1.1 The monsoon

The term ‘monsoon’ is derived from the Arabic word *mausam*, which means seasons. The term monsoon is widely used to denote any annual climate cycle with seasonal wind reversal in both tropical and subtropical regions resulting seasonal changes in both atmospheric circulation and associated precipitation. These changes arise from the reversal in heating and temperature gradient between continents and adjacent oceans with the progress of seasons. The extremes in the seasons are denoted as summer and winter. The name southwest monsoon is used for the rain and southwesterly winds of summer season and northeast monsoon to the winter.

Ramage (1971) formulated four criteria to define the monsoon areas as:

i. the prevailing wind direction shifts by at least $120^0$ between January and July;

ii. the average frequency of prevailing wind direction in January and July exceeds 40%;

iii. the mean resultant wind in at least one of the months exceeds $3 \text{ms}^{-1}$ and;

iv. fewer than one cyclone-anticyclone alternation occurs every two years in either months in a $5^0$ latitude-longitude rectangle.

Based on the above criteria, monsoon regions include almost half of the African continent, south and east Asia and northern Australia (shaded portions in figure 1.1), in which more than half of the world population resides.
As monsoons are such dominant features, the social and economic welfare of many tropical countries is intimately linked to the vagaries of monsoon cycle.

1.1.1 The monsoon makers

The three general mechanisms account for the cause and existence of monsoons are as follows

i. *The differential heating of the oceans and continents*: Over the vast ocean regions, tropical circulation has little variations, but over continents it has a seasonal rhythm. The relative warming of land over ocean produces heat lows, which gradually forms regions of convergence. The air is drawn into the low from the winter hemisphere crossing the equator. In winter the condition reverses. Thus the tropical oceans and continents experience a semi-annual reversal in wind direction, which characterises the monsoon.

ii. *Moisture processes in the atmosphere*: As moist warm air rises over summer time heated land surfaces, the moisture eventually condense, releasing energy in the form of latent heat of condensation. This extra energy raises summer land-ocean pressure differences adding vigor to monsoon.

iii. *Earth’s rotation*: As the earth rotates on its own axis and produces a rotation force called Coriolis force, air in the monsoon currents moves in curved paths. Inter-hemispheric difference in Coriolis force also causes wind to change direction as they cross the equator.
1.1.2 The annual cycle and mean monsoon

The Asian monsoon system contains two distinct seasons: the wet and the dry. The wet occurs during the boreal summer when warm and moist winds blow across south Asia from southwest. During dry winter season, the winds reverse, blowing cool and dry air from the winter continent across Indian subcontinent from the northeast. The two conditions are expressed schematically in figure 1.2.

Figure 1.2 Boreal summer and winter season wind pattern and ocean heat transport.

The relationship between the general mechanism that generate monsoon, the seasonal climate cycle and monsoon annual cycle is described by Webster (1987). In the transition months between the southern and northern hemisphere summer (April and May), the intertropical convergence zone (ITCZ), which is the region of surface low pressure, rising air movements and convergence of air masses is located in the equatorial region, which is the region of maximum heating at this time. At this period of seasonal cycle, the northern hemisphere tropical-subtropical latitudes are beginning to warm up with weak vertical motions. The northern hemisphere Hadley cell is still predominates with offshore air flow. With the northward movement of the Sun in May to June, heating in the northern hemisphere land asses increases strengthening vertical motion. Northern hemisphere moisture content increases as a result of intensification of southern hemisphere Hadley cell and predominant wind direction is offshore. By this time ITCZ and associated precipitation has moved well north of the equator along with the onset of monsoon in many areas.
During June-July period sensible heat input at surface becomes maximum, as vertical motion and moisture content over land masses. The monsoon attains its maximum intensity with maximum precipitation and northward extent. By September the heating again comes close to equator as in the April period. This indicates the cessation of northern hemisphere wet monsoon season. By December, maximum heating and ITCZ has moved well south of the equator. Northern hemisphere Hadley cell starts strengthening and zone of maximum heating attains its maximum southward position.

