon flow and the ocean transport results from a coupled ocean-atmosphere feedback. A critical element of the meridional heat transport is that it is accomplished to a large degree by Ekman processes, a point that has been noted in previous work. For example, Levitus (1987) was the first to note that meridional Ekman transports are southward in the Indian Ocean on both sides of the equator and in both the boreal summer and the annual mean. Wacongne and Pacanowski (1996) describe a process where the upwelled water within the western boundary currents are transported across the equator via the action of Ekman transports. Loschnigg and Webster (2000) showed that the seasonal reversal of the Ekman transports are central to the maintenance of the heat balance of the Indian Ocean.

Figure 6 shows a schematic of the monsoon system. The two panels represent summer and winter with relatively warm water (shaded) in the northern basin during summer and in the southern basin during winter. Surface winds similar to those shown Figure 3 are superimposed on the figure. In each season there is a strong flow from the winter to the summer hemisphere with a characteristic monsoon “swirl”. The divergent part of the wind field (not shown: see Webster et al. 1998) is also from the winter to the summer hemisphere. The broad gray arrows represent the ocean Ekman transports associated with the surface wind forcing. Irrespective of the season, the Ekman transports are from the summer to the winter hemisphere. The total effect of the feedback is to cool the summer hemisphere and warm the winter hemisphere, thus reducing the SST gradient between the summer and winter hemispheres. These transports are sufficiently large to be responsible for reducing the heating of the upper layers of the summer hemisphere to values less than shown in Table 2. In the manner described in Figure 6, the amplitudes of the seasonal cycle of the monsoon are modulated through the negative feedbacks between the ocean and the atmosphere.

3. Interannual Variability of the Monsoon

Figure 7 shows the difference of the near-surface wind fields between strong and weak monsoon years, defined in terms of the monsoon index shear index defined by Webster and Yang (1992). Using the index criteria, 1979, 1983, 1987, and 1992 were categorized as “weak” monsoon years and 1980, 1981, 1984, 1985, and 1994 as “strong” monsoon years. In agreement with earlier analyses of Webster and Yang (1992) and Webster et al. (1998), the difference fields show a tendency for an increase in westerlies across the Indian Ocean region during strong years or, alternatively, an increase in low-level easterlies during weak years. There are important local manifestations of these difference fields. In strong monsoon years the southwesterlies towards East Africa are enhanced while, at the same time, there is an increase in onshore flow towards Sumatra. In weak monsoon years there is the reversal of the wind vectors. The form of the difference fields shown in Figure 7 will turn out to be of critical importance in understanding the manner in which the monsoon is regulated. This is because the changes in winds between the monsoon extremes occur in regions where major upwelling occurs.

3.1 Modes of interannual variability in the monsoon

Statistical analyses indicate specific bands of monsoon variability (e.g. Webster et al. 1998). Here we consider major frequency bands longer than the annual cycle. These are the biennial period (first discussed by Yasunari 1987, 1991, Rasmusson et al. 1990, and Barnett 1991), and the multiyear variability that appears on ENSO time scales. Within the biennial period we include the newly discovered Indian Ocean dipole (Webster et al. 1999, Saji et al. 1999, Yu and Rienecker 1999, 2000).

3.1.1 <u>Biennial variability</u>:

The interannual variability of monsoon rainfall over India and Indonesian-Australian shows a biennial variability during certain periods of the data record (Figure 2). It is sufficiently strong and spatially pervasive during these periods to show prominent peaks in the 2-3–year period range, constituting a biennial oscillation in the rainfall of Indonesia (Yasunari and Suppiah, 1988) and east Asia (Tian and Yasunari, 1992; Shen and Lau, 1995) as well as in Indian rainfall (Mooley and Parthasarathy, 1984). The phenomena, referred to as the tropospheric biennial

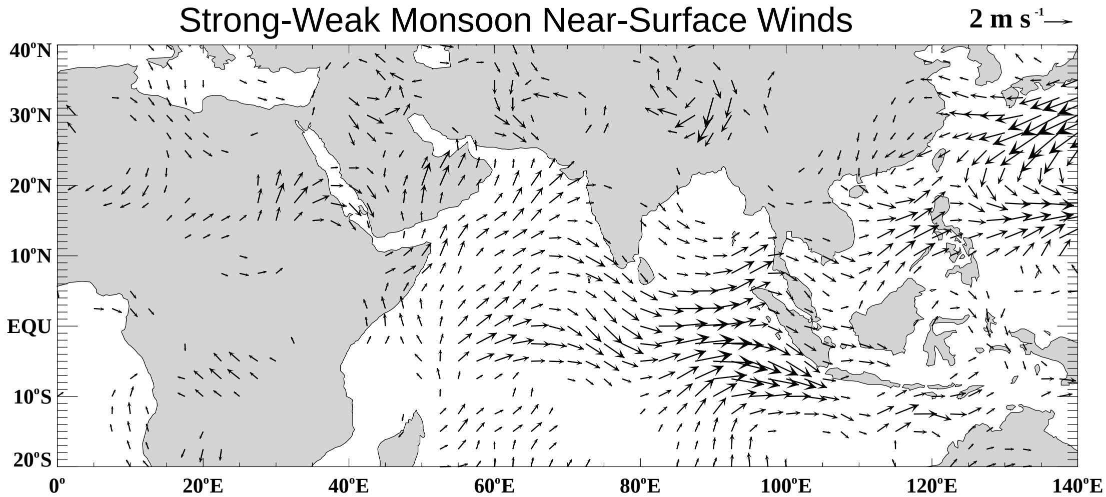


Figure 7 Difference in the surface winds in JJA between strong and weak monsoon years. Strong and weak monsoon years were determined using the monsoon shear index of Webster and Yang (1992) which specified 1979, 1983, 1987, and 1992 as "weak" years and 1980, 1981, 1984, 1985, and 1994 as "strong" monsoon years.

oscillation (TBO) in order to avoid confusion with the stratospheric quasi-biennial oscillation (QBO), appears to be a fundamental characteristic of Asian-Australian monsoon rainfall.

The rainfall TBO, apparent in Figure 2, appears as part of the coupled ocean-atmosphere system of the monsoon regions, increasing rainfall in one summer and decreasing it in the next. The TBO also possesses a characteristic spatial structure and seasonality (Rasmusson et al. 1990, Ropelewski et al., 1992). Meehl (1994a) stratified ocean and atmospheric data relative to strong and weak Asian monsoons. He found specific spatial patterns of the TBO with a distinct seasonal sequencing. Anomalies in convection and SST migrate from south Asia toward the southeast into the western Pacific of the southern hemisphere following the seasons. Lower-tropospheric wind fields associated with the TBO in the SST fields possesses an out-of-phase relation between the Indian Ocean and the Pacific Ocean basins (Ropelewski et al., 1992) with an eastward phase propagation from the Indian Ocean toward the Pacific Ocean (Yasunari, 1985; Kutsuwada, 1988; Rasmusson et al., 1990; Ropelewski et al., 1992; Shen and Lau, 1995) providing possible links between monsoon variability and low-frequency processes in the Pacific Ocean (Yasunari and Seki, 1992, Clarke et al. 1998).

