Chapter – 1
INTRODUCTION

Monsoon, generally defines as a system of winds characterized by a seasonal reversal of its direction. Indian Meteorological Department (IMD) defines it as the seasonal reversal of the wind direction along the shores of the Indian Ocean, especially in the Arabian Sea, that blows from the southwest during one half of the year and from the northeast during the other half. In other words, seasonal wind, which blows with consistency and regularity during a part of the year and which is absent or blows from another direction for the rest of the year. Such seasonal changes of wind are primarily the result of differences in the quantity of heat received from the sun by different parts of the earth. Most of the solar energy received by the land is used up in heating the air rather than the earth’s surface. On the other hand, solar energy is able to penetrate too much greater depths in the oceans because of the stirring which go on under the action of the wind.

One may describe the monsoon as a system of winds with the following features:

1. A system, with marked seasonal wind shifts caused by the differential heating of the land and the sea i.e. by the different response of land and ocean to incoming radiation from the sun.
2. Winds that are largely confined to the tropics i.e. the region between 20° N and 20° S on both sides of the equator.
3. Summer monsoons over the northern hemisphere may be thought of as the Southeast trades or trade winds of the southern hemisphere which, on crossing the equator, are deflected to the right by the earth’s rotation and as a consequence approach the land from a south westerly direction.

The temperature contrast between land and sea must be very pronounced and persistent, to be able to generate a monsoon. But, if the thermal contrast is on a smaller scale and lasted a short while, a phenomenon that is similar to the monsoon called a sea-breeze is often observed near a coast. The monsoon occurs over the tropics because it is the region of the world that receives most solar energy and is able to build up intense thermal gradients between land and oceans. The similarity between the monsoon and the land or sea breeze has been noticed by the English astronomer
and a remarkable scientist Edmund Halley in 1686 and later extended by George Hadley in 1735 with an explanation of the trade winds which is essentially true even today. The circulation over the tropics is named as Hadley cell in honour of George Hadley. Monsoon is considered broadly due to the differential heating of northern hemisphere land and southern hemisphere ocean. That is why, there is a reversal of circulation in summer and winter and they are not meridional in nature because the Coriolis force tilts them. This is one of the factors why the summer monsoon enters Indian subcontinent from southwest direction and is termed also as south west monsoon (SWM).

C S Ramage (1971) has suggested four features of monsoon winds:

1) The wind direction should shift by at least 120° between January and July.
2) The average frequency of prevailing wind directions in January and July should exceed 40 percent.
3) The mean resultant winds in at least one of the months should exceed 3 m/s.
4) There should be less than one cyclone-anticyclone alteration every two years, in either month, over a five degree latitude/longitude grid.

These characteristics emphasize the seasonal nature of the change in wind direction and the persistence of the wind regime in each season. The change must not be a temporary one, which is brought about by variations in the track of moving pressure systems, such as tropical cyclones.

A precise definition of the monsoon is a matter of choice, because one need hardly frame a set of rules which embraces all facets of the monsoon. Monsoon winds are most pronounced in the summer season of northern hemisphere, i.e. during the months of June, July, August and mid September (JJAS). During June the trade winds from the southern hemisphere penetrate deep into the northern hemisphere towards India, the wide stretches of Southeast Asia and to a lesser extent towards Africa. The mountains of the eastern coast of Africa are also important because they help to deflect the southeast trades towards India. The seasons of long and short rains over east Africa, which correspond to the summer and winter monsoons of the northern hemisphere, are generated not so much by differential warming but by seasonal shifts of trade winds.
The monsoons associated with other continental masses of the earth are not well marked as the Indian monsoon but seasonal changes in wind direction are known to occur over north Australia, western and eastern Africa and southern USA also. Most pronounced monsoon winds of the world are those flows over the Indian sub-continent in the hundred–day period between the beginning of June and middle of September each year. It brings relief to a dry and parched land in the form of rain and affects Indian agriculture in a very substantial measure. Indeed, the impact of the monsoon on Indian economy is very pronounced. The vagaries of the monsoon are now proverbial. Excessive rain has led to floods in certain areas, while little or no rain in other parts has brought in its wake drought and famine resulting in acute distress to millions. Such fluctuations in rainfall have engaged the attention of people. The earliest systematic study of rainfall data by H F Blanford (1889) yielded much valuable information on the distribution of rainfall. Sir Gilbert walker (1910) relies on finding associations between a meteorological element, such as rainfall and some other variable at an earlier time has discovered significant relation between monsoon rains over the peninsular India and South America pressure April and May. When the pressure is high over the southern pacific, it is low over the Indian Ocean but once every few years the pattern is reversed i.e. the pressure over the Indian Ocean became high while lower pressures prevailed over the southern pacific.