The schematic model of annual cycle described here is for mean annual conditions. But in reality, there is a considerable variation in onset, duration and magnitude of monsoon.
The seasonal climatology of precipitation and wind at 850 hPa during the summer (June to September, JJAS) and winter (December to February, DJF) are illustrated in figure 1.4. During northern summer, the low level circulation over Indian Ocean and the subcontinent is dominated by strong cross equatorial flow and southwesterly winds. The monsoonal circulation acts as a moisture conveyor belt, transporting moisture from the south Indian Ocean and the Arabian Sea towards the south Asian land masses and the Bay of Bengal (Fasullo and Webster, 1999). The relatively steady moist, unstable air adected over the south Asian landmasses enhances convection and precipitation across south Asian monsoon region. Heavy rainfall experiences over south and southeast Asian region as well as through much of the northern Indian Ocean including the eastern Arabian Sea and Bay of Bengal. Maximum precipitation is also found in the winter hemisphere just south of the equator.

The mean convection and precipitation during the northern winter shifts from southern Asia to along and south of the equator over the maritime continent and northern Australia and extending into the South Pacific Convergence Zone (SPCZ). The 850 hPa circulation has northeasterly winds over the Indian subcontinent and cross equatorial flow from north to south over the maritime continent which provides moisture for the Australian monsoon rainfall (figure 1.4b) The boreal summer south Asian monsoon tends to be stronger than boreal winter Australian monsoon both in terms of precipitation and the strength of monsoon circulation. The difference is in large part due to the presence of an elevated heat source in the form of Tibetan Plateau affecting the south Asian monsoon. The onset of summer monsoon is coincident with the reversal of the meridional temperature gradient in the upper troposphere south of the Tibetan Plateau (Flohn 1957; Li and Yanai, 1996). In the southern hemisphere the reversal of temperature gradient is not evident. But still the cross-equatorial pressure gradient that drives the winter monsoon are the result of intense radiative cooling over north Asia during winter (Webster et al. 1998).
Figure 1.4: Seasonal climatology of precipitation (shaded, mm day$^{-1}$) and 850 hpa wind (vector, m s$^{-1}$) for (a) summer season (June to September) and (b) winter (December to February).

Figure 1.5: Seasonal climatology of sea surface temperature (shaded, K) and 200 hpa wind (vector, m s$^{-1}$) for (a) summer season (June to September) and (b) winter (December to February).
The climatology of 200 hPa wind and Indian and Pacific Ocean sea surface temperature (SST) are shown in figure 1.5 for both the seasons considered above. The predominant feature of 200 hPa wind is the persistent upper level westerly jet poleward of 20° latitude in both the hemispheres. Over South Asia and Africa, there is an upper level easterly jet during summer that is replaced by an upper level westerly jet during the winter. Thus on average the flows in both the seasons are generally opposite in direction to low level flow yielding an easterly wind shear during summer and westerly wind shear during winter. Cross-equatorial return flow is also evident during summer, large part of which is divergent contributing to the local Hadley cell (Krishnamurthi, 1971).

Mean SSTs in excess of 28°C are confined to equatorial latitudes. In the Indian Ocean basin and western Pacific Ocean, a clear latitudinal shift of warm SSTs into the summer hemisphere is evident. The entire Bay of Bengal and the eastern Arabian Sea possess warm SSTs during summer and is evenly distributed along the equator in the winter season. The expansion of warm SST eastward in the central equatorial Pacific Ocean and westward in the Indian Ocean during the northern winter is evident in the figure 15a and b.

1.2 The Indian summer monsoon

Indian summer monsoon, which is an active part of south Asian monsoon, gives more than 75% of the annual rainfall to the Indian land masses affecting the agriculture and thus economy of the country. The Indian summer monsoon is made up of the following components suggested by Krishnamurthi and Bhalme (1976) and is shown in figure 1.6

1) The monsoon trough over north India- It is formed in the summer months as extension of global ITCZ and is region of surface low pressure and wind shear. South of the trough has southeasterly winds and north has northwesterly winds.

2) The Mascarene anticyclone system and cross-equatorial jet- It is high pressure system formed over southeast Indian Ocean. From this high large outflow of air takes place and it crosses the equator as cross-equatorial Somali jet and becomes southwesterly wind. Reaching at maximum
intensity in the summer months, it crosses the southern Arabian Sea and reach over central western and southern coast of India. Variation in the intensity of this jet is important in determining the rainfall over India.

3) **The Tibetan high pressure system**- This is an upper level anticyclone found above the surface monsoon trough located over north India. By July it is established over Tibetan High lands and is well developed at 200 hPa level. It remains upto the end of summer season and then moves south-southeast direction with the movement of maximum heating to south.

4) **The tropical easterly jet**- The outflow of air from the southern flank of Tibetan anticyclone gives rise to tropical easterly jet and it remains from June to September.