Explanations for the TBO fall into two main groups:

(i) The TBO as results from feedbacks in the seasonal cycle of the atmosphere-ocean interaction in the warm water pool region, especially in the western Pacific Ocean. For example, Clarke et al. (1998) suggests the oscillation may be produced by an air-sea interaction instability involving the mean seasonal wind cycle and evaporation. They argue that similar instabilities are not possible in the Indian Ocean and that Indian Ocean oscillations found there are the result of Pacific instabilities.

(ii) Biennial oscillations occur as a natural variability of the monsoon coupled ocean-atmosphere-land monsoon system. As distinct from the views expressed in (i), the source of the biennial oscillation is thought to reside in the Indian Ocean. Nicholls

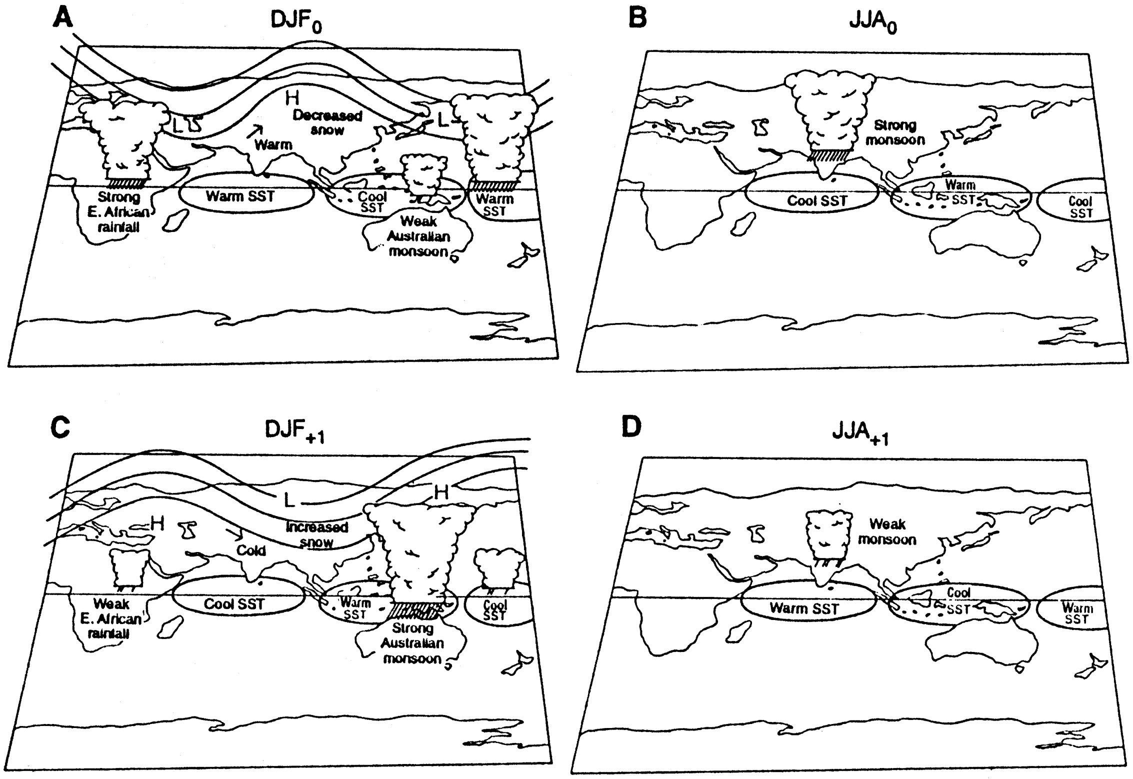


Figure 8 Schematic diagram of the Meehl biennial mechanism. Diagram shows the evolution before a strong monsoon (A), through a strong monsoon season (B), to the northern winter after the strong monsoon before the weak monsoon (C), to the following weak monsoon (D). From Meehl (1997).

(1983) noted a seasonal change in the feedback between the wind field and surface pressure. In the monsoon westerly (wet) season the wind speed anomaly is negatively correlated to the pressure anomaly, while in the easterly (dry) season it is positively correlated. The wind speed anomaly, on the other hand, is negatively correlated to the SST change throughout the year through physical processes such as evaporation and mixing of the surface ocean layer. Nicholls suggested that a simple combination of these two feedbacks in the course of the seasonal cycle induces an anomalous biennial oscillation. Meehl (1994a, b, 1997) substantiated Nicholls' hypothesis but focused on the memory of oceanic mixed layer. That is, when large-scale convection over the warm water pool region, associated with seasonal migration of ITCZ and the monsoon, is stronger (weaker), the SST will eventually become anomalously low (high) through the coupling processes listed above. The anomalous state of the SST, thus produced, would be maintained through the following dry season and even to the next wet season. In turn, the SST anomaly produces weaker (stronger) convection. In

