CHAPTER 1

INTRODUCTION

1.1 Relevance of the topic

The Asian monsoon system is an integral component of the earth's climate system, involving complex interactions of the atmosphere, the hydrosphere and the biosphere. The majority of the population of the planet reside in the monsoon regions. India, with a population of nearly a billion people, is located in the central portion of South Asia and is predominantly within the monsoon regime. The annual variation of rainfall and temperature has a major influence on the livelihood and well being of the country. The change in the large scale atmospheric and oceanic circulations due to the redistribution of monsoon heat sources and sinks is also affecting weather and climate in regions far away from the monsoon region. Droughts and floods due to the changes in monsoon affect agricultural and industrial production and cause property damages, human suffering and death and spread of diseases thus posing serious threats not only to the monsoon community but also on the global scale.

The summer monsoon contributes about 80% to the annual rainwater potential of India. Traditionally in India, agricultural practices have been tied to the annual cycle of the monsoon. Variations in the timing and quantity of rainfall significantly affect the productivity of crops. A monsoon year with less total rainfall than normal generally corresponds to low crop yields (*Gadgil*, 1996). An intense monsoon can cause considerable distress as a result of severe flooding and crop destruction . In addition, within the monsoon rainy season there are great variations of precipitation, the so called intraseasonal variations. The long periods of excess rainfall are often referred to as *active* phases of the monsoon and the dry periods of little rain as *break* phases. Since ploughing and planting periods are extremely susceptible to the changes in monsoon rains, the intraseasonal variability has a direct influence on the agricultural sector. Even if the average seasonal monsoon rains are normal, an ill-timed arrival or cessation of rainfall can cause crop destruction (*Webster et al.*, 1998). With accurate forecasts, the impact of variability of the monsoon on agricultural practices, water management, etc., could be optimised. Since the Asian monsoon system is a dominant manifestation of a strongly interactive ocean-atmosphere-land system, understanding the mechanism that produce the variability in the monsoon is very much needed for developing accurate prediction methods.

1.2 Justification for the study

A strong cross-equatorial Low Level Jetstream (LLJ) with core around 850 hPa exists over the Indian Ocean and south Asia during the boreal summer monsoon season, June to September. LLJ has its origin in the south Indian Ocean, north of the Mascarene High as an easterly current, it crosses the equator in a narrow longitudinal belt close to the east African coast as a southerly current with speeds at times even as high as 100 knots, turns into a westerly current over the Arabian Sea and passes through India and enters the western Pacific Ocean.

LLJ has strong horizontal shear (vorticity) in the planetary boundary layer (maximum vorticity at 850 hPa) and is important for generation of convective rainfall and for the development of rain producing weather systems. The LLJ is the main conduit for transporting moisture generated over Indian Ocean to the monsoon area. The intraseasonal oscillation of LLJ controls this moisture distribution. In active monsoon condition a large convective region over the Bay of Bengal is maintained by the LLJ passes through peninsular India. It is found that this convection maintains the LLJ. When convection over Bay of Bengal decreases and it gets established over the equatorial Indian Ocean at the time of break monsoon, the LLJ bypasses India and is located south of India. It then transports bulk of the moisture evaporated over the Indian Ocean to the west Pacific Ocean and the moisture supply to India is cut off. Thus LLJ plays an important role in monsoon rainfall distribution over the Indian subcontinent.

1.3 Monsoon

The term *monsoon* appears to have originated from the Arabic word *Mausam* which means *season*. With change of season there is also generally a change of wind direction; hence the word monsoon also came to be associated with wind directions. In a true monsoon climate, seasonal wind shifts typically cause a drastic change in the general precipitation and temperature patterns. Monsoons have been subject of considerable research work for more than a century. According to *Ramage* (1971) the main characteristics of monsoon regions are as follows

- *The prevailing wind direction shifts by at least 120° between January and July.*
- *The average frequency of the respective prevailing wind directions in January and July exceeds 40%.*
- *The mean resultant wind speed in at least one of the months exceeds 3 ms^{-1}.*
- *Fewer than one cyclone-anticyclone alternation occurs every two years in either month in a 5° latitude-longitude region.* This **monsoon region** which includes parts of the African continent, South Asia and North Australia is shown as shaded area in Figure 1.1.

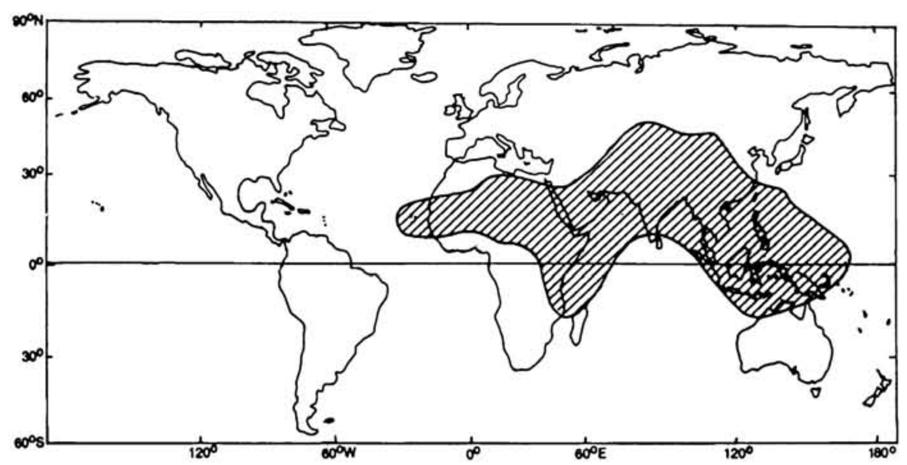

Figure 1.1: Areas with monsoon circulations according to Ramage criteria (*Ramage*, 1971)

There are three fundamental driving mechanisms of the planetary-scale monsoon. These are *(i) the differential heating of the land and the ocean and the resulting pressure gradient (ii) the impact of the rotation of the planet and (iii) moist processes in the atmosphere*. Thus monsoon is a complex, nonlinear phenomenon involving atmosphere, oceanic, and land-based processes (*Webster*, 1987). Ocean can store energy more efficiently than land and therefore retain heat longer than land mass. So over large ocean basins, seasonal changes in tropical circulation are limited to minor latitudinal shifts and small variations in the intensity. However, over land important seasonal temperature and pressure changes take place that produce seasonal reversal of pressure gradient force. As a result there are major seasonal wind reversals (Figures 1.2 and 1.3).