5) **Monsoon cloudiness and rainfall**- Cloud cover is the manifestation of moist convective processes over the Indian subcontinent and it varies both in space and time. During active phase of monsoon, cloud cover is maximum in a belt running from western shore of Bay of Bengal to northern shore of Arabian Sea and minimum over foot hills of Himalaya, south India and Sri Lanka. The pattern reverses during break phase. Rainfall distribution closely flows the cloud distribution.

Figure 1.6 : Components of Indian summer monsoon system (Krishnamurthi and Bhalme, 1976)
Chapter 1  

1.3 Variability of the monsoon

The Asian monsoon exhibits high amplitude variability on all time scales from synoptic to interdecadal. The main causes for the variations are internal dynamics and boundary forcing (Shukla, 1987). Internal dynamics controlling monsoon includes a variety of aperiodic variations in the atmospheric circulations such as traveling disturbances, thermal and orographic forcing, non-linear interaction between different scales of atmospheric motions etc. Boundary forcing refers to changes in surface conditions, like spatial extent of snow cover, surface hydrological effects and sea surface variations. These changes can alter the geographical distribution of sources and sinks of heat and moisture in the atmosphere and thus patterns of tropical winds.

On the basis of time scales of variation summer monsoon variability can be divided generally into three categories:

- **Intraseasonal** (with in the season)
- **Interannual** (between years)
- **Inter decadal** (more than a decade)

### 1.3.1 Intraseasonal variability

While defining the variability of a monsoon system in its seasonal character, its variability about a typical season is of most interest and important. The southwest monsoon is marked by episodes of prolonged abundant precipitation, active periods separated by periods of reduced rainfall, break days. The transition from active to break periods and vice versa evolve slowly such that typically 3 to 4 active periods are seen over a typical single monsoon season (Webster et al., 1998). Relative frequency of occurrence of active and break phases could influence the seasonal mean and contribute to interannual variability of intraseasonal oscillation (ISO).

The active and break periods of south Asian monsoon or the wet and dry spells over the Indian continent, are manifestation of repeated northward propagation of the Tropical Convergence Zone (TCZ) from the equatorial position to the continental position (Yasunari, 1979; Sikka and Gadgil, 1980) and results from superposition of a 10-20 day and a 30-60 day oscillations. Both the 10-20 day oscillation and the 30-
60 day oscillation contribute almost equally to the total intraseasonal variability in this region. While 30-60 day oscillation has a very large zonal scale encompassing both the south and east Asian and west Pacific monsoon regions, the 10-20 day oscillation has a smaller zonal scale and is regional in character. The 30-60 day mode is characterized by a northward propagation, while the 10-20 day mode is characterized by a westward propagation.

Two mechanisms seem to contribute to the temporal scale selection of the 30-60 day mode. One is a 'convection-thermal relaxation feedback mechanism' (Goswami and Shukla, 1984), according to which convective activity results in an increase of static stability, which depresses convection itself. As convection weakens, dynamical processes and radiative relaxation decreases moist static stability and brings the atmosphere to a new convectively unstable state. This mechanism does not involve wave dynamics and may be responsible for the northward propagating 30-60 day oscillations not associated with eastward propagation of convection in the equatorial region. The other mechanism involves eastward propagation of convection the equatorial Indian Ocean in the form of a Kelvin wave and west northwest propagation of Rossby waves emanated from the equatorial convection over the western Pacific (Wang and Xie, 1997). The time scale is determined in this case by propagation time of the moist Kelvin wave from the eastern Indian Ocean to western Pacific, the moist Rossby waves from western Pacific to the Arabian Sea where they decay and a new equatorial perturbation is generated.

The monsoon ISO is crucial building blocks of the Asian summer monsoon. Through multi-scale interactions with synoptic activity on one hand and the seasonal cycle on the other, they determine not only the probability of occurrence of daily precipitation but also the interannual variability of the seasonal mean. Thus, the ISOs also have the potential to produce interannual variability of the seasonal mean precipitation.

1.3.2 Interannual variability

Among the wide range of variabilities of Asian summer monsoon, the interannual variability (IAV) is most extensively studied. The interannual variability of Asian monsoon is the yearly deviation of seasonal transition from mean annual cycle. The
interannual variability is characterized by a set of seasonally and spatially varying characteristic features. Many large-scale phenomena like ENSO which affect the monsoon also fluctuates significantly in interannual time scales. The interannual variability of the south Asian monsoon is rather modest with the interannual standard deviation being about 10% of the seasonal mean. However, larger excess or deficit of all India rainfall are associated with large spatial scale covering most of the country (Shukla, 1987). Extremes in monsoon rainfall leads to devastating floods and droughts (Shukla, 1987; Webster et al., 1998) leading to enormous economic loss and human misery. Therefore, understanding of the physical processes responsible for the observed IAV of south Asian monsoon is crucial for advancing the capability for predicting the IAV.