1.1 Regional aspects of the monsoon:

1.1.1 Monsoon Trough:

With the advance of the monsoon, a zone of low pressure builds over northwest India as a result of excessive solar insolation, gradually extends eastwards until it forms an elongated low pressure zone running parallel to the Himalayan mountains in a west to east direction. Its axis is roughly parallel to the Himalayan foothills. Monsoon trough is not a quasi-stationary system but shows periodical movements from its normal position. When it moves north and lies close to the Himalayan foothills, a remarkable change in the rainfall pattern occurs. The rains cease abruptly over the plains of northern India but increase equally and rapidly in intensity over the foothills of northeast India known as break in monsoon rains. Another feature of a break is the westward passage of low pressure system across the Indian peninsula. Parts of the peninsula which lie in the rain shadow of the Western
Ghats derive much of their monsoon rainfall during break situations. On the other hand, when the axis of the monsoon trough moves south and tends to dip into the Bay of Bengal, conditions become favorable for the formation of a low or a depression. Consequently, a southward position of the monsoon trough is usually an indication of well distributed rain over central India and Indo–Gangetic plains.

1.1.2 Monsoon Depressions:

A good part of the rainfall during the monsoon is generated by the westward passage of depressions or low pressure systems in Bay of Bengal. In the monsoon months especially in July and August, one to three systems are observed on an average. The life time of these systems is around a week.

1.1.3 Sub Tropical Jet Streams:

Towards the end of May, a narrow stream of air which moves from the west to the east over northern India, suddenly weakens and moves to a new location far to the north of the Himalayas. This is known as sub-tropical westerly jet stream. Its movement towards the north is one of the main features associated with the monsoon onset over India. It occurs at a lower altitude of 300 hPa. Another jet stream sets in over the southern half of the Indian peninsula, which flows from the east to the west called the tropical easterly jet, exhibits periodic movements to the north and south of its mean location during the hundred days monsoon season beginning with the first of June and ending around mid-September occurs at an altitude of 150 hPa.

1.1.4 Tibetan High:

Thermal convection is stronger over the Tibetan plateau because of its higher elevation. Tibetan high is one of the important factors influencing Indian summer monsoon (ISM). There is an anti cyclone at 200 hPa which in other words leads to divergence at higher levels and induces convergence at lower Levels. This is a large anticyclone during the northern summer months known to have its largest amplitude near 200 hPa. The high pressure belt is to the south of Tibet during June and September and over Tibet during July and August between 500 hPa to 200 hPa.
1.2 Onset and Withdrawal of Summer Monsoon:

There are three main phases of monsoon. First and foremost is the monsoon onset, second is the progress of the monsoon and third is the withdrawal of monsoon. Observing the monsoon climatology it is evident that the monsoon onset is on June 1st over Kerala coast (IMD 1943; IMD weather information is available online at http://www.imd.gov.in). The normal onset dates of summer monsoon over India (IMD) are shown in figure in 1.1. By the end of June a monsoon is set over the entire country. Fieux and stommel (1977) have reported that 25 May, with a ± 5-day uncertainty is the average date of onset of monsoon winds in the Arabian Sea about 500 Km north east of Somalia. Monsoon onset time is not the same throughout the Arabian Sea (Halpen and woiceshyn 1999). It takes some six weeks for the rains to be fully established over the entire India. There are substantial year to year variations in the date of onset of the monsoon over Kerala. The date of the Monsoon onset over Kerala (MOK) is declared by IMD every year based on subjective estimates prepared by operational forecasters. The IMD forecaster, while declaring the date of the MOK, has been taking into consideration subjectively the following features (Rao 1976):

1) The rainfall should be widespread over Kerala and adjacent areas with large rainfall amounts at individual stations;
2) This rainfall persists over several days;
3) The lower-tropospheric westerlies in and around Kerala should be strong and deep; and
4) The air should be rich in moisture (high relative humidity) up to at least 500 hPa.