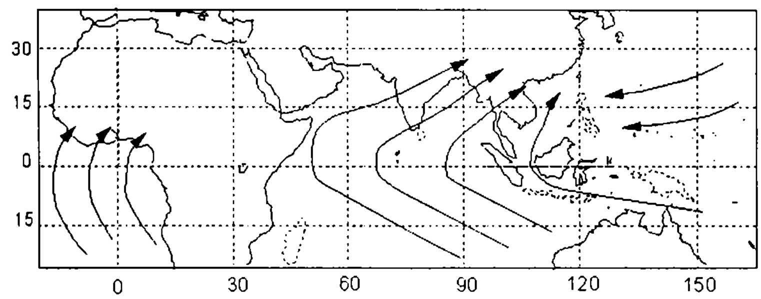

Figure 1.2: Surface wind flow during northern hemispheric summer (*Webster*, 1987)

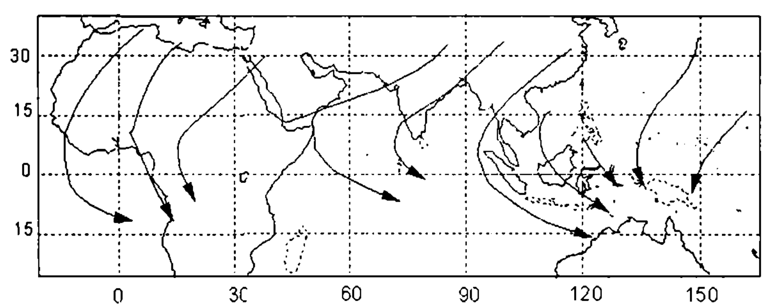

Figure 1.3: Surface wind flow during northern hemispheric winter (*Webster*, 1987)

Because of the Coriolis force due to earth's rotation, air in the monsoon currents moves in curved paths. Inter–hemispheric difference in the direction of the Coriolis force also cause winds to change direction as they cross the equator. As moist warm air rises over summer–time heated land surfaces, the moisture eventually condenses, releasing energy in the form of latent heat of condensation. This extra heating increases the summer land–ocean pressure differences to a point higher than they would be in the absence of moisture in the atmosphere. Moisture processes therefore add to the vigour of the monsoons (*McGregor and Nieuwolt*, 1998).

Monsoon is essentially an annual oscillation in the state of the atmosphere. The relationship among the general mechanisms that generate the monsoons, the seasonal climate cycle and the annual monsoon cycle are shown in Figure 1.4. In the transitional months between the southern and northern hemispheric summers, the Inter Tropical Convergence Zone (ITCZ) is located in the equatorial regions (Figure 1.4 (a)), where maximum surface heating can be found. At this stage, the northern hemispheric tropical–subtropical latitudes are begin to warm up and weak vertical motion is present. The northern hemispheric Hadley cell still predominates at this stage. With the northward movement of the Sun in May to June, the heating of northern tropical land masses intensifies, as does the vertical motion over these land masses (Figure 1.4 (b)). At this time, precipitation belts associated with ITCZ have moved north of the equator, signaling the onset of the summer monsoon.

From June to July, sensible heat input at the surface is close to a maximum, as also the vertical motion and atmospheric moisture over the northern hemispheric tropical land masses (Figure 1.4 (c)). Maximum values of pressure gradient force have also been attained at this stage and the monsoon reaches its maximum intensity with maximum amount of precipitation. By September, surface temperature has decreased markedly with maximum insolation positioned close to that of the April position. The structure of the monsoon at this time of the year is therefore similar to that of April (Figure 1.4 (d)). September heralds the cessation of the northern hemispheric monsoon season. By December, the southern hemispheric wet season is well under way as precipitation belts associated with ITCZ have moved south of the equator (Figure 1.4 (e)) (*McGregor and Nieuwolt*, 1998).

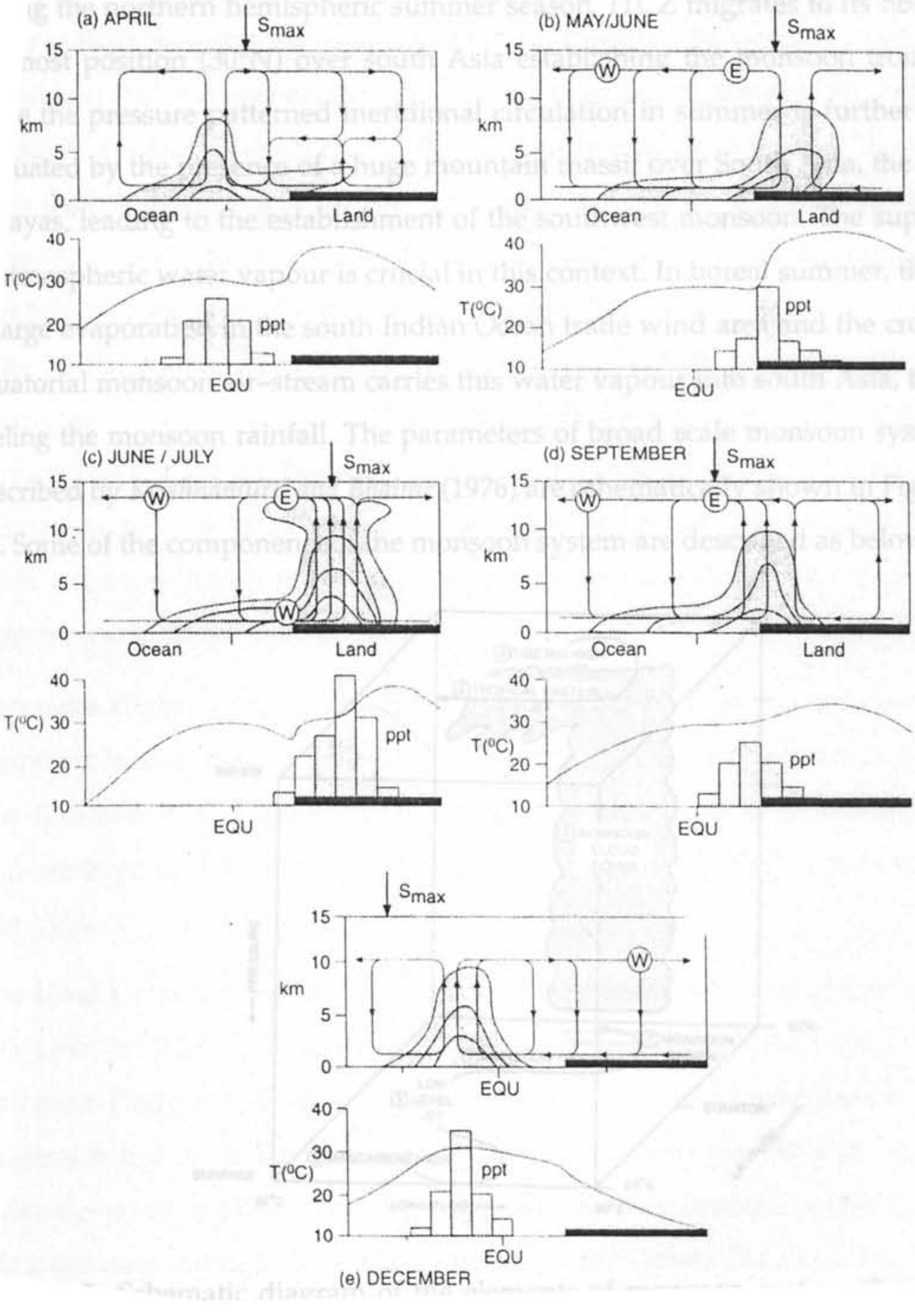

Figure 1.4: The annual monsoon cycle (*Webster et al.*, 1998)

1.4 Asian Summer Monsoon

During the northern hemispheric summer season, ITCZ migrates to its northern most position (30°N) over south Asia establishing the monsoon trough. There the pressure patterned meridional circulation in summer is further accentuated by the presence of a huge mountain massif over South Asia, the Himalayas, leading to the establishment of the southwest monsoon. The supply of atmospheric water vapour is crucial in this context. In boreal summer, there is large evaporation in the south Indian Ocean trade wind area and the cross–equatorial monsoon air–stream carries this water vapour into south Asia, thus fueling the monsoon rainfall. The parameters of broad scale monsoon system described by *Krishnamurti and Bhalme* (1976) are schematically shown in Figure 1.5. Some of the components of the monsoon system are described as below.