Although the variability of the monsoon is largely controlled by the internal dynamics of the atmosphere, which is related to the amount and distribution of solar energy, the slowly varying boundary forcing from underlying ocean and land also plays a major role in affecting the interannual variability of monsoon. The seasonal mean tropical circulation is thought to be influenced to a greater extend by the boundary conditions rather than internal dynamics (Charney and Shukla, 1981; Shukla, 1981).

Among the external causes of interannual variability of Asian monsoon, sea surface temperature (SST) is perhaps the leading impacting factor (Chao and Chen, 2001). Effect of SST on monsoon can be the remote effect from tropical central and east Pacific and local effect from oceans near Asian continent. There are several studies, both observational and modelling indicating that the interannual variability of Indian summer season (June to September) monsoon rainfall is linked to the SST variation in Pacific (Rasmusson and Carpender, 1983; Mooley and Parthasarathy, 1983; Ju and Slingo, 1995; Soman and Slingo, 1997). The relationship between the east Pacific SST and monsoon can be regarded as problems of ENSO-monsoon relationship. There is a tendency for the El Nino to be associated with drought and La Nina to be associated with above normal conditions over India. This interaction is primarily through the change in the equatorial Walker circulation influencing the regional Hadley circulation associated with the Asian monsoon (Webster et al., 1998, Goswami, 1998; Lau and Nath, 2000).
SSTs in the west Pacific and Indian Ocean are also believed to be important for South Asian monsoon. Although the variabilities of these regional SSTs are not independent from ENSO, there is considerable effect of regional SST on interannual variability of monsoon. Regional SSTs influence monsoon through changes in surface heat and moisture fluxes, direct moisture supply, thermal difference between land and ocean and waves generated anomalous atmospheric heat sources.

1.3.3 Interdecadal variability

The south Asian monsoon (SAM) precipitation does not seem to have any climatic trend but has epochs of roughly three decades when the precipitation has the tendency to be more above than below normal followed by a roughly three decades when it has the tendency to be more below than above normal. The large scale circulation changes associated with the interdecadal variability may lead to change in teleconnection patterns in these time scales. For example, the ENSO-monsoon relationship is known to undergo low frequency variations in these time scales. It may be noted that the simultaneous as well as the lag relationship between all India rainfall and Nino3 SST anomalies has undergone major changes during the recent years compared to earlier decades. Our understanding of the interdecadal variability of the SAM remains much poorer than our understanding of the mean annual cycle and its intraseasonal and interannual variabilities. Modulation of interannual variability by interdecadal variability influences the predictability of seasonal mean monsoon. Correlation between several predictors and Indian summer monsoon has been found to undergo interdecadal variability (Kumar et al., 1999). A better understanding of interdecadal variability may therefore, be very important in improving the predictability of mean seasonal monsoon climate.

1.4 Modes of interannual variability

In the periodicities longer than annual cycle, we have interannual variability corresponding to biennial periodicity (Yasunari, 1987; 1991; Rasmusson et al., 1990; Barnett, 1991) and multiyear variability corresponding to ENSO scale (see figure 1.6 b).
One notable connection with the interannual variability of the south Asian monsoon is that with the ENSO. The occurrence of El Niño is generally associated with a weak monsoon, and La Niña is associated with a strong monsoon (e.g., Webster and Yang, 1992). During normal periods, the warm pool (SST> 27°C) extends from the eastern Indian Ocean to the western Pacific Ocean and is associated with a broad precipitation maximum. During an El Niño event, the locus of maximum SST in the Pacific Ocean shifts eastward, typically bringing more precipitation over the central and eastern Pacific Ocean. During these periods the eastern Indian Ocean, Indonesia, and south Asia are in the subsiding part of the Walker circulation that has shifted eastward from its climatological position (Webster et al., 1998). The strong heat source associated with the south Asian monsoon could indeed influence the atmospheric circulation in a significant way and could modify the surface stresses over the central and western Pacific and influence the strength and evolution of the ENSO (Yasunari, 1990; Chung and Nigam, 1999; Kirtman and Shukla, 2000). These independent studies of ENSO influence on the ASM and ASM influence on the ENSO, made it clear that the ENSO and the ASM are not independent phenomena but part of a coupled ocean-atmosphere oscillation.