Studying the onset of the monsoon over India during the four years 1979 to 1982 (Pearce and Mohanty 1984) have concluded that the monsoon onset consists of two main phases: 1) Moisture build-up over the Arabian Sea during which synoptic and mesoscale transient disturbances develop. 2) A rapid intensification of Arabian Sea winds and a substantial increase in Latent heat release over a large region, essentially a large-scale feedback process. According to their study a major part of the moisture required for this is supplied by evaporation over the Indian Ocean, depending on the strength and configuration of the low-level winds (Chen and Dell’osso 1986).
During monsoon onset over India, dramatic changes are known to occur including a rapid increase in daily precipitation rate, an increase in vertically integrated moisture and an increase in kinetic energy, especially of the low-level flows (Krishnamurti 1985). The moisture required for this large area of rainfall is mainly produced in the south Indian Ocean and carried to the convective heat source associated with MOK by a strong cross-equatorial low-level jet (LLJ) stream (Findlater 1969; Joseph and Sijikumar 2004). The 850 hPa zonal wind over 5° N–15° N, 40° E–80° E is such a sensitive parameter to quantify the abrupt transition of large-scale atmospheric preconditions associated with the ISM onset. Monsoon onset has duration of 6 days for wind in the Arabian Sea (Halpen and woiceshyn, 1999) and 3 days for rainfall in India (Das 1987). The onset of SWM which is accompanied by a steep increase of rainfall along the west coast is often referred to as ‘the burst of the monsoon’. The west coast of India, which is on the windward side of the mountain range of Western Ghats, receives very heavy rainfall during the summer monsoon season (Rao, 1976; Soman and Krishnakumar, 1990). Over 70% of the annual rainfall
in India occurs during the summer monsoon season from June to September (Parthasarathy et. al., 1994). The MOK is associated with a large area of organized rainfall caused by deep convection extending eastward a few thousand kilometers from the low-latitude regions of the Arabian Sea in the north Indian Ocean (Joseph et. al., 1994). The intraseasonal variations of the ISM rainfall and associated features have been documented extensively in several studies such as Yasunari (1979), Sikka and Gadgil (1980), Gadgil and Guruprasad (1989), Lawrence and Webster (2002). Withdrawal of monsoon is relatively rapid. The withdrawal of monsoon starts on late August or early September which is completed by September. Cooling of the land masses of northern India and further north and a shift in the activity of the troughs in the westerly wind belt to a relatively lower-latitude belt result in the southward shift of the monsoon trough and withdrawal of the monsoon from northwest India. In general, the withdrawal of the monsoon takes place during the first fortnight of September and withdrawal from most of the parts of the country occurs during the period mid-September to mid-October (Mooley and shukla 1987).

1.3 Intraseasonal Variability:

The fundamental driving mechanisms of the monsoon cycle are the cross-equatorial pressure gradients resulting from differential heating of land and ocean, modified by the rotation of the earth and the exchange of moisture between the ocean, atmosphere and land (e.g., Webster 1987). Number of interesting changes occurs in the circulation of the atmosphere when the summer monsoon sets in over India. The Indian subcontinent experiences vigorous intraseasonal variability during summer with quasi-periodic wet episodes (active phases) and dry episodes (break phases) (Raghavan 1973; Krishnan et. al., 2000; Gadgil and Joseph 2003; Krishnamurthy and Shukla 2007, 2008; Rajeevan et. al., 2008). Since it is recognized that rainfall is the most important facet of the monsoon which gives its direct socio-economic impact, it has been the basis for the identification of active spells and breaks in many studies. According to Pisharoty and Asnani (1960), the break monsoon is caused by eastward extension of the subtropical anticyclone at 500 mb over the Sahara desert and Arabia into central India where the ridge inhibits ascending motion. While corroborating this, Ramaswamy (1962) has attempted to explain how the subtropical anticyclone extends as a ridge to central India in the rear of extra tropical westerly waves. According to
Dixit and Jones (1965), the break-monsoon is the result of westward extension of the Pacific subtropical anticyclone across the Central Bay of Bengal into central India at 500 mb and above. Since the study of Ramamurthy (1969), active spells and weak spells/breaks of the ISM have been extensively studied, particularly in the last decade, however, different scientists have used the same term ‘break’ to denote different features of convection and/or circulation over different regions. The early works of keshavamurty (1973) and Murakami (1977) have indicated the existence of a 10–20 day oscillation in the meridional wind data over north Indian stations. Krishnamurti et. al., (1973) for the first time point out that the spectrum of the Tibetan high has shown a dominant periodicity around 13 days. Krishnamurti and Bhalme (1976) have made a substantial contribution by identifying a quasi-biweekly oscillation in most of the salient elements of the summer monsoon system. They postulate that the alternations between active and break spells could result from cloud feedback and large-scale radiative effects. Fluctuations between active and break monsoon spells on a quasi-biweekly timescale have also been noted by Murakami and Frydrych (1974) and Murakami (1976).

The active-break cycles are linked to observe northward propagation of convection from Indian Ocean on to the Asian subcontinent in summer (Keshavamurty et. al., 1980; Sikka and Gadgil, 1980). This northward propagation has a time scale of 30-60 days and has been noted in many studies (Yasunari, 1981; Krishnamurti and Subrahmanyan, 1982; Lau and Chan, 1986; Wang and Rui, 1990; Gadgil and Asha, 1992). Hartman and Michelsen (1989) have analyzed daily precipitation from Indian stations during 1901-70 and confirmed the existence of 30-50 day variability over peninsular parts of the country. Yasunari (1981) has suggested that the northward migrating monsoon cloud bands are maintained by a transient local Hadley cell and also may be related to the low-frequency Madden Julian Oscillation (MJO) (Madden and Julian, 1972, 1994). The 30–60-day oscillation is linked to the MJO, which is one of the dominant modes of tropical variability on intraseasonal time scales. The MJO can be defined as a 30-50 day oscillation in the large scale circulation cells that move eastward from at least the Indian Ocean to the central Pacific Ocean. It shows significant effect on the atmospheric circulation of the global tropics. Even though the association of active and break periods of the monsoon with MJO is not fully understood, there is abundant evidence of frequency peaks in south
Asian rainfall and wind in the same period bands as the MJO (Julian and Madden, 1981; Wang and Rui, 1990; Madden and Julian, 1994). Recently, Pai et al., (2009) have examined the impact of MJO on the active and break spells over the monsoon core region using the Wheeler–Hendon indices (Wheeler and Hendon 2004). The public safety and economy of India are very vulnerable to the occurrence of extreme active and break phases of the ISM intraseasonal oscillation (ISO), which gives rise to devastating floods, landslides and prolonged severe droughts (Gadgil 1995; Sikka 1999; Webster et al., 1998; Gadgil and Rao 2000; Gadgil et al., 2002; Webster and Hoyos 2004; De et al., 2005). Numerous observational studies during the past few decades have been of considerable use in understanding active and break monsoon spells in terms of subseasonal oscillations of the summer monsoon.