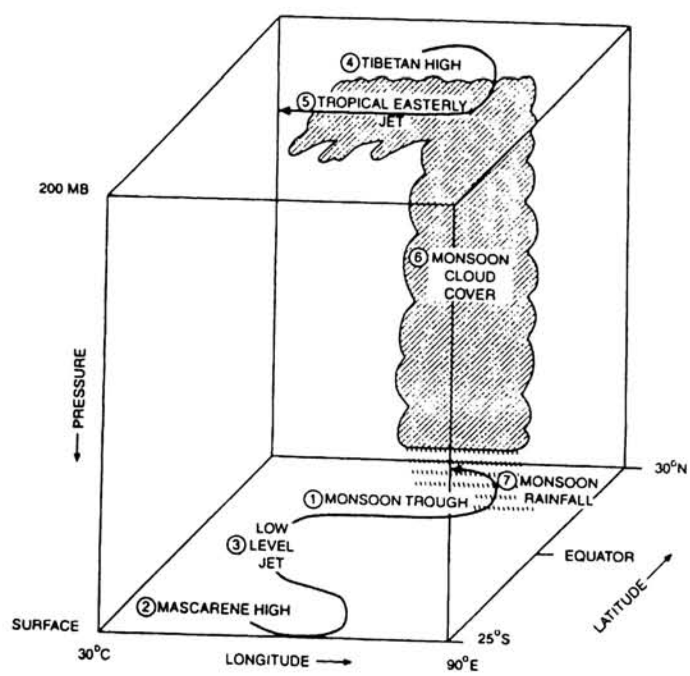

Figure 1.5: Schematic diagram of the elements of monsoon system (*Krishnamurti and Bhalme*, 1976)

Monsoon Trough : During the summer monsoon season, at the surface there is a trough extending southeastwards from the heat low over Pakistan upto Gangetic West Bengal. It is called the Monsoon Trough. The heat low over central parts of Pakistan and neighboring regions is generally linked to the region of maximum heating which are out of reach of the maritime air mass. The heat low is shallow extending upto about 1.5 km and above it is a well marked ridge extending to the upper troposphere, which is part of the subtropical high pressure belt. The monsoon trough is regarded as a part of the equatorial trough of the northern summer in Indian longitudes. The position of this trough line varies from day–to–day and has a vital bearing on the monsoon rains. No other semi–permanent feature has such a control on monsoon activity. When monsoon trough is south of its normal position, we get the active monsoon phase. When monsoon trough moves to the foot hills of the Himalayas we get the break monsoon phase (*Rao*, 1976).

Mascarene High : This is a high pressure area south of the equator near the Mascarene Islands east of Madagascar. The centre of this anticyclone is located near 30°S and 50°E. Variation in the location and strength of the Mascarene High are important in relation to the summer monsoon circulation and rainfall over India.

Low Level Cross Equatorial Jet : This is the well known northern summer Low Level Jet (LLJ). LLJ has its origin in the south Indian Ocean north of the Mascarene High as an easterly current, it crosses the equator in a narrow longitudinal belt close to the east African coast as southerly current with speeds at times even as high as 100 knots, turns into Arabian Sea as a westerly current and passes through India to the western Pacific Ocean. The LLJ is the main conduit for transporting moisture generated over the Indian Ocean into the monsoon area.

Tibetan High : This is a large anticyclone known to have its largest amplitude near 200 hPa during the northern summer months. Between 500 hPa and 200 hPa, the high pressure belt is well to the south of Tibet during June and September and over Tibet during July and August. The combination of the Tibetan High in the upper troposphere and the Monsoon Trough at sea level is accompanied by warm hydrostatic tropospheric columns over the northern India and over the foot hills of the mountains. The warm troposphere is another important feature of broad-scale monsoon system.

Tropical Easterly Jet : A strong easterly flow of air south of the Tibetan High, a tropical easterly jetstream (TEJ), develops in the upper troposphere at around 150 hPa during the monsoon season. This jet has winds of 80–100 knots and has its axis approximately around 10°N and 100°E. Normally, TEJ is in an accelerating stage from south China Sea to south India and decelerates to the west. Upper divergence associated with TEJ is regarded as favourable for convection upstream of 70°E. Subsidence occurs downstream. The fact that the tropical easterly jet only occurs in the summer suggests that its development is related to the seasonal cycle of surface heating and convective heating in the area over which jet lies.

Monsoon Cloudiness and Rainfall : The Indian longitudes are characterized by a large seasonal excursion of the maximum cloud zone (MCZ). It is found that during June–September there are two favourable locations for MCZ over these longitudes. On a majority of days the MCZ is present north of 15°N in the monsoon zone. Often a secondary MCZ occurs in the equatorial region between the equator and 10°N. The monsoon MCZ gets established by northward movement of the MCZ occurring over the equatorial Indian Ocean in April and May. The secondary MCZ appears intermittently, and is characterized by long spells of persistence only when the monsoon MCZ is absent. The monsoon MCZ cannot remain for more than a month without re–establishment by the the secondary MCZ (*Sikka and Gadgil*, 1980). The rainfall shows similar temporal variation like the cloudiness in MCZ

All these components are connected through the monsoon Hadley circulation. *Krishnamurti and Bhalme* (1976) found that interaction among solar radiation, conditional instability of the monsoon atmosphere and the cloudiness in the monsoon trough region lead to a quasi biweekly oscillation of the components of the monsoon system.

1.4.1 Onset and Withdrawal of Monsoon

Onset of Asian Summer Monsoon is one of the most dramatic seasonal transition exhibited by the atmosphere. The *onset of monsoon* refers to the burst of rains over Kerala on the southern tip of the peninsula and is associated with the commencement of organised monsoon rainfall over Indian region and the establishment of the monsoon trough. The onset phase of monsoon commences in late May/early June. The mean date of onset over Kerala is 1 June, with standard deviation of 8 days. It takes about six weeks from the time of onset over Kerala for the monsoon rains to cover the entire country (India). Figure 1.6 shows the normal onset dates of southwest monsoon over India.