While a connection between the south Asian monsoon and the ENSO exits, but it is not possible to predict the strength of the monsoon solely from the phase of ENSO, as these monsoon–ENSO correlations have variable lag-lead times (Webster and Yang, 1992). It is worth noting here that many droughts and floods of the south Asian monsoon occur without El Nino or La Nina. The correlation between Indian

Figure 1.6: (a) Indian summer monsoon rainfall anomaly (cm) from 1950-2004. (b) wavelet analysis of ISMR anomalies from 1950-2004, dotted line indicates 95% significance level.
rainfall and Pacific Oceans SST varies from 0.4 to 0.8 for different decades from 1900 through to the present (Torrence and Webster, 1999). Kumar et al., (1999) gives the evidence of weakening of the reverse monsoon-ENSO relationship during the last two decades.

The failure of ENSO-monsoon relationship in the recent decades forced to look into the relationship of monsoon with the newly discovered tropical phenomena like tropospheric biennial oscillation (TBO) and Indian Ocean Dipole (IOD). The biennial tendency of IAV of the asian summer monsoon has been known for a long time (Mooley and Parthasarathy, 1984; Yasunari, 1990; Clarke et al., 1998; Webster et al., 1998). Meehl (1987) first identified the quasi-biennial signal in summertime all India rainfall. Subsequent studies identify the spectral peak that exists in AIR between 2.5 and 3 yr and interpret it as a tendency for the monsoon to alternate often between successive “strong” and “weak” years, a sequence attributed to large-scale coupled land–ocean–atmosphere interactions in the Indian and Pacific Ocean regions (e.g. Meehl 1997).

Both observational (Yasunari, 1990; 1991; Ropelewski et al., 1992; Yasunari and Seki, 1992; Kane, 1995; Yang et al., 1996; Tomita and Yasunari, 1996; Harzallah and Sadourny, 1997; Webster et al., 1998) and modeling (Goswami, 1998; Ogasawara et al., 1999; Kitoh et al., 1999; Chang and Li, 2000; Li et al., 2001) studies have documented characteristics of tropical biennial variations. Strong biennial signal is observed in both precipitation and circulation features of Asian summer monsoon. Numerous conceptual models of the oscillation have also been developed (Meehl, 1987; 1993; 1994; Goswami, 1995; Webster et al., 1998; Kawamura et al., 2001; Clarke et al., 1998; Loschnigg and Webster, 2000; Kim and Lau, 2001; Loschnigg et al., 2003). More recently, an active role for equatorial eastern Pacific Ocean interactions has also been incorporated (Meehl and Arblaster 2002a, b; Sahai et al. 2003; Wang et al. 2003).

Various mechanisms have been proposed for the TBO (e.g. Nicholls, 1978; Meehl, 1987; Meehl, 1993; Clarke et al., 1998; Goswami, 1995; Chang and Li., 2000). The modeling studies mentioned above provides a synthesis of the TBO studies and the ENSO-monsoon connection studies and shows that they are linked and part of the
same air-sea coupled oscillation involving both the Indian Ocean and the Pacific Ocean basins

1.4.1 Tropospheric Biennial Oscillation

The tropospheric biennial oscillation (TBO) is defined as the tendency for a relatively strong monsoon to be followed by a relatively weak one, and vice versa, with the transitions occurring in the season prior to the monsoon involving coupled land–atmosphere–ocean processes over a large area of the Indo-Pacific region (Meehl, 1997, Meehl and Arblaster, 2002a; Meehl et al., 2003). Thus the TBO is more a tendency for the system to flip-flop back and forth from year to year, and not so much an oscillation. The more of these interannual flip-flops or transitions, the more biennial the system. One of the most remarkable features of TBO is its characteristic seasonal progression and dynamically coherent spatial structure. The conditions of a strong (or weak) monsoon over India and southeast Asia in the northern summer often continues to the succeeding autumn and winter over the maritime continent and Australia monsoon region. So that a strong (weak) Asian monsoon is often followed by a strong (weak) Australian monsoon.