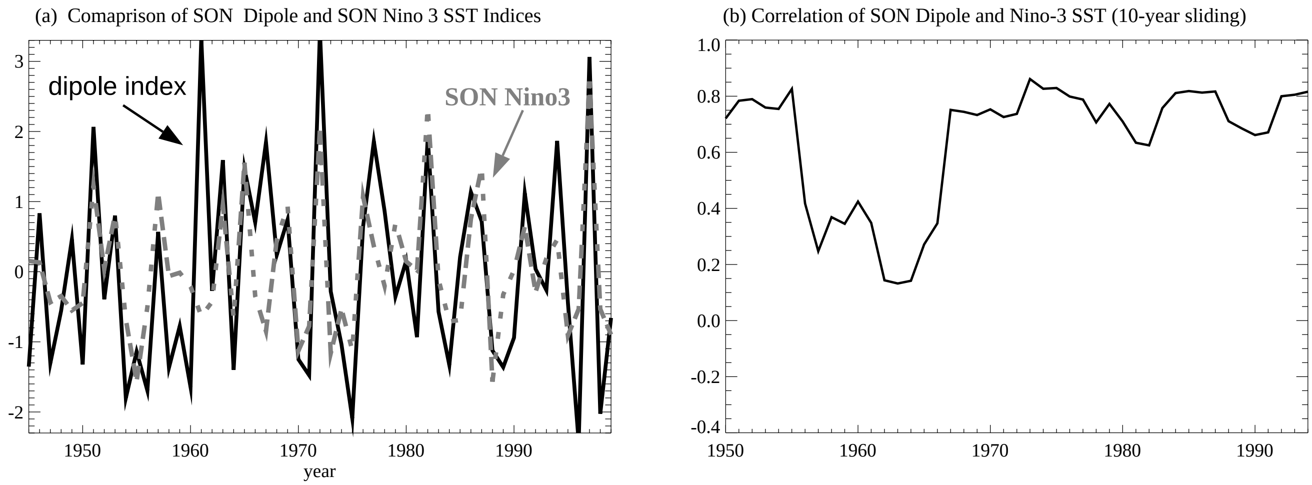


Figure 9 Statistics of the Indian Ocean dipole. (a) Time section of the dipole index. Index is defined as the SST difference in the areas 5°S-10°N and west of 60°E, and 10°S-5°N and east of 100°E for the October-November period. The index, defined by Clark et al. (2000b) is similar to that defined by Saji et al. (2000). (b) 10–year running correlations between the dipole index and the October-November Niño-3 SST.

this class of hypotheses the ocean–atmosphere interaction over the warm water pool appears to be of paramount importance.

A schematic diagram of the four phases of Meehl's biennial monsoon system (Meehl 1994a) is shown in Figure 8. The first panel depicts the winter season prior to the first monsoon season showing anomalously warm SST in the central and western Indian Ocean and cooler SSTs in the eastern Indian Ocean and the Indonesian seas. Anomalously warm SSTs in the Indian Ocean herald a stronger monsoon supposedly by a heightened surface hydrological cycle (panel 2). This is consistent with the empirical results of Clark et al. (2000a). A stronger monsoon is accompanied by stronger wind mixing and evaporation which leads to cooler SSTs in the central and eastern Indian Ocean. A reversal of the east-west SST gradient produces a stronger North Australian monsoon (panel 3). In turn, the colder than normal Indian Ocean leads to a weak Indian summer monsoon (panel 4). The theory also notes that a strong South Asian monsoon is preceded by a strong East African monsoon and a weak South Asian monsoon by a weak East African monsoon. Presumably, the oscillation of the East African monsoon is associated with the change of the longitudinal SST gradient.

The sequence of SST change shown in Figure 8 follows observations quite closely including the oscillation of precipitation from year to year (Figure 2). Whereas it is very clear that the oscillation of the SST in the Indian Ocean is indelibly tied to the variability of the monsoon rains, there are two problems with the theory. First, there is no satisfactory explanation for the change in the SST gradient along the equator. Second, it is difficult to account for the persistence of SST anomalies for the 9 months between the end of one summer season and the start of the next. Clearly, this cannot be accomplished by thermodynamical processes alone. In fact, the e–folding time of 50 m mixed layer with a 1 K anomaly at 303 K is between 40–60 days. There must be dynamical ocean processes at work in the Indian Ocean that increase the persistence of the anomalies.

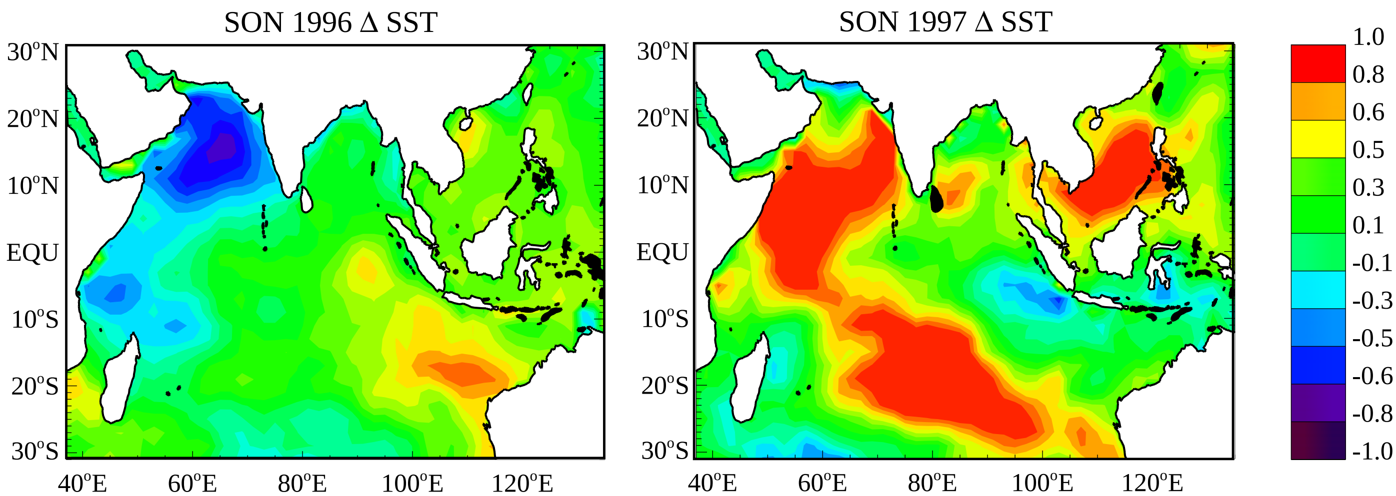


Figure 10 The anomalous SST fields for the boreal autumn (SON) over the Indian Ocean for a positive dipole year (1997) and a negative dipole year (1998).

3.1.2 The Indian Ocean Dipole:

Between July, 1997, and the early summer of 1998, the strongest seasonal SST anomalies ever recorded occurred in the Indian Ocean. During this period, the equatorial gradients of both SST and sea surface height (SSH) reversed with cooler surface temperature in the eastern basin and warmer in the west. These anomalies occurred in conjunction with strong easterly wind anomalies across the equatorial Indian Ocean. The anomalies in SST and SSH persisted for almost a year and coincided with the 1997-1998 El Niño. This event has received considerable attention (e.g., Webster et al. 1999, Saji et al. 1999, Yu and Rienecker 1999, 2000). Following the summer of 1998, the pattern changed polarity with anomalously warm water occupying the eastern Indian Ocean with cold water in the west. In 1996, the SST distribution was very similar to that found in 1998. In fact, over the last few years there has been a sequence of positive and negative dipole events in the Indian Ocean.

Analysis of Indian Ocean SST data reveals that the 1997–98 and 1961 events were members of a longitudinal SST anomaly oscillation in the tropical Indian Ocean. Reverdin et al. (1986) and Kapala et al. (1994) refer specifically to an event of similar magnitude in 1961. Figure 9a plots a September-November (SON) dipole index defined by Clark et al. (2000b) which is very similar to the index defined by Saji et al. (1999). The index oscillates between positive (anomalously warm west, cold east) and negative (anomalously cold west, warm east) values. Figure 10 compares mean seasonal anomalous SON SST distributions for the 1997 positive dipole event with the negative dipole occurring the previous year. In both years the anomalous SST difference exceeded 2°C.