The nature of the onset of the monsoon and the significant factors in the energetics have been extensively studied (*Krishnamurti et al.*, 1981; *Pearce and Mohanty*, 1984; *Ananthakrishnan and Soman*, 1988; *Joseph et al.*, 1994). The main circulation features of the onset can be summarised as follows *(i) formation and northward movement of a cyclonic system (onset vortex) in the southwest Arabian Sea in many years; (ii) strengthening and deepening of westerlies in the lower troposphere and organisation and strengthening of easterlies in the upper troposphere over peninsular India; (iii) the subtropical westerly jet over north India tending to weaken and shift*

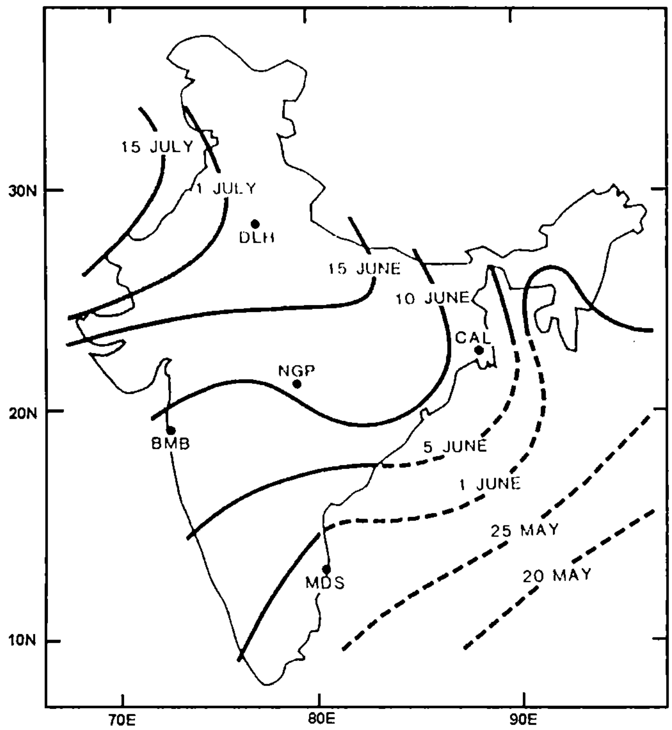

Figure 1.6: Mean dates of onset of the summer monsoon over India. Broken lines denote isolines based on inadequate data. (From India Meteorological Department)

northward; (iv) persistent heavy cloudiness over the southeast Arabian Sea (*Soman and Krishnakumar*, 1993).

On the synoptic scale, the onset of monsoon over Kerala is associated with the genesis and movement of either a depression/cyclone in the Indian seas, or a low pressure area off the coast of Kerala or mid–tropospheric disturbances in the east–west trough zone at the tip of the peninsula. The onset involves the genesis of a well–defined onset vortex (*George and Mishra*, 1993) in about 50% of the years. It usually forms on the cyclonic shear side of the low–level jet in the lower troposphere over eastern Arabian Sea and subsequently its motions are more westward, usually toward the Arabian coast, where it is known to dissipate. It has been noted to first form in the middle troposphere over the eastern

Arabian Sea and subsequently cyclogenesis occurs in the lower troposphere (*Krishnamurti*, 1985).

Krishnamurti and Ramanathan (1982) showed that convective heating has by far the largest effect in determining the structure and strength of the divergent wind. They also demonstrate that unless the moisture is present in the initial profile, no onset occurs in the model. *Sikka* (1980) suggested that passage of deep mid-latitude westerlies across the Mozambique channel in late May triggers the surge in cross–equatorial flow leading to the onset of the monsoon. Analysis of Outgoing Longwave Radiation (OLR) data by *Joseph et al.* (1994) has shown that around the time of monsoon onset over Kerala (MOK), an active belt of convection extends from the south China through the Bay of Bengal, with suppressed convection in the equatorial trough region of the western North Pacific. They also point out that the delayed MOK is associated with El Niño particularly in its year +1. Of the 22 years between 1870–1989 when MOK was delayed by 8 days or more, 16 cases were associated with moderate or strong El NiñoȮf the 13 strong El Niños during the same period, 9 were associated with moderate to large delays in MOK.

While the onset of the monsoon over India is invariably gradual, its withdrawal is relatively rapid. Withdrawal from the northwestern parts of India normally commences by September 1 (Figure 1.7). Cooling of the 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. As these factors vary from year to year, the withdrawal of the monsoon undergoes interannual variation. In general, the withdrawal of the monsoon from western Rajastan, Haryana, and Punjab takes place during the first fortnight of September, and withdrawal from most of the remaining

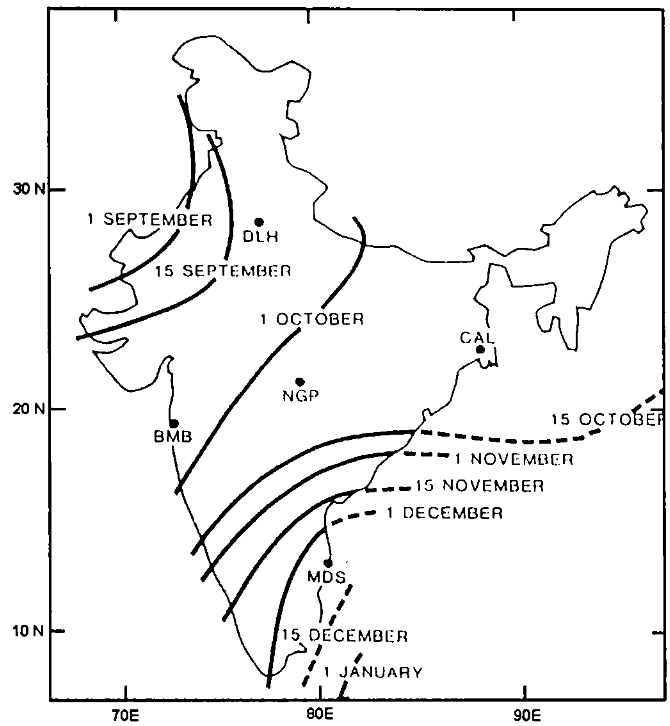

Figure 1.7: Mean dates of withdrawal of the summer monsoon from India. Broken lines denote isolines based on inadequate data. (From India Meteorological Department)

parts of the country occurs during the period mid–September to mid-October (*Mooley and Shukla*, 1987).

1.4.2 Annual Cycle

During Asian summer monsoon season, there are two principal rainfall maxima. The first lies around 15°N and contains two principal sub–maxima: over the Bay of Bengal and in the eastern Arabian Sea next to Ghat mountain range. The second and weaker maximum lies close to and south of the equator and extends from 60°E to 110°E. The northern precipitation zone matches the location of the monsoon trough. The southerly maximum coincides with the weaker surface pressure trough lying just to the south of the equator. During winter the precipitation patterns extend as a broad zone from the western Indian Ocean to the date line to the south of the equator and is located closer to the equator than during the summer. The heating gradients in both seasons associated with precipitation and radiative processes are largest on the planet, dominating the planet's annual cycle and defining circulation structures on a planetary scale (*Webster et al.*, 1998).