Rodwell (1997), Annamalai and Slingo (2001) use the term ‘break’ to denote weak spells of the daily all-India average rainfall calculated operationally by the IMD. Webster et al., (1998) use the term “break (active) spells” to denote weak (strong) spells of convection and 850 hPa zonal winds over a large scale region (65°–95° E, 10°–20° N). Krishnan et al., (2000) defines the break days as days with positive OLR anomalies over northwest and central India (i.e. only over the western part of the monsoon trough zone), provided the average OLR anomaly over 73°–82° E, 18°–28° N exceeds 10W/m². On the other hand, Goswami and Ajayamohan (2001) have defined ‘breaks’ on the basis of the strength of the 850 hPa wind at the single grid-point 15° N, 90° E. Gadgil and Joseph (2003) have defined breaks (and active spells) on the basis of the daily rainfall over the monsoon trough zone. The break composite of rainfall of Gadgil and Joseph (2003) is very similar to that of Ramamurthy (1969) with positive rainfall anomalies over the Himalayan foothills and south-eastern peninsula. Krishnamurthy and Shukla (2000, 2007) have used the all-India daily rainfall index from the 1°× 1° rainfall data for 1901–1970, prepared from the IMD rain gauge data. Mandke et al., (2007) have identified the active/break days on the basis of the precipitation anomaly over an area 73°–82° E, 18°–28° N, which they called the Indian core region. Hartmann and Michelson (1989), Rajeevan et al., (2006) and Krishnamurthy and Shukla (2008) have used the IMD gridded 1°× 1° rainfall data for identifying the active and break spells.
In the present thesis the criteria given by Rajeevan et. al., (2010) is considered. According to them active and break events are identified by averaging the daily rainfall over the core monsoon zone (the core region is roughly from 18.0° N-28.0° N, and 65.0° E-88.0° E) and standardizing the daily rainfall time series by subtracting from its long term normal (1951–2000) and by dividing with its daily standard deviation. The break spell has been identified as the period during which the standardized rainfall anomaly is less than −1.0, consecutively for three days or more. Similarly, the active periods are identified as the periods during which the rainfall anomaly is more than +1.0 the Standard deviation, consecutively for three days or more. Identification of active spells and breaks is based on the updated version of the IMD gridded rainfall dataset. The criteria are derived from the rainfall over the region over which significant rainfall fluctuations between the active and break spells are observed, viz., the core monsoon zone. Hence Rajeevan et. al., (2010) have considered only July and August months in identifying active and break spells.

1.4 Interannual Variability:

Variability in the southwest Indian Ocean is not just a local phenomenon. A large fraction of interannual variability is determined by the slowly varying surface boundary conditions such as sea surface temperature (SST), surface albedo and soil moisture (charney and Shukla, 1981). Several workers have shown that there is a significant relationship between deficit of rainfall during summer monsoon and El Nino–southern oscillation (ENSO) (Sikka, 1980; Rasmusson and carpenter, 1983; Shukla and paolina, 1983). El Nino and La Nina have long been identified with interannual variations of ISM rainfall (Kane 1998). Goswami (1994) has proposed a conceptual model of how the ISOs influence the seasonal mean and interannual variability of the Indian monsoon. Several recent modeling studies show that a significant fraction of the interannual variability of the seasonal mean ISM is governed by internal chaotic dynamics (Goswami 1998; Hazrallah and Sadourny 1995; Rowell et. al., 1995; Stern and Miyakoda 1995). Most of these studies, however, do not provide any insight regarding the origin of the internally generated interannual variability. Goswami (1997) shows that modulation of the ISOs by the annual cycle could give rise to an internal quasi-biennial oscillation in the tropical atmosphere and influence the interannual variability of the Indian monsoon. To the
extent that the ISOs are intrinsically chaotic and unpredictable, the predictability of the ISM would depend on relative contribution of the ISOs to the seasonal mean compared to the externally forced component. Goswami et. al., (1998) have considered daily surface winds, illustrates that the spatial structures of the intraseasonal mode and that of the dominant interannual mode are similar and made a start in relating the ISOs with the interannual variability from observations. Annamalai et. al., (1999) have examined the relationship between the ISOs and interannual variability. Unlike the results of some earlier model studies, they conclude that there is not a common mode that describes intraseasonal and interannual variability. The Indian Ocean zonal dipole mode (IOD); (Saji et. al., 1999; Webster et. al., 1999) is a basin-scale pattern of surface and subsurface temperature that seriously affects the interannual climate anomalies of many nations around the Indian Ocean rim, as well as the global climate system (Yamagata et. al., 2004).