Because of the relative abundance of data and the magnitude of the event, the 1997–1998 positive event was studied extensively. There are a number of points that have emerged from these studies which are pertinent to subsequent discussion:

(i) The association of the Indian Ocean dipole with the El Niño of 1997–1998, or with the ENSO phenomena in general, remains unclear. Saji et al. (1998) for example, states that there is little or no relationship between ENSO and the dipole. However, there are moderate correlations between the dipole index and the Niño 3 SST index which may be seen form the 10–year sliding correlation between the SON dipole and the Niño 3 SST (Figure 9b). An overall correlation of about 0.55 exists. Based on correlations between ENSO and the dipole, Reason et al. (2000) argue that the dipole is simply an extension of the ENSO influence in the Indian Ocean.

(ii) It is clear that ENSO variability may be associated with some dipole events, but it may not be involved in all. Figure 9b shows significantly reduced correlation in the 1950s and 1960s. Furthermore, during the 1997–1998 the climate patterns around the Indian Ocean rim were very different from those normally associated with El Niño. Although the El Niño was the strongest in the century, monsoon rains were normal in South Asia and North Australia when drought may have been expected. Instead of slightly increased rainfall in East Africa, the rainfall was the largest positive anomaly of the century, even larger than the 1961 excursion. Arguably, in 1997–1998, climate anomalies around the basin could be more associated than with the anomalous conditions in the Indian Ocean than in the Pacific. Finally, large dipole events have occurred that are not matched by ENSO extrema, most notably the 1961 event documented by Reverdin et al. (1986) and Kapala et al. (1994).

(iii) Saji et al. (1999) note that the signature of the dipole commences in the late summer. Such a commencement occurs for both the positive and the negative phases of the dipole.

(iv) Webster et al. (1999) suggest that the dipole is a coupled ocean-atmosphere instability that is essentially self-maintaining. Analyses suggest that the initial cooling of the eastern Indian Ocean sets up an east to west SST gradient that drives near-equatorial anomalous easterlies. In turn, these winds change the SSH to tilt upwards to the west. Relaxation of sea surface height anomalies, in the form of westwardly propagating and downwelling ocean Rossby waves, depress the thermocline in the west and enhance the warming of the western Indian Ocean. The slow propagation of these modes (1–2 m s^{-1}), and the manner in which they maintain the warm water by deepening the thermocline, assures that the dipole is a slowly evolving phenomena.

3.1.3 Interannual variability and monsoon-ENSO relationships:

Table 2 shows the long term relationships between Indian and North Australian rainfall and ENSO variability. It is clear from the statistical analyses displayed in the table that the monsoon undergoes oscillations in conjunction with the ENSO cycle. Specifically, when the Pacific Ocean SST is anomalously warm, the Indian rainfall is often diminished in the subsequent year. The correlation of the AIRI and the SOI over the entire period is –0.5 (Torrence and Webster, 1998). Shukla and Paolina (1983) were able to show that there was a significant relationship between drought and ENSO. In fact, all El Niño years in the Pacific Ocean were followed by drought years in the Indian region. Not all drought years were El Niño years, but out of the total of 22 El Niño years between 1870 and 1991, only two were associated with above average rainfall. La Niña events were only associated with abundant rainy seasons, and only two were associated with a monsoon with deficient rainfall. A large number of wet years were not associated with cold events, just as many drought years were not associated

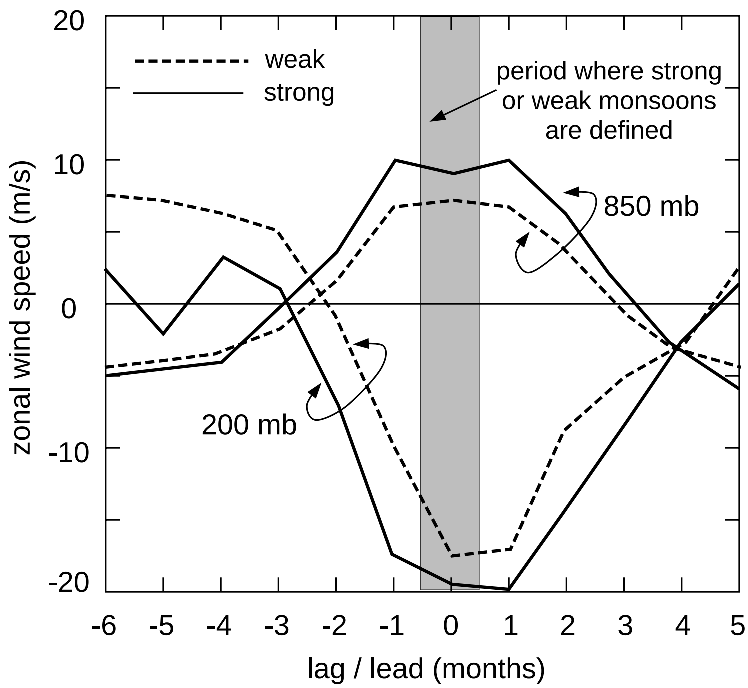


Figure 11 Variation of the anomalous 850 mb and 200 mb zonal wind components relative to strong and weak monsoon years defined at month zero (July). The curves indicate a different circulation structure in the Indian Ocean region prior to strong and weak summer monsoons up to two seasons ahead. After Webster and Yang (1992).

	All India Summer			North Australian Summer		
Rainfall	Total	El Niño	La Niña	Total	El Niño	La Niña
Below Average	53	24	2	49	20	4
Above Average	71	4	19	58	5	17
Deficient	22	11	2	18	9	0
Heavy	18	0	7	17	2	5

Table 2: Relationship between the phases of ENSO (El Niño and La Niña) and the variability of the Indian summer monsoon (June-August) and the North Australian summer monsoon (December-February). Rainfall in these two monsoon regimes are represented by indices of total seasonal rainfall. “Deficient” and “heavy” rainfall seasons refer to variability less than, or greater than, one standard deviation, respectfully.