The major precipitation zone in northern summer is in a tilted position with its center in the Indian region (between 15°N and 20°N) while its position in the western Pacific remains around 5°N. This mean displacement of the precipitation zone over the Indian region represents a strong asymmetric heat source. The heating gradients associated with such a heat source drives a regional Hadley circulation with ascending motion around 20°N, and descending motion from the equator to the southern hemisphere subtropics. The precipitation maximum and hence the ascending part of the Hadley circulation during the northern summer over the rest of the globe is closer to the equator and centered around 5°N. It was first noted by *Schulman* (1973) that the regional monsoon Hadley cell with reverse meridional circulation in the Indian monsoon region is strong enough to make the zonally averaged Hadley cell appear to be very week during the northern summer. The importance of the regional Hadley circulation in the monsoon rainfall was recognised by *Joseph* (1978) who showed that the mean meridional wind at 150 hPa in June, July and August over India was related to the seasonal monsoon precipitation. However, there is also an east–west Walker circulation with major ascending motion in the equatorial western Pacific and Indonesia and subsidence over the equatorial Indian Ocean. Thus, the Indian monsoon may be viewed as a superposition and interaction between a regional Hadley circulation and a planetary-scale Walker circulation (*Goswami*, 1994; *Goswami et al.*, 1999).

Through out the annual cycle, the Sea Surface Temperature (SST) in the Indian Ocean undergoes major changes. The SST is the most important characteristics of the coupled ocean–atmosphere system and determines, to a large degree, the manner in which the ocean and the atmosphere interact. The atmosphere feels the SST but the future SST depends on the surface thermal structure and particularly on the heat content of the upper ocean. A warm SST anomaly without warm heat content will quickly be eliminated by the atmosphere. On the other hand, warm SST region with warm heat content can be very persistent to the climate system. The north Indian Ocean SST reaches maximum in spring and early summer (more substantially in the Arabian Sea). With the arrival of the monsoon due to mixing and enhanced Ekman transports SST seems to fall rapidly. Negative feedbacks occur between the atmospheric component of the monsoon and the ocean.

1.4.3 Interannual Variability

The year to year variation of the Asian monsoon is one of the strongest signals of the earth's climate variability. The mean rainfall over Indian peninsula (June–September) is about 852 mm and the standard deviation of the seasonal mean is about 84 mm (*Parthasarathy et al.*, 1994). A large fraction of interannual variability is determined by the slowly varying surface boundary conditions such as SST, surface albedo and soil moisture (*Charney and Shukla*, 1981). Several workers have shown that there is a significant relationship between drought in the Indian summer monsoon and El Niño-Southern Oscillation (ENSO) (*Sikka*, 1980; *Rasmusson and Carpenter*, 1983; *Shukla and Paolina*, 1983). *Webster et al.* (1998) point out that nearly all El Niño years are drought years in India but not all drought years correspond to El Niño years. Although the relationship is not fully understood, it is clear that the monsoon and ENSO are related in some fundamental manner.

Soman and Slingo (1997) showed that the modulation of the Walker circulation (additional subsidence) is the dominant mechanism by which El Niño weakens the Asian summer monsoon for the years 1983 and 1984. However, the delayed onset during El Niño may be associated with the complementary cold SST anomalies in the western Pacific that delay the northward transition of the Tropical Convective Maximum (TCM). They further suggest that during cold events it is primarily the warm SST anomalies in the western Pacific that enhances the TCM and lead to an early onset and stronger monsoon. Although numerous studies have focused on ENSO links to the Indian monsoon, *Lau and Wu* (1999) mentioned that India is not a major center of action in the dominant coupled precipitation–SST relationships. *Arpe et al.* (1998) studied the differences between 1987 and 1988 summer monsoons and suggest that, while large scale dynamics over India are mainly governed by Pacific SST, the variability of precipitation over India is impacted by a number of other factors including SST anomalies over the northern Indian Ocean, soil wetness, initial conditions, and the quasi–biennial oscillation. They show that the two direct effects of El Niño are to reduce precipitation over India and reduce the surface winds over the Arabian Sea. They suggest that the latter leads to an increase in SST and more precipitation over India acting to counteract the direct effect of El Niño .

Another school of thought believe that the monsoon in turn influence ENSO (*Yasunari*, 1990). Several recent studies have examined the possibility of modification of the ENSO characteristics by the summer monsoon by simple coupled models. *Wainer and Webster* (1996) argued that the interannual variation of the summer monsoon may contribute to irregularities of El Niño. *Chung and Nigam* (1999) showed that, based on results from an intermediate ocean–atmosphere coupled model, that monsoon forcing may increase the frequency of occurrence of El Niño. *Lau et al.* (2000) suggested that boreal spring warming in the north Arabian Sea and subtropical western Pacific may play a role in the development of strong South Asian monsoon. *Kumar et al.* (1999) showed that the inverse relationship between ENSO and Indian Summer Monsoon, that was clearly evident before 1980, weakened considerably in recent decades. They suggest that this is associated with a southward shift in Walker Circulation, and increased Eurasian surface temperature during winter and spring seasons. Using NCEP/NCAR reanalysis data and Atmospheric General Circulation model *Goswami and Jayavelu* (2001) have shown that the Indian Monsoon, by itself, does not produce significant surface wind anomalies in the the equatorial Pacific either during or following the monsoon season and thus the Indian monsoon by itself is unlikely to influence the ENSO in a significant way.

The interannual variability of monsoon rainfall shows a biennial variability during certain periods of the data record. This biennial oscillation, referred to as tropospheric biennial oscillation (TBO) is reported 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 rainfall TBO appears as a part of the coupled ocean–atmosphere system of the monsoon regions, increasing rainfall in one summer and decreasing in the next.

A strong biennial tendency has also been noted in ENSO cycles for some time (*Lau and Sheu,* 1988; *Rasmusson et al.,* 1990). Except for the different time scales, the evolutionary features of the biennial oscillation in SST, sea level pressure, wind and precipitation are very similar to that of ENSO. In the last several years, there have been increasing evidences showing the presence of the TBO in the monsoon regions (*Yasunari,* 1990; *Shen and Lau,* 1995; *Meehl,* 1997). Recently a number of theories have emerged suggesting that the TBO may be related to the air–sea interaction in the Asian Summer Monsoon region, modified by coupled ocean–atmosphere processes (*Meehl,* 1997; *Chang and Li,* 1999). Recent studies have also suggested that strong monsoon-ENSO interactions may result in a strong biennial tendency in ENSO cycles in the form of rapid development of La Niña approximately one year after an El Niño (*Lau et al.*, 2000; *Lau and Wu*, 2001). *Kim and Lau* (2000) investigated the mechanism of the quasi-biennial tendency in ENSO-monsoon coupled system using an intermediate coupled model. They found that the strong coupling of ENSO to monsoon wind forcing over the western Pacific is the key mechanism.