1.5 Jet streams and Monsoon:

Jet stream is a current of fast moving air found in the upper levels of the atmosphere. Any jet stream is a result of horizontal temperature gradient in the perpendicular direction. The main types are the polar jets, the subtropical westerly jets and the less common tropical easterly jets. The two jet streams which are at the tropopause level and influence monsoon are subtropical westerly jet stream and easterly jet stream. These are high level jets observed at 200 mb which follow the principle of geostrophic winds. Towards the end of May, a narrow Stream of air, which moves from west to east over northern India, suddenly weakens and moves to a new location far to the north of the Himalayas known as sub-tropical westerly jet stream. Another Jet stream sets in over the southern half of the Indian peninsula. This flows from east to west known as tropical easterly jet. They are variant throughout the globe. Whenever there is a continental elevation, there are variations in jet stream. If there are Rocky Mountains, because of more height, jet stream will be also at greater height. Jet streams moving in the same direction, while passing through Himalayas (orientation of east–west and westerly jet stream is also in the same direction), splits into two. One moves towards India (south of Himalayas i.e. near Uttar Pradesh) and other moves towards china. Joseph, P.V and P. L. Raman (1966) have pointed out the existence of a LLJ stream over peninsular India. There is another jet which can be
observed in south India termed as Monsoon Low Level Jet (MLLJ) (Sam and Vittal Murty 2002). MLLJ is closely related to monsoon intensity.

1.6 Findlater Jet with reference to Indian summer Monsoon:

According to a definition suggested by Reiter (1961), LLJ should have marked gradients of wind speed in the horizontal and vertical. The LLJs are generally located in the lowest 1 to 2 Km of the troposphere. These are strongly influenced by orography, friction, diurnal cycle of heating and corresponding variations of pressure gradient and static stability (Asnani, 1993). LLJ over peninsular India plays a vital role in maintaining and controlling the ISM (Joseph and Sijikumar, 2004) first observed by World War II pilots and developed just below the tropopause over areas of steep pressure gradient on the surface.

During the late sixties and early seventies, J. Findlater, a British meteorologist has observed very strong winds (LLJ) in the form of a narrow current of air off the coast of east Africa. This LLJ is found to be most pronounced between 1.0 and 1.5 km. It is observed to flow from Mauritius and the northern part of the island of Madagascar before reaching the coast of Kenya at about 3° S. Subsequently, it ran over the plains of Kenya, Ethiopia and Somalia before reaching the coast again around 9° N. Using monthly mean winds Findlater (1971) has shown that the LLJ splits into two branches over the Arabian Sea, the northern branch intersecting the west coast of India near 17° N, while the southerly branch passes eastward just south of India and this LLJ is termed as Findlater jet (FLJ) or Somali jet which is named after the scientist J. Findlater, who discovered its existence over Arabian Sea. It is this FLJ associated with SWM. The geographical location of FLJ is off the coast of Somalia. So, it is also known as Somali jet. The FLJ appears to be fed by a stream of air which moves northwards from Mozambique Channel. The major part of the FLJ penetrates into east Africa during May and, subsequently, traverses the northern parts of the Arabian Sea before reaching India in June.

The importance of the FLJ also arises from the fact that its path around 9° N coincides with a zone of coastal upwelling. This upwelling is brought about by strong low level winds. Coastal upwelling is responsible for creating a narrow strip of cold sea SSTs off Somalia. The SSTs in this region could be as low as 15° C, while off Mumbai on the west coast of India the SSTs are between 25° C-30° C. The strong
temperature gradient that is set up has an impact on monsoonal winds. FLJ plays a vital role in SWM dynamics and the only jet stream which can be seen on ocean and influence the land. It contributes lot of moisture and momentum flux to the monsoon system. A number of interesting changes occur in the circulation of the atmosphere when the summer monsoon sets in over India. The summer monsoon is inarguably an important facet of life in India, whether the aspect is economic (Webster et. al., 1998) or cultural (Zimmermann 1987). Intense south westerly surface winds in the Arabian Sea and heavy rainfall along the west coast of India are annual occurrences of the summer monsoon. A linkage between SWM winds in the Arabian Sea and rainfall over India has been indicated more than a quarter–century ago (Findlater 1969). Halpen and woiceshyn (1999) have shown that eastward expansion of the Somali jet raise the intensity of surface wind convergence and increases the amount of integrated cloud liquid water in the eastern Arabian Sea, which, presumably influences the rainfall of the west coast of India.