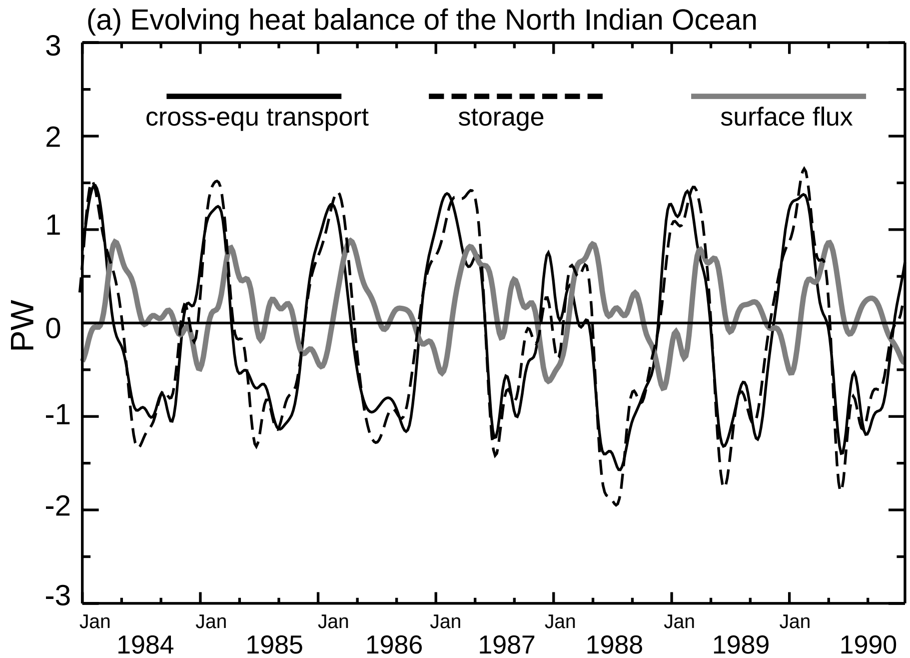
with warm events. Although the relationship is far from perfect, it is clear that the monsoon and ENSO are related in some fundamental manner. Similar relationships occur between the north Australian rainfall and ENSO.

The statistical relationships discussed above indicate a common co-occurrence of monsoon variability and ENSO extremes. But at what part of the annual cycle do anomalously strong and weak monsoon seasons emerge? To help answer this question, mean monthly circulation fields were composited for the weak and strong monsoon years. The upper and lower tropospheric zonal wind fields in the south Asian sector for the composite annual cycle of the strong and weak monsoons are shown in Figure 11. At the time of the summer monsoon when the anomalous monsoon is defined, both the low-level westerlies and the upper level easterlies are considerably stronger during strong monsoon years than during weak years as depicted in Figure 7. But what is very striking is that the anomalous signal of upper level easterlies during strong years extends back until the previous winter with 5–6 m s^{-1} less westerly during strong years. However, in the lower troposphere the difference between strong and weak years occurs only in the late spring and summer. Thus there is a suggestion that the anomalies signify external influences from a broader scale into the monsoon system. This surmise is supported by what is known about tropical convective regions. Generally, enhanced upper tropospheric winds will be accompanied by enhanced lower tropospheric flow of the opposite sign. But this is clearly not the case prior to the strong monsoon, suggesting that the modulation of the upper troposphere probably results from remote influences. This is troublesome in relation to ENSO-monsoon relationships. The ENSO cycle is phase-locked with the annual cycle and an El Niño, for example, develops in the boreal spring. Figure 11 suggests that there are precursors to anomalous monsoons that develop before the growth of an ENSO event. Whether or not this result is of problems in the definition of an anomalous monsoon or a result from which insight may be gained regarding monsoon variability is not known.

3.2 Interannual modes in ocean heat transport

If the monsoon is indeed a coupled ocean-atmosphere system then there should be oceanic variability occurring on the same time scales as atmospheric variability. To test this hypothesis, Loschnigg and Webster (2000) forced the McCreary et al. (1993) dynamic ocean model with NCEP winds. Radiative forcing was obtained from Bishop and Rossow (1991). In the results discussed here, 5–day average fields were used to force the model.

Figure 12a shows time sections of the mean monthly cross-equatorial heat transports, storage change and net heat flux into the Indian Ocean north of the equator for the period 1984–1990 (Loschnigg and Webster 2000). The period was chosen because it included the years 1987 and 1988 which corresponded to an intense El Niño and a weak monsoon followed by a La Niña and a strong monsoon, respectively. Figure 12b shows a sequence of annual averages of these fields. Averaged over the entire 7 years, the net heating of the North Indian Ocean is balanced to a large degree by the southward heat transport across the equator. Storage change averages roughly to zero through the period. However, these long term balances do not occur during 1987 and 1988. During 1987, the net surface heat flux is about 30% higher than the 7–year average, presumably because of the weaker than average winds (less evaporation and mixing) and greater surface radiational flux because of less cloud and convection. During 1997 the rises in storage indicates an increase in the mean temperature of the upper North Indian Ocean. Also, the southward heat transport is less than average which is consistent with lighter winds and less southward Ekman transports. In 1988, a strong monsoon year, the net surface heat flux is lower than the 7–year average by 30% as a result of greater evaporation and by the reduced solar heating. Cross-equatorial heat transports are almost double the previous year again reflecting the impact of the stronger than average winds. Storage decreases indicates a net cooling of the North Indian Ocean. The results of this experiment indicate clear differences in ocean behavior between strong and weak monsoon years.



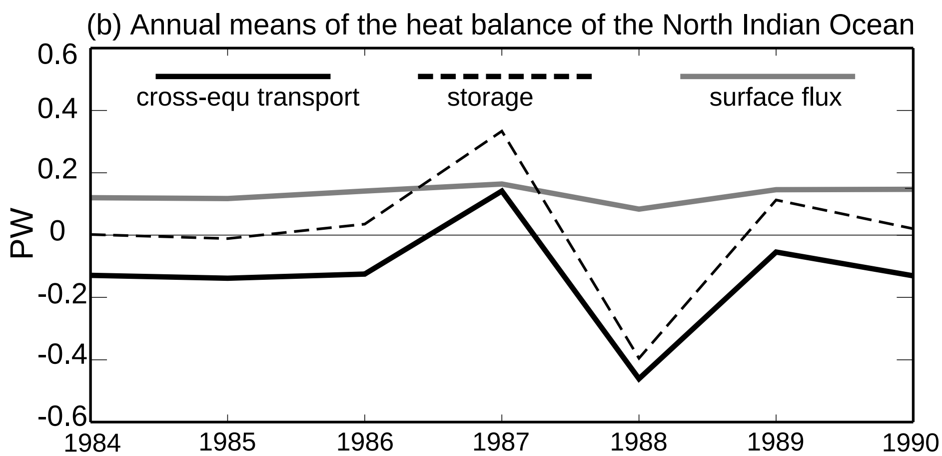


Figure 12 (a) Evolution of the components of the heat balance of the North Indian Ocean. mean monthly values are plotted from January 1994 through December 1990. (b) Annual averages of the components of the heat balance of the Indian Ocean basin north of the equator for model simulations of the period 1984-1990 showing the cross-equatorial heat transport, and the heat storage and surface flux into the North Indian Ocean. (c) Time-latitude section of the total northward transport for 1987. This year was a weak monsoon year coinciding with a strong El Niño. (d) Same as (c) but for 1998 which was a strong monsoon year and also the year of a strong La Niña. Model was forced by 5–day average NCEP surface winds (Kalnay et al. 1996) and ISCCP fluxes (Bishop and Rossow 1991).