1.4.4 Intraseasonal Variability

Among the many time scales of Asian Monsoon variations, the intraseasonal variation is most distinctive. Observational evidence shows the existence of three different quasi-periodic oscillations with periods of 4–6 days, 10–20 days and 30–60 days (*Krishnamurti and Bhalme*, 1976). The 4–6 day oscillations are largely observed in monsoon trough region. The dynamical system associated with the 4–6 day scale is the monsoon disturbance. A dominant characteristic of intraseasonal fluctuations during summer in the monsoon region is the active–break cycles of precipitation with periods of 10–20 days or 30–60 days . Active spells of the summer monsoon region are associated with an intense trough over India with heavy rainfall over monsoon trough zone. During the break monsoon condition, the monsoon trough moves northward to the foot of the Himalayas, resulting in decrease in rainfall over much of India but enhanced rainfall in the far north and south (*Ramanadham et al.*, 1973). These anomalies are large scale and extend across the entirety of South Asia.

The active–break cycles are linked to observed northward propagation of convection from Indian Ocean on to the Asian subcontinent in summer (*Kesavamurty 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 Subrahmanyam*, 1982; *Lau and Chan*, 1986; *Wang and Rui*, 1990; *Gadgil and Asha*, 1992). Similar northward propagation is also found over the west Pacific during northern summer (*Murakami*, 1984; *Wang and Rui*, 1990). *Hartman and Michelsen* (1989) analysed 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) 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 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. 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).

Wang and Rui (1990) classified the intraseasonal movement of convection anomaly in three categories. They are (i) eastward (ii) independent northward and (iii) westward. The eastward propagating convection anomaly exhibit three major tracks: (a) equatorial eastward from Africa all the way to the mid-Pacific, (b) first eastward along the Indian Ocean, then either turning northeast toward the northwest Pacific or southeast toward southwest Pacific at the maritime continent, and (c) eastward propagation along the equator with split center(s) moving northward in the Indian and/or west Pacific Oceans. Independent northward propagation which is not associated with eastward propagation is found over two longitude sectors: the Indian monsoon region and western Pacific monsoon region. The mechanism responsible for meridional propagation may differ from that for the eastward propagation. However, climatologically, the most active period of the MJO (at least in its equatorial manifestation) is in the boreal fall and winter (*Webster et al.*, 1998).

A number of theories have been proposed to explain the northward movement of convection during summer. *Lau and Peng* (1990) showed that the interaction of tropical diabatic heating associated with the 30–60 day oscillation and the summer monsoon mean flow induces the development of westward propagating synoptic scale cyclonic vortices over the monsoon region leading to the active and break phases of the monsoon. They suggest that the rapid development of these disturbances, and associated weakening of equatorial convection *via* changes in the low–level moisture convergences accounts for the observed northward migration of the monsoon trough. *Webster* (1983); *Srinivasan et al.* (1993) emphasized the important role of land surface heat fluxes in the boundary layer that destabilize the atmosphere ahead of the ascending zone, causing a northward shift of convective activity. *Goswami and Shukla* (1984) suggested that the northward propagation is due to a convection–thermal relaxation feedback where in the convective activity increases static stability while dynamic and radiative relaxation decreases the moist static stability, bringing the atmosphere to a convectively unstable state. Based on results of a modelling study of summer ISO's, *Wang and Xie* (1997), described the northward *propagation* as a convection *front* formed by the equatorial Rossby waves emanating from the equatorial convection. *Krishnan et al.* (2000) suggest that monsoon breaks are initiated by rapid northwest propagating Rossby waves emanating from convectively–stable anomalies over the Bay of Bengal.

Using principal oscillation pattern (POP) technique *Annamalai and Slingo* (2001) describes the origin and propagation of 15 day and 40 day mode of intraseasonal oscillations. The 40 day mode originates and intensifies over the equatorial Indian Ocean and it has poleward propagation on either side of the equator, as well as eastward propagation into the equatorial west Pacific. The Rossby waves emanating from the west Pacific appear to be responsible for the northwestward propagation of convection. The 15 day mode originates over the equatorial west Pacific, associated with westward propagating Rossby waves, amplifies over the northwest tropical Pacific and modulates both the continental and oceanic TCZs over the Indian longitudes. By separating summertime ISOs based on their zonal propagation characteristics in the Indian Ocean and the western Pacific Ocean, *Lawrence and Webster* (2002) have shown that the eastward propagation of convection along the equator is a fundamental feature of the majority of summertime ISOs. The eastward propagation appears to be directly related to the northward movement of convection that is associated with active and break cycles of precipitation across India.

There is some evidence that interannual variations of intraseasonal oscillation activity may influence monsoon strength. Several modelling studies show that a significant fraction of the interannual variation of the seasonal mean Indian summer monsoon is governed by internal chaotic dynamics (*Goswami et al.*, 1998; *Harzallah and Sadourny*, 1995; *Rodwell and Hoskins*, 1995). *Goswami and Ajayamohan* (2001) have shown that the intraseasonal and interannual variations are governed by a common mode of spatial variability. The spatial pattern of standard deviation of intraseasonal variability of low-level vorticity and spatial pattern of the dominant mode of intraseasonal variability of the low level winds are similar to that of interannual variability. In another recent study *Lawrence and Webster* (2001) have shown that intraseasonal activity–Indian monsoon relationship is essentially independent of the ENSO–Indian monsoon relationship.

1.5 Low Level Jetstream

Low Level Jetstream (LLJ) according to a definition suggested by *Reiter* (1961) *should have marked gradients of wind speed in the horizontal and vertical.* There are several places where strong low–level currents are observed. 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). The following geographical locations are favourable for the occurrence of these LLJs.

1. *Slopes of the mountains parallel to the anti–cyclonic flow around the sub–tropical anti–cyclones; for example, low–level jet over west central USA, along the Peru coast in South America and along Namibian coast in South African continent.*
2. *Narrow mountain gaps, like Marsabit (North Kenya) Jet stream.*
3. *North–south oriented continental coasts near cross–equatorial flow, e.g., East–African LLJ during Asian summer monsoon.*

Bunker (1965) using Aircraft observations of wind in the Arabian Sea during the International Indian Ocean Expedition (IIOE) traced a LLJ with large vertical wind shears off Somalia and across the Central parts of Arabian Sea. He showed that monsoon winds attained a speed of 50 knots in the southwestern parts of the Arabian Sea at the top of a 1000 meter layer of air cooled by contact with the upwelled water off the Somali–Arabian coasts.