The TBO has become recognized as a candidate for understanding some of the processes that can contribute to interannual variability of a variety of parameters in the Indian and Pacific regions in both observations (Yasunari, 1990; Ropelewski et al., 1992; Yasunari and Seki, 1992; Yang et al., 1996; Tomita and Yasunari, 1996; Meehl, 1997; Meehl and Arblaster, 2001) and models (Ogasawara et al., 1999; Chang and Li, 2000). Coupled climate interactions between ocean and atmosphere contribute to a mechanism that produces biennial variability (TBO) in the troposphere and upper Ocean in the tropical Indian and Pacific Ocean regions. The TBO is associated with modulations of seasonal cycle, with maximum in the TBO is manifested as warm (El Niño) and cool (La Niña) in the tropical Pacific and have connections to Indian monsoon.

1.4.1.1 Role of Oceans and Asian-Australian monsoon in TBO

An important part of any biennial mechanism is anomalous heat storage in the ocean and the associated SST anomalies that can occur in certain regions. Thus, the ocean retains the “memory” of ocean–atmosphere interaction over the course of a
year to affect the atmosphere the following year (e.g., Brier, 1978; Nicholls, 1978; Meehl, 1987; 1993; Chang and Li, 2000; Li et al., 2001). Several studies have suggested that coupled ocean dynamics plays a role in the formation and maintenance of these heat content and SST anomalies associated with the TBO (Meehl, 1993; Clarke et al., 1998; Webster et al., 1999; Saji et al., 1999; Meehl and Arblaster, 2002a; Loschnigg et al., 2003). Meehl (1993) also noted significant upper-ocean heat content anomalies in regions of the tropical Pacific that contribute to the memory of the system and, thus, the TBO. Yu et al., (2003) through modeling study argued that east Pacific SST is important for inphase Indian to Australian transition in TBO cycle and Indian Ocean plays major role in out of phase transition from Australian to Indian monsoon. Pacific Ocean SST anomalies maintain large amplitude during inphase transition in northern autumn and reverses sign in spring season, in which out of phase transition is occurring. On the other hand, the Indian Ocean SST anomalies maintain large amplitude during out-of-phase transition and reverses sign in northern spring. These seasonally dependent anomalies in these two oceans allow these two oceans to play different role in transition of phase of TBO.

Li et al (2001) and Wu and Kirtman (2007) are of the opinion that biennial transition of Indian summer monsoon can produce through the local air-sea interaction in the northern Indian Ocean when Pacific ENSO is suppressed. Wu and Kirtman (2007) showed that the local SST anomalies in the Indian Ocean induce monsoon transition through low level moisture convergence. Surface evaporation anomalies affect the SST changes. The anomalous condition in the Indian and Pacific Ocean through regional and remote effect in spring season can produce anomalous monsoon conditions.

A plausible mechanism through which ocean-atmosphere coupling leads to a TBO and the role of Asian summer monsoon is given by Wu and Kirtman (2004). A strong asian summer monsoon (ASM) during boreal summer (June to September) can enhance surface easterlies in the central equatorial Pacific, induces an eastward propagating upwelling Kelvin wave and gives rise to negative SST anomalies in the eastern Pacific that amplifies through air-sea interactions. Colder SST in the eastern Pacific is also associated with warmer SST in the western Pacific. A strong ASM also cools the Indian Ocean through enhanced evaporation and upwelling.
Associated intensification of the Walker circulation leads to divergence of moisture supply in the western Indian Ocean. Reduced moisture supply at low levels together with upper level subsidence leads to a weaker ASM during the next summer. A weak ASM induces opposite affects and can lead to a stronger monsoon next year. Thus, the ocean-atmosphere interaction generates IAV of the ASM via generation of TBO signal.

1.4.1.2 Theories of TBO

Understanding the TBO in the Asian-Australian, the Pacific Ocean and the Indian Ocean sector often relies on the knowledge of variations of Asian-Australian monsoon, ENSO and Indo-Pacific SST as well as the transitions of these phenomena with seasons and geographic locations. The origin of the TBO is a subject of debate. The existing theories on TBO emphasis the following three aspects of ocean-land-atmosphere coupling viz; (a) Local air-sea interaction (Brier, 1978; Nicholls, 1978; Meehl, 1987; Clarke et al., 1998), (b) Remote ocean-atmosphere interaction (Meehl, 1987; Chang and Li, 2000; Kim and Lau, 2001; Li et al., 2001, Meehl and Arblaster, 2002 a,b), and (c) Tropical- extratropical teleconnection (Yasunari and Seki, 1992; Meehl, 1994, 1997; Tomitta and Yasunari, 1996; Ogasawara et al., 1999).