P.V. Joseph and S. Sijikumar (2004) have done a detailed study of FLJ and have improved on the observations of Findlater. According to them,

1) The core of the cross equatorial FLJ crosses the equator in a geographically fixed narrow longitudinal belt close to the East African coast as a southerly current and it crosses India as a westerly current at latitudes varying from the equator to 25º N. In active monsoon conditions, the core of the FLJ passes through peninsular India around latitude 15º N. In break monsoon conditions the FLJ from the central Arabian Sea moves south eastward and passes eastward close to Sri Lanka in the latitude belt from the equator to 10º N. There is often seen at this time a weaker FLJ axis through north India around latitude 25º N.

2) FLJ does not show splitting into two branches over the Arabian Sea as suggested by Findlater (1971). His suggestion which is widely accepted since then is based on the analysis of monthly mean winds. Such an analysis is likely to show the FLJ of active and break monsoons as occurring at the same time, suggesting a split of the FLJ over the Arabian Sea. Two branches of FLJ through India are however seen during break monsoon spells but the northern branch is at around latitude 25º N and not at about 17º N as found by Findlater.

3) Convective heating of the atmosphere over the Bay of Bengal has a high and significant linear correlation coefficient with the zonal component of the wind at 850
hPa over peninsular India (70° E–80° E) and the Bay of Bengal (80° E–100° E) all between latitudes 10° N and 20° N. The intensity of FLJ and the convection intensity have a lag of 2-3 days that means first convection starts and after 2-3 days the maximum is observed. It is speculated that active convection occurring over the Bay of Bengal between latitudes 10° N and 20° N accelerates the whole inter hemispheric FLJ and takes the monsoon to an active spell splitting of jets as a barotropic phenomenon has been mentioned in several studies (Thompson, 1957; Arakawa, 1961; Wiin-neilsen, 1961; and others).

One of the main objectives of the present thesis is to investigate the influence of FLJ on SWM. Sometimes FLJ changes its orientation from north-south to east-west. So, the influence on monsoon onset in general and monsoon circulation is examined. As this FLJ contributes lot of moisture and momentum in flux to the monsoon system an attempt is made how the FLJ is important and the exact correlation between FLJ intensities and rainfall over Kerala, South India and India is examined. The relation between FLJ and SST, Outgoing Longwave Radiation (OLR), Latent Heat Flux (LHF), Sensible Heat Flux (SHF) are also examined.

1.7 Sea Surface Temperature and Monsoon Rainfall:

SSTs have become a well-known acronym since 1997 when widespread media coverage has been given to the El Nino phenomenon and rightly so, because scientists agree that SSTs do influence Indian summer monsoon rainfall (ISMR). It has been postulated in earlier studies (viz., Saha 1970a, 1974; Ellis, 1952), that the SST over the Arabian Sea may have important influences on the monsoon flow and associated rainfall. Several workers viz., Saha (1970 a, b, 1974) have postulated the possible effects of the SST anomaly (SSTA) and can be qualitatively described as follows:

1. Warmer SST anomalies (SSTAs) and stronger winds may cause higher evaporation and the monsoon current may be more moist and unstable.

2. Colder SSTAs may cause higher surface pressures and less evaporation over the Arabian Sea and this may reduce the cross-equatorial moisture flux and thus reduce the rainfall over India.
3. Higher pressures over the western Arabian Sea and lower pressures over the eastern Indian Ocean may set up west circulation cells.

Indian Ocean SST variability is most important to understand and assess the impacts of air sea interactions on the evolution and the potential predictability of climate and monsoon in the adjacent region. Indian Ocean SST is significantly correlated with the All-India Rainfall index several months preceding the onset of the monsoon ((Shukla and Mooley (1987), Rao and Goswami (1988), Harzallah and Sadourny (1997), and Sadhuram (1997)). Gradients of SST within the oceans are important in determining the location of precipitation over the tropics, including the monsoon regions (Lindzen and Nigam 1987). A cool Arabian Sea SST leads to reduced Indian rainfall and vice versa (Shukla 1975; Kershaw 1988; Yang and Lau 1998). Earlier studies (Shukla, 1975; Rao and Goswami, 1988) have suggested that the Arabian Sea SST is important because of its possible role on the interannual variability of ISM. Model simulations also confirm that the Arabian Sea SST influences subsequent monsoon rainfall on timescales less than a month. A General Circulation Model (GCM) study in which the solar forcing of the land and ocean are incorporated separately shows that the annual cycle of SST in the Indian Ocean is crucially important in establishing the monsoon circulation and rainfall (Shukla and Fennessy 1994). Recent studies have identified that Indian Ocean SST has a strong seasonal variability (Wolter and Hastenrath 1989; Krishnamurthy and Kirtman 2003) and its variability is related to the ENSO variability (Klein et. al., 1999; Krishnamurthy and Kirtman 2003).