Analysing the wind data of five years collected by the radiosonde/rawinsonde network of India, *Joseph and Raman* (1966) established the existence of a westerly low level jet stream over peninsular India with strong vertical and horizontal wind shears (Figure 1.8). This jet is seen over peninsular India on many days in the month of July with core at about 1.5 km above mean sea level and core speeds of the order of 40-60 knots. It showed persistence of a few days at one latitude. South to north movement of the jet core was also reported. *Findlater* (1969a) found that the Asian summer monsoon LLJ has its origin in the south Indian Ocean north of the Mascarene High as an easterly current, it crosses the equator in a narrow longitudinal belt close to the east African coast as southerly current with speeds at times even as high

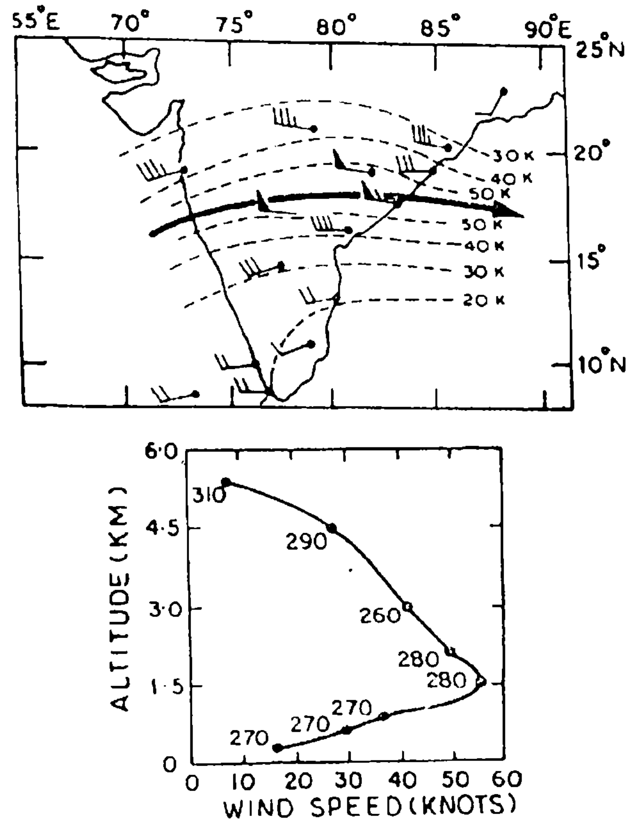

Figure 1.8: Low Level Jet stream over peninsular India at 00 UTC on 9 July 1961. (a) winds and isolines of wind (magnitude) in knots at 850 hPa level. LLJ axis is marked by a thick line. (b) vertical profile of wind speed at Visakhapatnam (17.1°N, 83.3°E) on the jet axis . The three digit numbers marked are the directions of wind. (adapted from *Joseph and Raman* (1966))

as 100 knots, turns into a westerly current over the Arabian Sea and passes through India. This jet according to their computations accounts nearly for half the inter–hemispheric transport of air in the lower troposphere. Using monthly mean winds *Findlater* (1971) showed that the LLJ splits into two branches over the Arabian Sea, the northern branch intersecting the west coast of Indian near 17°N, while the southerly branch passes eastward just south of India. Figure 1.9 shows the monthly mean airflow at the 1.0 km level for August.

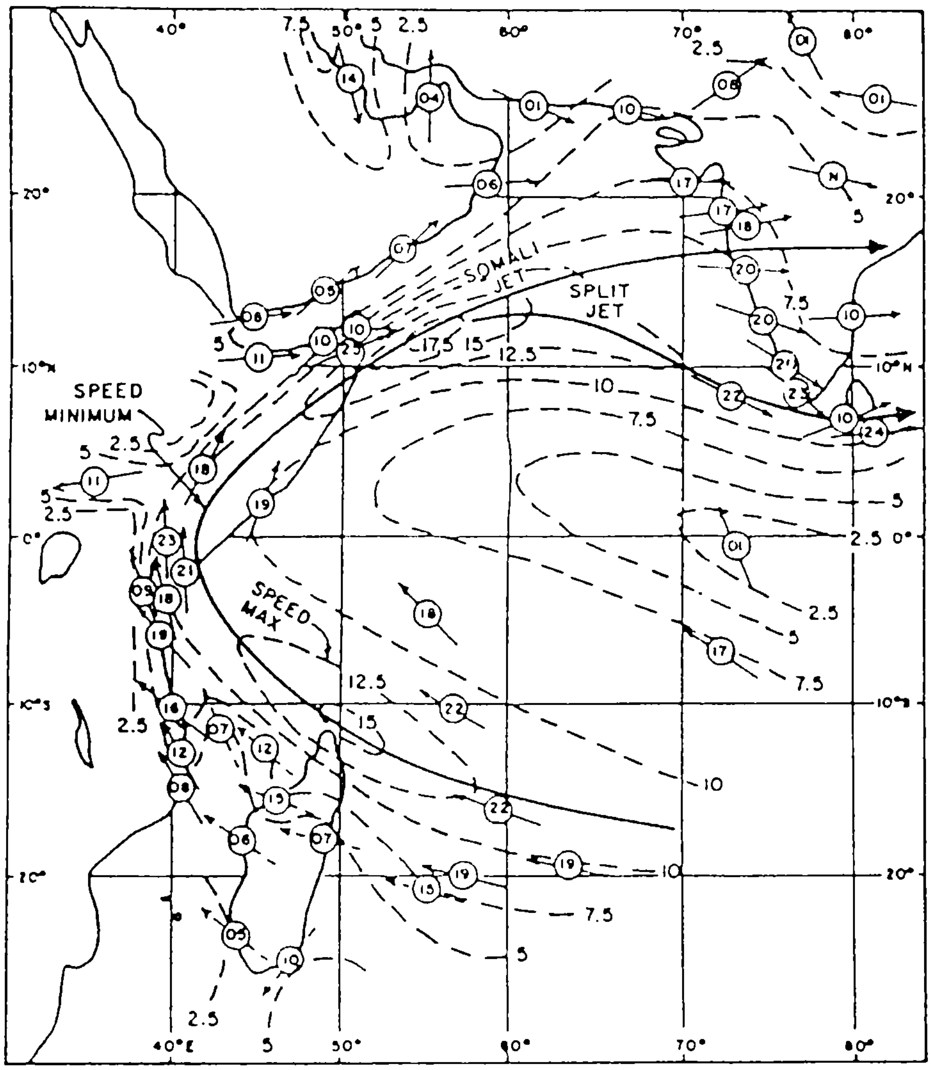

Figure 1.9: Wind field at 1 km for August over the Indian Ocean from *Findlater* (1971). Thick lines marked are the LLJ axes. Isotachs in ms^{-1} are shown as broken lines.

Krishnamurti et al. (1976) studied the LLJ using a single–level, time–dependent, primitive equation model with bottom topography and a prescribed flow on the eastern boundary. They concluded that the broad–scale low–level jet is forced by the differential heating over India and barotropic instability is a possible mechanism for the splitting of the jet. Other one layer models such as those of *Anderson* (1976); *Hart* (1977); *Bannon* (1979, 1982) have indicated how a cross equatorial jet can be formed and have described some of the dynamics involved. *Anderson* (1976) invoked the effects of lateral friction along the mountains located at the East African coast and showed that a balance between the β effect and lateral friction leads to a reasonable strength of the East African Low Level Jet and a location at a reasonable distances away from the mountains. *Hart* (1977) presented a number of analytical models in which the stratification of the observed atmosphere was represented by a free surface under the influence of a reduced gravity. An impermeable western wall barrier simulated the effects of the East African Highlands, and a realistic zonal inflow was prescribed at the eastern boundary. This simplified picture of the observed situation represented the idea that it is the inversion that forces the incident flow to go around rather than the mountain barrier. One of the models of *Hart* (1977) represented the idea of potential vorticity conservation and the advection of the potential vorticity across the equator resulting in the formation of a low level jet in the presence of western boundary mountains.