The local air-sea interaction theory emphasizes the importance of basic flow and the interactive processes of atmospheric and oceanic anomalies (see Nicholls, 1978). A westerly wind anomaly superimposed on westerly basic flow intensifies the total wind and cools the ocean. By this same argument, westerly wind anomaly weakens a easterly basic flow and warms upper ocean. The changes in the atmosphere associated with these changes are considered as a delayed feedback of the SST anomaly. On the other hand, Meehl (1987, 1993) stressed the role of initial SST anomalies in local air-sea interaction leading to TBO. Meehl (1997) described an air–sea negative feedback mechanism in which warm spring SSTs in the Indian Ocean enhance atmospheric convection. The ensuing stronger monsoon leads to greater than average wind strength, increased Ekman transports and vertical mixing, and higher heat loss by evaporation throughout the summer monsoon season, which causes subsequent cooling of the ocean surface. The low ocean temperatures persist.
for one year until the next pluvial season. The lowered SSTs are associated with less convection than the previous spring, producing weakened winds, and reduced heat loss by evaporation, and less mixing, which leads to higher SSTs than the year before. The cycle is thus repeated. Figure 1.7 shows the biennial cycle due to land-atmosphere and ocean-atmosphere interactions as illustrated by Meehl (1997).

One of the prominent features of Asian-Australian monsoon system is the migration of monsoon convection from northern Australia to southern Asia from boreal winter to summer and in opposite direction from summer to winter. In this migration...
process, a strong (weak) Asian monsoon is followed by a strong (weak) Australian
monsoon. Meehl and Arblaster (2002a) proposed a remote forcing mechanism
emphasizing the importance of eastern Pacific SST for the large scale east-west
circulation and the Asian monsoon. The southeastward migration of a strong Asian
monsoon through Australia weakens the south Pacific High and thus the trade
winds. This increases eastern Pacific SST and decrease zonal SST gradient and
east-west atmospheric cells across tropical Pacific. As a result of the diminished
upward motion in the western Pacific weakens next monsoon.

Tropical–extratropical teleconnection mechanism (Meehl, 1997) emphasis the
response of extratropical atmosphere to tropical forcing for TBO. The response of
the extra tropical circulation causes anomalies in land surface temperature.
According to Ogasawara et al. (1999), the convective activity around Indonesia and
northern Australia is the main tropical forcing and anomalous atmospheric
circulation over Asia is direct Rossby wave response to tropical heating. Favorable
interaction between the land and atmosphere processes prolongs the anomalies of
the coupled land-atmosphere system.

Using a simple five-box model that considers SST-monsoon, evaporation-wind,
monsoon-Walker circulation, and wind-thermocline feedback, Chang and Li (2000)
showed that these processes can give rise to the seasonal evolution of TBO. Their
theory explains why TBO can maintain the same phase from northern summer to
northern winter and why a reversed phase of TBO can last three locally inactive
seasons to affect the next year's monsoon. Without the interference from and
interaction with motions of other scale, Chang and Li (2000) obtained a pure,
regular biennial oscillation. Differ from Meehl (1997), Chang and Li (2000) put
importance to Asian monsoon than east Pacific SST for the TBO mechanism.
According to them a strong monsoon produces anomalous westerlies in the Indian
Ocean and increases east-west circulation in the Pacific. As a result of the westerly
anomalies SST cools in the Indian Ocean and it persists to next monsoon making it
weak. Increase in west Pacific SST due to the deepening of west Pacific
thermocline favoring strong Australian monsoon in the forthcoming boreal winter.
In addition, strong trade winds over Pacific decrease east Pacific SST which
maintain west Pacific warm through easterly anomaly and shoaling thermocline.
The mechanism proposed by Chang and Li (2000) is given in figure 1.8.

Figure 1.8: Schematic diagram indicating the interactive processes leading to TBO. Land and Ocean regions are shaded. Non shaded boxes indicate atmospheric (oval) and oceanic (rectangular) processes. The strong monsoon phase starts with warm Indian Ocean SST leading to strong south Asian monsoon. The reverse phase is in the upper right corner separated by thick ribbon (Chang and Li, 2000).

Another potential explanation for biennial monsoon variability involves land surface processes, especially Eurasian snow cover. Extensive snow cover typically precedes a weak monsoon and light snow cover precedes a strong monsoon (Barnett et al. 1989; Vernekar et al., 1995; Yang, 1996). Large and persistent winter snow cover over Eurasia following a strong monsoon can delay and weaken the spring and summer heating of landmass that is necessary for the establishment of large scale monsoon flow (Shukla, 1987; Barnett et al., 1989; Yasunari et al., 1991). In this manner, snow cover may act as a negative feedback upon the Asian monsoon, imparting biennial variability. Clearly a better understanding of the ocean–atmosphere and land–atmosphere feedbacks that may give rise to the TBO can improve predictions of monsoon strength.