Many researchers have pointed out that the SST in the tropical Indian Ocean effects not only the rainfall in India (Shukla and Mooley, 1987) and Australia (Nicholls, 1985), but also the East Asian monsoon (Xiao and Yan, 2001), the north-western Pacific subtropical high (Wu et. al., 2000) and the summer rainfall in China (Chen, 1991). So, more and more studies have focused on the Indian Ocean. Recently, the discovery of the Indian Ocean dipole event (Saji et. al., 1999) which shows that the Indian Ocean can develop an independent ENSO-like variability mechanism (Webster et. al., 1999) gives rise to new interests in this region (Murtugudde and Busalacchi, 1999; Li and Mu, 2001). Studies on SST dynamics in the Pacific (e.g., McPhaden and Hayes, 1991; Feng et. al., 1998; Qiu, 2000), many scientists focus on
the heat budget and SST in the tropical Indian Ocean. These studies certainly shed light on the physics of SST variation but they are limited to small regions in the tropical Indian Ocean and the results are region means or at a single point although Rao and Sivakumar (2000) have investigated the heat budget and SST for the whole tropical Indian Ocean, based on observations, the equation they used is an approximate one. In fact, all heat budget equations used to “explicitly” diagnose SST change need to be extended. The ISM rainfall is most strongly associated with Indian Ocean SSTs in the seasons that precede the monsoon (Clark et. al., 2000, Li et. al., 2001). Lying on both sides of the Indian subcontinent, air-sea interaction over the Arabian Sea and the Bay of Bengal are found to influence precipitation variability associated with ISM.

It is well known that the El Nino is associated with devastating droughts over the western tropical Pacific (called Tuarang in Indonesia), torrential floods around the eastern tropical Pacific rim and unusual weather patterns over various parts of the world (Philander; 1990). However, Saji et. al., (1999) have recently pointed out that the Indian Ocean gives birth to another unique coupled ocean-atmosphere mode, which may induce unusual rainfall in the surrounding area including the tropical East Africa. They have called the new climate signal the "Dipole Mode (DM)" based on SST and corresponding wind anomalies over the tropical Indian Ocean. Webster et. al., (1999) and Yu and Rienecker (1999) have also reported the DM event being occurred during 1997-98. According to Saji et. al., (1999), the DM event has also occurred in 1961, 1967, 1972, 1982, 1994 and 1997. The DM structure is characterized by the cold SSTA in the south eastern tropical Indian Ocean (SETIO) and the warm SSTA in the western tropical Indian Ocean (WTIO) (Vinayachandran et. al., 1999; Behera et. al., 1999; Saji et. al., 1999; Webster et. al., 1999; Yu and Rienecker, 1999; Murtugudde et. al., 2000). DM event may occur independently of the ENSO as a unique Air-Sea coupled phenomenon over the tropical Indian Ocean (Saji et. al., 1999; Webster et. al., 1999).

The recently much discussed Indian Ocean Dipole/Zonal mode (IODZM), whose dynamics shows some similarities to ENSO, seems to be strongly damped and climatically significant events have only been reported 3 times since 1960 (Annmalai and Murtugudde; 2004). More recently, Li et. al., (2001) have shown that the Indian
Ocean SST, including that of the Arabian Sea, plays the dominant role on the biennial oscillation of the ISM. These studies have suggested that the pre monsoon SST could be used as a predictor for the ISM. It has been noted that a weak ISM is followed by warm SSTAs in the following winter and spring (Li et. al., 2001; Babu and Joseph, 2002), which implies that the influence of the memory of the Arabian Sea on the atmosphere is significant. Therefore, it is necessary to understand the summer cooling accurately.

1.8 Outgoing Longwave Radiation and Monsoon activity:

The OLR data has been examined to study the annual and interannual atmospheric variability in the tropics by Heddinghaus and Krueger (1981). Prasad and Verma (1985) have found OLR data is useful in the study of large scale monsoon circulation and associated cloudiness and rainfall over the Indian regions. The OLR data have been used in estimating quantitative precipitation over the tropical Pacific by Motell and Weare (1987), and over the tropical Atlantic Ocean by Yoo and Carton (1988). Over the tropical ocean, because of relatively stable surface temperature, OLR variations are almost entirely owing to changes in the distribution of cloudiness, and hence, are related to changes in precipitation (Arkin et. al., 1989). Muthuvel and Arkin (1992) have examined the interannual and long-term climatic variations in the tropics using OLR data. Janowiak and Arkin (1991), Xie and Arkin (1998) have investigated the relationship between precipitation and OLR, and developed a new technique to estimate monthly precipitation for the entire globe. Prasad et. al., (2000) have also found OLR over the Indian Ocean useful in long range prediction of the ISMR.