With a fine–mesh (200m) vertical resolution numerical model *Krishnamurti and Wong* (1979) studied the planetary boundary layer dynamics of the low level monsoonal flow over the Arabian Sea. They concluded that the meridional motion of air across the equator from the Southern to the Northern Hemisphere toward lower pressure results in an acceleration and an enhancement of the horizontal advective terms in the balance forces. As an extension of this study *Krishnamurti et al.* (1983)–using a three dimensional model that removes the restriction of symmetry–modeled the jet's vertical structure and simulated more realistically its curvature and the position of its maximum strength over the Arabian Sea. The quadratic bulk formulation drag coefficient used was dependent on orographic height. Away from the equator in both hemispheres, the modeled momentum balance was found to be mainly geostrophic at the level of 1 km, with surface friction producing an Ekman balance at 200 m. Near the equator, the balance was cyclostrophic above with friction again important near the ground. An important feature of their results was that the confluence associated with the intertropical convergence zone (ITCZ) over the Arabian Sea did not appear to be accounted for by geostrophic flow. The imbalance between Coriolis and pressure gradient forces to the south of the ITCZ tended to accelerate air parcels northward and so produce the confluence.

Krishnamurti and Ramanathan (1982) showed that the overall development and strengthening of the low–level zonal flow during onset is highly sensitive to the large scale field of differential heating. *Joseph et al.* (1994) suggested that the intensification of the Low Level Jetstream occurs only after the ITCZ over the Indian Ocean has moved north of the equator. A time depended primitive equation model with specified zonal flow, mountains and diabatic heating was used to study the LLJ by *Hoskins and Rodwell* (1995); *Rodwell and Hoskins* (1995). The east African highlands and a land–sea contrast in surface friction are shown to be essential for the concentration of the cross equatorial low level flow into LLJ. They found that surface friction and diabatic heating provide mechanisms for material modification of potential vorticity (PV) of the flow and both were found important for the maintenance of the LLJ. The study identified the strong sensitivity of the LLJ to changes in convective heating over Indian Ocean. When there is a little modification of the PV, the LLJ turns anticyclonically over the Arabian Sea and the flow tends to avoid India.

Arpe et al. (1998) explained the teleconnection between the Somali Jet over the Arabian Sea and El Niño/ La Niña. During El Niño the convection area over the Indian subcontinent, which is the attractor for the Somali Jet, would move eastward and consequently subsidising air would replace rising air and the intensity of the rising air would be reduced. *Halpern and Woiceshyn* (1999) defined the onset of the Somali Jet to be the date when National Aeronautics and Space Administration (NASA) Scatterometer (NSCAT) surface wind speeds off Somalia reached 12 ms^{-2} over a 3°× 3°region in the western Arabian Sea for at least six days. The minimum duration was about three inertial periods, which is the approximate time for development of Ekman currents. The zonal component of wind direction must be eastward. Using Special Sensor Microwave Imager (SSM/I) wind data *Halpern and Woiceshyn* (2001) studied the interannual variations of the Somali Jet in the Arabian Sea during 1988-99 linked with El Niño and La Niña episodes. According to them the average date of Somali jet onset was two days later in El Niño events in comparison with La Niña conditions. Monthly mean strength of the Somali Jet 0.4 ms^{-1} weaker during El Niño episodes than during La Niña intervals. They also reported that the monthly mean intensity of the Somali Jet is above (below) normal, there is an excess (deficit) of rainfall along the Indian west coast.

1.6 Monsoon Depressions

The monsoon season in summer over Asia is punctuated by intermittent emergence and subsequent decay of well-defined, synoptic scale propagating disturbances. One of such synoptic scale tropical disturbance is the Monsoon Depression (MD). These are the important rain producing disturbances of Indian southwest monsoon. Climatology shows that monsoon depressions generally form in the northern portion of the Bay of Bengal and move west-northwestwards (*Rao*, 1976; *Sikka*, 1977). The general features of monsoon depressions were studied by several workers (*Koteswaram and George*, 1958; *Pisharoty and Asnani*, 1957; *Sikka*, 1977; *Mak*, 1987). The horizontal wavelength of a depression is typically about 2000 km and generally extend vertically up to about 10 to 12 km. The radial gradient of surface pressure ranges from 2 to 5 hPa per 100 km and a surface wind of 8 to 16 ms^{-1}. The disturbance has cold core in the lower troposphere and warm core above (*Sikka*, 1977). Maximum rainfall and cloudiness is found southwest of the depression center (*Pisharoty and Asnani*, 1957). *Lindzen et al.* (1983) ascribed the growth of a Bay of Bengal monsoon depression to the horizontal shear flow instability mechanism. *Douglas* (1992a,b) investigated the structure and dynamics of the monsoon onset vortex and Bay of Bengal depression by special observing systems of MONEX 1979. They found many similarities between the structure of the mature onset vortex and the Bay of Bengal monsoon depression.The observed lower tropospheric positive vorticity tendency west of both depressions (indicating the direction of motion) was primarily a result of an excess of cyclonic vorticity generation by convergence over anticyclonic vorticity advection. Heat-budget calculations for the rain areas showed an approximate balance between warm advection and adiabatic cooling at 850 hPa, though diabatic heating was large above this level. For both depressions the region of maximum rainfall was coincident with the location of the maximum warm advection in the lower troposphere.

1.7 Heavy Rainfall Events Along West Coast of India

Weather during the monsoon period varies from one area to another and from one day to another over the same area. These variations are connected with synoptic patterns in surface and upper air. Formation of off–shore vortices in the trough in westerlies along the west coast of India and associated heavy rainfall is an important synoptic system during monsoon (*Rao*, 1976). Offshore vortices are mesoscale in character with linear dimensions of the order of 100 km or even less and their presence is detected by weak easterly winds at coastal stations. Notwithstanding their small dimension, they are effective in giving spells of very heavy rain in their vicinity. The peculiarity of the rainfall associated with these vortices is that the rainfall over the coast is heavier than that over hill stations a few kilometers east. Their normal duration is of the order of 1 to 3 days. The dynamics of these vortices has not been examined in much detail so far. Existence of these vortices has not been firmly established because of the lack of mesoscale data. They are suspected to be forming when the monsoon is normal or strong over the Arabian Sea. Available data show that they are very small both horizontally and vertically and can be located by coastal surface winds. For the forecasting of heavy rainfall the approach was mainly a statistical correlation, lacking the proper understanding of their dynamics (*George*, 1956; *Mukherjee*, 1980).