The transitions (from relatively strong to relatively weak monsoon) occur in northern spring for the south Asian or Indian monsoon and northern fall for the Australian monsoon involving coupled land atmosphere–ocean processes over a large area of the Indo-Pacific region. Transitions from March-May (MAM) to June–September (JJAS) tend to set the system for the next year, with opposite sign in the
following year. Indian and Pacific SST forcing are more dominant in the TBO than
circulation and meridional temperature gradient anomalies over Asia (Meehl and
Arblaster, 2002a). A fundamental element of the TBO is the large-scale east–west
atmospheric circulation (Walker circulation) that links anomalous convection and
precipitation, winds, and ocean dynamics across the Indian and Pacific sectors. This
circulation connects convection over the Asian–Australian monsoon regions both to
the central and eastern Pacific (eastern Walker cell), and to the central and western
Indian Ocean (western Walker cell). From all these above discussions Meehl and
Arblaster (2002b) proposed a mechanism involving convection, SST, wind, extra
tropical circulation, east-west circulation etc and is illustrated in figure 1.9 starting
from the winter season (DJF) before the strong monsoon to the next winter after the
monsoon.

In the winter (DJF) season, warm SST is seen in the Indian Ocean and east Pacific
and cool SST is seen north of Australia. This will look like an El Nino condition.
Under these circumstances, the Eurasian continent is characterized by less snow and
warmer conditions as a result to the tropical heating. In the following spring
(MAM) the conditions of coupled ocean-atmosphere system persists because of the
memories in the ocean and thermocline conditions. As a result summer monsoon
increases. Because of these, the Eurasian land surface is cool and SST decreases in
the Indian Ocean and west Pacific. These summer conditions are maintained to next
post monsoon season (SON). These above conditions lead to the strong Australian
monsoon in the next winter (DJF). The strong convection associated with
Australian monsoon and weak convection over the western Indian Ocean and the
central Pacific contribute an anomalous trough and cold land temperature over
Eurasia (as a Rossby wave response), which are followed by a weak south Asian
summer monsoon.
Figure 1.9: schematic picture representing TBO features associated with anomalies of convection, SST, surface winds, extratropical circulation and equatorial Pacific and Indian Ocean thermocline orientations (Meehl and Arblaster 2002b)
But in reality, TBO exhibits a rich, irregular spectrum. Such irregularity may, in large part, result from nonlinear interaction with ENSO. Since TBO is an inherent mode of the coupled ocean-atmosphere system in which monsoon is a major part, the prediction of anomalous rainfall over the Asian-Australian monsoon region based on ENSO forecast must consider the combined effects of TBO and ENSO. The strength of the TBO is modulated by interdecadal variability. For example, Torrence and Webster (1999) find the biennial oscillation to be weaker in some decades than others. Difficulties in fully understanding the relationship between the TBO and monsoon arise at least partly due from lack of understanding the internal and external features of the TBO. The TBO in Asia, Australia and the adjacent tropical regions can be considered as inherent mode of monsoon itself. However, the variability of SST in the tropical central Pacific is also characterized by quasi-biennial signal and is coupled with the Asian-Australian monsoon. It is even more difficult to understand the relationship between the internal monsoon characteristics and external ENSO features of TBO, which is the result of interaction of annual cycle and ENSO cycle. Quantifying the role of these conditions associated with the transition of TBO and their relationship with ENSO and the accurate representation of these in models is needed for better forecasting of monsoon.

Eventhough a good account of TBO mechanism is in the literature, it is not differentiated from ENSO in the Pacific. The role of monsoon, land surface processes, ocean etc are not well studied. The ENSO also has a good biennial cycle. The ENSO- biennial cycle and TBO are not differentiated up to now. The present study tries to answer all these questions by analyzing the following parts of TBO-monsoon relationship

- Mechanism of biennial and low frequency variability of monsoon
- Atmosphere- Ocean pattern associated with TBO
- Circulation features of TBO.
- The difference of TBO in the absence of active Pacific Ocean
- Role of Indian Ocean processes like Indian Ocean Dipole (IOD) in TBO
- Is intraseasonal Oscillation is modified during TBO period?
- Possibility of QBO-TBO interaction
- Effect of climate shift on TBO and interannual variability of Indian summer monsoon