Figures 12c and d show details of the meridional heat transport as a function of time and latitude for 1987 and 1988. The format is the same as in Figure 5b except 5–day average forcing is used. The general annual cycle of heat transport is apparent in both years with northward transport during the winter and early spring and southward transport during summer. Both years also have strong intraseasonal variability suggesting a response by the ocean to higher frequency forcing. However, there are also significant differences. The stronger southward transport in 1988 noted in Figure 12a can be seen clearly in the enhanced transports during the summer. However, in 1987, the weaker southward transports are made up by a combination of stronger northward heat transport in winter and early spring and weaker southward transport during summer.

Cherikova and Webster (2000) have repeated the interannual integrations of the ocean model using the 40–year period of the NCEP reanalysis data. As independent surface radiative flux data is not available for the entire period, the NCEP reanalysis data is used for the radiative heating fields as well as for surface wind forcing. Figure 13a shows a 40–year time section of the annually averaged anomalous meridional ocean heat transports plotted between 30°S and 25°N. The long-term 40–year mean has been removed leaving anomaly fields. Three specific features can be noticed:

(i) The magnitude of the variability from year to year is the same order of magnitude as found by Loschnigg and Webster (2000). Variations from year to year are about $\pm$ 0.1 PW.

(ii) With only a few exceptions, the anomalies are coherent in latitude. That is, the anomalies for a particular year are usually either positive or negative from 30°S to 25°N.

(iii) The variability is strongly oscillatory. Figure 13b shows the spectra of the transport across the equator. Three significant bands located at 2, 2.6 and 4 years are apparent suggesting a strong biennial and ENSO periodicity. Each is significant at the 95% level.

In summary, the ocean shows variability in both its thermodynamical and dynamical structures on interannual time scales. Furthermore, the response of the ocean appears to be compensatory suggesting that there exist negative feedbacks on interannual time scales that regulate or govern the system.

3.3 Interannual regulation of the monsoon

Figure 6 suggested a coupled ocean-atmosphere regulatory system of the monsoon annual cycle that reduced the amplitudes of seasonal variability in both hemispheres. It is a simple matter to extend this system to explain the oceanic compensations acting from year to year noted in the last section. For example, let us assume that the North Indian Ocean SST were warmer than normal in the boreal spring. The ensuing stronger monsoon flow would produce greater fluxes of heat to the southern hemisphere in the same manner found in Figure 12b. If the North Indian Ocean were cooler then one would expect a reversal in compensation. In a sense, the interannual regulation described here is very similar to the Meehl (1994a) theory. Like the Meehl theory, the natural time scale of the oscillation is biennial. In addition, because ocean heat transport is an integral part of the theory, it adds a dynamic element to Meehl's theory.

4. General Theory of Regulation of the Coupled Ocean-Atmospheric Monsoon System

In the last section, we modified Meehl's theory by noting that the anomalous monsoon winds will induce ocean heat transports that will reverse the sign of the monsoon anomaly. However, there are still a number of issues that need to be considered. For example:

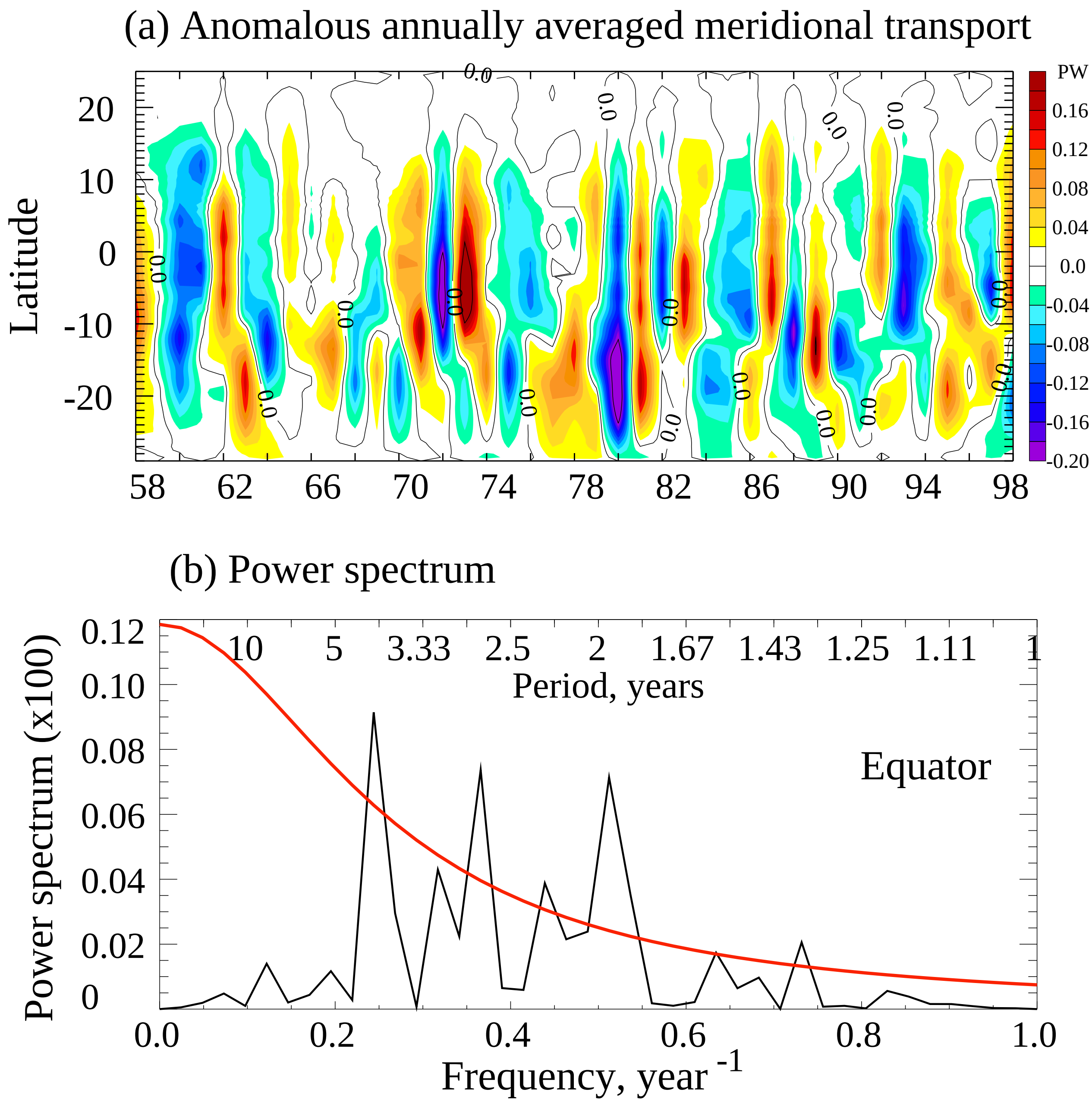


Figure 13 (a) Time-latitude section of the interannual anomalies of the annually averaged northward heat flux in the Indian Ocean. Field was computed by subtracting out the long term average annual northward heat transport from the individual yearly values. Variations from year to year are about ± 0.05 to 0.1 PW. Note the latitudinal coherence of the anomalies. (b) The spectra of the anomalies of the annually averaged northward heat flux in the Indian Ocean. The 95% significance line relative to a red noise null hypothesis is shown. Three major significant peaks emerge at 2, 2.5 and 4 years.

(i) Whereas there are dynamic elements added to the Meehl theory, one is still faced with the problem of maintaining an upper ocean temperature anomaly from one year to the next.

(ii) The regulation theories, either the Meelh theory or the modified Meehl theory, do not involve the Indian Ocean dipole. It could be possible, of course, that the dipole is an independent phenomena. However, the similarity of the basic time period of the dipole to that of monsoon variability (essentially biennial) and the fact that the dipole emerges during the boreal summer monsoon suggests an interdependence. The problem, though, is how to incorporate an essentially zonal phenomena (the dipole) with the essentially meridional phenomena (the oceanic heat transport) into a general theory of the monsoon.

A theory which takes into account these two problems is now developed. It is shown schematically in Figure 14. The figure displays a sequence through two monsoon seasons starting (arbitrarily) in the boreal spring. There are three columns in the figure representing the anomalous meridional oceanic heat transports (column 1), the influence of the anomalous monsoon circulation on the ocean (column 2), and the evolution of the dipole (column 3). During the two year period, the monsoon goes through both weak and strong phases. At the same time, the dipole progresses through a positive and negative phase. We will now argue that the morphology of the dipole and the monsoon are intimately related.

(i) The left hand column of Figure 14 describes essentially the regulation theory discussed in section 3.3. The sequence starts with an anomalously cold North Indian Ocean in the boreal spring (March-May of the first year: MAM:1) which often precedes a weak monsoon (e.g., Sadhuram 1997, Hazzallah and Sadourny 1997, and Clark et al. 2000a) in the summer of the first year (JJA:1). A weak monsoon is associated with a reduced southward heat transport leading to the SST distribution in the first boreal fall (SON:2) shown in the second figure of the row. If the anomaly persists through to the second spring (MAM:2), it will lead to a strong monsoon in the second summer (JJA:2). Enhanced southward transports and reduced net heating of the Indian Ocean leads to an anomalously cold northern Indian Ocean.

(ii) Stronger and weaker summer monsoons also influence the ocean system in other ways. For example, the anomalous monsoon circulation (Figure 7) will influence the upwelling patterns in two major areas: Along the east Africa coast north of the equator, and along the western coast of Sumatra. Reduced southwesterly flow of Africa will decrease upwelling while offshore flow near Sumatra will enhance upwelling. Thus, during the weak monsoon of JJA:1 the changes in the monsoon circulation will create anomalously warm water in the west and colder water in the east. On the other hand, the circulation associated with the strong monsoon in JJA:2 will produce cooler water in the eastern basin (enhanced southwesterlies) and warmer water in the east (onshore winds along the Sumatra coast). In summary, changes in the monsoon winds between strong and weak monsoons can create zonal anomalies in the SST distribution.

(iii) The east-west SST gradients caused by the anomalous monsoon intensities can lead enhancements of the zonal SST gradients by coupled ocean-atmosphere instabilities as described in detail in Webster et al. (1999) and summarized in section 3.1.2. Simply, the SST gradients force zonal wind anomalies which change the distribution of low-latitude sea-level height distribution. During JJA:1, the sea level height will slope upwards to the west while during JJA:2 it will slope upwards to the east.