The interannual variations of the ISMR are found to be significantly related to the interannual variation of OLR of west–central India, south Equatorial Indian Ocean is in the opposite phase with the ISMR, and east of Madagascar Sea is in phase with the ISMR. Hence, low OLR (or high convective activity) of these two regions represents high ISMR whereas cloud free region and OLR, in this case, appears to be modulated owing to variation in the SST. Therefore, high OLR of the region represents high ISMR (Prasad and Bansod 2000).
1.9 Heat Fluxes and Monsoon activity:

Evaporation at the Air–Sea interface results in a transport of energy and water vapor into the atmosphere. The energy transport partly compensates losses of energy through radiation processes in the atmosphere. The global mean of this energy transport is equivalent to 26% of the incoming solar energy at the top of the atmosphere. On this account, the exchange of the energy between the sea surface and the atmosphere is a major energy source for the atmospheric circulation. The exchange of water vapor and heat at the surface takes place simultaneously and connects the energy cycle to hydrological cycle. Thus, the LHF causes a cooling of the upper layer of the ocean and through the loss of water, an increase of the salinity in the oceanic mixed layer. Good estimates of LHF at the sea surface (together with fluxes of momentum and sensible heat) with global coverage could be very useful for verifying coupled ocean–atmosphere models, as well as for driving ocean models. They are also very useful for understanding the basic physics of Air–Sea interaction.

Manabe et. al., (1975) have shown that air-sea fluxes act as lower boundary condition for the atmosphere and as an upper boundary condition for the ocean in the large scale coupled model. Mohanty and Mohan Kumar (1990), using MONEX-79 (Monsoon Experiment) data have shown that air-sea fluxes of heat and moisture play an important role in relating the marine boundary layer fluxes and the monsoon. Mohanty et. al., (1994) has discussed in their study that it is important to understand the connection between the air-sea fluxes and monsoon activity rather than the SSTA and the monsoon activity. The SST and its variation effect air-sea fluxes during the summer monsoon season, the LHF anomaly shows significant positive anomaly over parts of the Arabian Sea. During May significant negative anomalies (heat loss) are found in the south-west Arabian Sea off Somali coast, North Bay of Bengal and some areas of south Indian Ocean. The negative anomalies (oceanic heat loss) are mainly attributed to strong low level wind and enhanced evaporation, leading to upwelling, cooling of sea surface and enhanced pre-monsoon convective activity with more cloud cover and reduction of incoming short-wave radiation.

The results in the pre-monsoon month of May confirm an earlier study by Mohanty et. al., (1996) and serve as a useful predictor for the subsequent monsoon season rainfall activity. The significant positive difference of LHF in May over
southwest Arabian Sea may be considered as a useful advance indicator of the possible behavior of the subsequent monsoon season. There are significant regions of higher LHF located over part of Arabian Sea off Somali coast and south Indian Ocean off Madagascar during the monsoon season. Kemball-Cook and Wang (2001) have examined the influence of LHF on the northward migration of OLR anomalies over the Indian Ocean. They suggest that the moisture flux from the ocean to the atmosphere indicated by negative latent heat flux anomalies favors convection by building up moist static energy. Several modeling studies have shown that land cover/use changes are a major component of change in regional hydrological cycles, affecting the partitioning of net radiation into sensible and latent heat fluxes in India (E. M. Douglas et. al., 2009). Recent observations suggest that there are plausible physical linkages between changes in surface heat and moisture due to vegetation and crop management and monsoon variability in India (T. W. Biggs et. al., 2008, E. Lee et. al., 2009) and East Asia (K. Takata et. al., 2009, E. Lee et. al., 2008).

Chapter 2 deals with the data used for the analysis over the study area. The study area includes the Arabian Sea from 0º N-30º N and 50º E-75º E, which has different wind pattern during SWM. To study the climatology one needs data for a longer time, so the data has been taken for a period of fourteen years i.e. from 1997-2010. The parameters used for the study are Wind, Rainfall, SST, OLR, LHF and SHF. The results of climatology, structure and interannual variability of FLJ during active and break spells (in the period 1997-2007) and also in El Nino and La Nina years are presented in chapter 3. It also includes the FLJ wind (intensities) over the Arabian Sea and has found the cross equatorial flow is stronger in active spells than in break spells. Variations of FLJ intensities before and after onset are discussed and correlated with rainfall over Kerala. In chapter 4, variations of monsoon rainfall over all India, South India and Kerala are examined with FLJ intensities. The FLJ intensities indicate the extent of moisture transport from Arabian Sea. This moisture transport reaches Indian subcontinent and results in rainfall. It also includes the intraseasonal variability of FLJ and its association with SST and various Air-Sea interaction parameters like OLR, LHF and SHF individually. Finally the conclusions are given in chapter 5. The significant outcome of the present thesis is that the FLJ plays an important role in the intraseasonal and interannual variability of Asian monsoon activity.