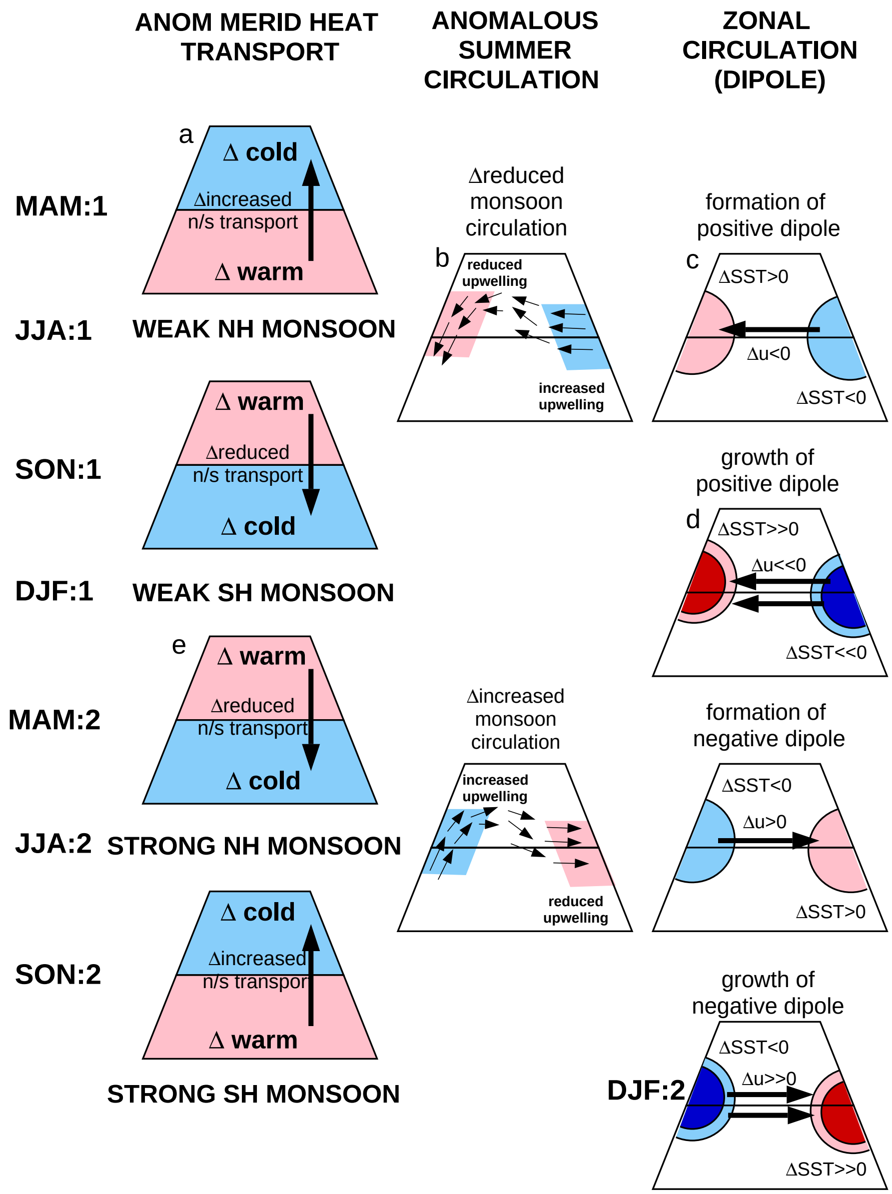


Figure 14 Schematic of a general theory of an ocean-atmosphere regulation system for the monsoon and the Indian ocean. Each column indicates a set of processes. The first column shows modulation of the monsoon variability by changes in the heat transport induced by the monsoon winds. In essence this sequence represents the Meehl (1997) biennial oscillation mechanism but with ocean dynamics. The second column shows the impact of the strong and weak monsoons on the upwelling regions of the ocean basin. The wind patterns are schematic representations of those computed in Figure 7. The third column represents the development of the Indian Ocean dipole relative to the upwelling patterns developed by the anomalous monsoon wind fields. Growth of the dipole anomaly is assumed to follow the coupled ocean-atmosphere instability described by Webster et al. (1999). Taken as a whole, the figure suggests that there are multiple components that regulate the monsoon with each component acting in the same sense. One important role of the dipole (either positive or negative) is to provide slow dynamics (or memory) to the SST anomalies induced by the strong or weak monsoons. For example, the sequence (a) to (e) helps the perpetuate the northern hemisphere anomalously warm temperatures created by the weak monsoon during the previous summer.

Relaxation of the sea-level height takes place in the form of equatorial modes. For example, during the period JJA:1 through DJF:1 the relaxation will be in the form of downwelling Rossby waves. Besides having a slow westward propagation they are downwelling and deepen and warm the western Indian Ocean. In turn, the enhanced SST gradient will produce stronger easterly winds which will continue to maintain the east-to-west slope of the surface. Between JJA:2 and DJF:2 the onshore winds towards Sumatra will deepen the thermocline and enhance the zonal west-to-east SST gradient. In turn, the winds themselves will be enhanced by a wind field responding to an increasing SST gradient. The important aspect of the dipole is that it introduces slow dynamics into the system.

(iv) Careful inspection of Figure 14 shows that the impact of the dipole is to enhance the SST distributions associated with meridional heat transports shown in the first column. For example, the dipole that develops in the period JJA:1 through DJF:1 will increase the SST in the northwest equatorial Indian Ocean. This SST enhancement can be seen by following the sequence "a" through "e" in Figure 14. On the other hand, the second dipole will cool the SST in the same location. This is the region found by Sadhuram (1997), Hazzallah and Sadourny (1997), and Clark et al. (2000a) to correlate most strongly in the winter with the following monsoon. Thus, the role of the dipole is to enhance and prolong the SST patterns necessary to regulate the intensity of the monsoon system.

5. Conclusions

In the preceding paragraphs, we have developed a theory that regulates the monsoon on both annual and interannual timescales. The study was motivated by noting that the surface heat balances in the Indian Ocean do not match the observed evolution of the SST indicating the importance of ocean heat transports. Furthermore, it appeared curious that the year-to-year variability of the South Asian monsoon is relatively small. Hence, we developed a theory of regulation of the monsoon rests on negative feedbacks between the ocean and the atmosphere. We are now in a position to address some of the questions raised in the introduction.

Perhaps the most important conclusion is that it is clear that the ocean involves itself in a dynamic manner. Although the ocean is responding to forcing from the atmosphere, the response is such that there is a strong feedback to the atmosphere. This feedback governs the amplitude and phase of the annual cycle and also modulates interannual variability.

One of the problems that emerged in earlier theories of monsoon amplitude regulation (e.g., Meehl 1997) is that it is difficult to understand how an SST anomaly pattern produced by an anomalous monsoon can perpetuate from one year to the next. By involving ocean dynamics in the regulation process we have managed to introduce mechanisms that allow SST anomalies to persist from one year to the next. This was accomplished by noting that the Indian Ocean dipole is also parented by an anomalous monsoon through the generation of zonal temperature gradients between upwelling regions. The slow dynamics of the dipole act to enhance the zonal SST gradient initiated by the anomalous monsoon irrespective of the sign of the initial perturbation. As the dipole grows, the SST anomalies produced occur in locations that are conducive to the generation of a reverse anomaly in the monsoon. In other words, the dipole adds the slow dynamics needed in the Meehl theory.

An immediate question is whether or not the dipole is an independent entity or a function of forcing from the Pacific Ocean. First, there is irrefutable statistical evidence (Table 2) that ENSO variability in the Pacific Ocean produces a response in monsoon variability. Also, asubstantial amount of the variance of the dipole can be explained in terms of the Niño-3 SST variability. However, there are periods when the dipole and ENSO are unrelated statistically when it is difficult to find an ENSO extrema to match the development of the dipole. Such a year was 1961. There may be a way of resolving this apparent paradox. Let us assume that the dipole is a natural mode of oscillation in the Indian Ocean in much the same way as El Niño is a natural oscillation of the Pacific. In fact. Webster et al. (1999) described a coupled ocean-atmosphere instability that supports this hypothesis. We also recall that the dipole is initiated by an anomalous monsoon. It is quite possible that weak monsoon induced by El Niño will induce, in turn, a positive dipole which then acts to reverse the impacts of ENSO during the following year. But, according to the hypothesis, a dipole will develop relative to an anomalous monsoon no matter how the monsoon is perturbed. Perhaps in 1961, and other such years, other factors could have perturbed the monsoon.

The results presented in this study are incomplete. They have depended on a number of empirical studies and experiments with stand-alone ocean models forced with atmospheric fields. The complete associations inferred in the conclusions above are difficult to establish from empirical studies because of the very convoluted nature of the phenomena. However, a considerable effort has been to ensure that the results are consistent with a multifaceted phenomena. Further work will probably have to await experimentation with fully coupled ocean-atmosphere land models.

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