CHAPTER 1

INTRODUCTION AND OVERVIEW
1.1 GENERAL

The summer monsoon rainfall over India contributes about 70-90% of the country’s total annual rainfall (Pant and Rupa Kumar, 1997) over different regions. India being an agricultural country, the seasonal monsoon rainfall significantly affects the socio-economic activities of the large population. Even though the contribution by agriculture towards the national economy is about 30%. Such events result in landslides, flash floods and crop damage that have significant impacts on society, economy, and the environment.

India being a land of diversities, climatic studies cannot consider India as a single unit (Normand, 1953). In addition to this, large differences in the spatial and temporal patterns of rainfall are observed over India. The variability in rainfall is made so complex by the presence of large differences in orography and landuse over the country, that the standard deviation of the seasonal rainfall is so different even over the homogenous regions and meteorological subdivisions.

That the large amount of rainfall obtained during the southwest monsoon season is due to the topographical features over the country is well known. The large accumulation of rainfall over the Western Ghats region is an example. The presence of the Himalayas and Tibetan plateau play significant role in the monsoon rainfall over India. Peninsular region of India is a peculiar region surrounded by ocean and with mountains over the landmass. Thus the variability and the role of topography on the climate of peninsular India need to be studied.

1.2 LARGE SCALE FEATURES AFFECTING THE INDIAN REGION

The main source of rainfall for the peninsular Indian region is the monsoon rainfall obtained during June-September months of a year. The components of the southwest monsoon are detailed below:

1.2.1 Monsoon

Monsoon denotes any annual climate cycle with seasonal wind reversal in tropical and subtropical regions. The term monsoon is derived from the Arabic word "mausam,"
which means season. The monsoon significantly changes the atmospheric circulation and associated precipitation over the region. The radiation absorbed by the oceans is transformed to different latitudes by ocean current whereas the radiation over landmass heats the earth surface, which results in the differential heating of oceans and land. Thus, during summer, the temperature gradients cause trade winds from winter hemisphere to cross the equator and move towards the low pressure area over the landmass. As the winds are southwesterly, the season is often called southwest monsoon season. During the winter season, low temperature and high pressure over continents makes the wind blow towards equator. Thus there is a seasonal annual reversal in wind direction. The winds are northeasterly during October-December months and hence known as northeast monsoon season. The southwest monsoon air mass is maritime and moist in great depth whereas northeast monsoon has dry continental air.

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

i. the prevailing wind direction shifts by at least $120^\circ$ between January and July;
ii. the average frequency of prevailing wind direction in January and July exceeds 40%;
iii. the mean resultant wind in at least one of the months exceeds $3\text{ms}^{-1}$ and;
iv. fewer than one cyclone-anticyclone alternation occurs every two years in either months in a $5^\circ$ latitude-longitude rectangle.

Thus, monsoon regions include almost half of the African continent, south and East Asia and northern Australia (Fig 1.1). About 55% of the world’s population is inhabited over these regions. The socio-economic security in many tropical countries are linked to the variation of monsoon cycle.

1.2.2 Causes of the monsoon

The three factors that account for the cause and existence of monsoons are as follows

i. **The land-ocean thermal contrast**: Only small variations are observed in the tropical circulation of large ocean basins, whereas this is entirely different over the landmass and adjacent seas. During summer, the warming over the
continents produces heat lows. Also, during summer, the trade wind from the southern hemisphere cross the equator and move towards the low pressure region over the landmass. These conditions found to reverse during the winter season and wind blow towards equator. Thus a semi-annual reversal in wind direction is experienced over the land-ocean, a characteristic of monsoon.

![Monsoon areas](image)

**Fig 1.1 Monsoon areas (shaded portions) from the criteria of Ramage 1971. (Figure from Monsoon Meteorology, Ramage 1971, Academic press, New York)**

ii. **The Moisture processes**: During Summer, as the moist air rises up and eventually condense, latent heat of condensation is released. This extra energy increases the land-ocean pressure differences, in the absence of moisture. Thus the moisture processes adds strength to the monsoon.

iii. **The Earth’s rotation**: Due to the rotation of earth on its own axis, a force called Coriolis force is produced. Therefore the monsoon current takes a curved path. The difference in the direction of Coriolis force in the two hemispheres causes the wind to change its direction as they move from one hemisphere to other.

### 1.2.3 The annual and seasonal monsoon cycle

The general mechanisms that produce the monsoon are the seasonal climate cycle and the annual monsoon cycle (*Webster, 1987*). These cycles and their relationships are illustrated in Fig 1.2. During the transitional months, between the summers of both hemisphere, the Inter Tropical Convergence Zone (ITCZ) is located over the equatorial region. ITCZ is a region of surface low pressure, rising air movements and convergence of air masses. Therefore, maximum surface heating is observed over this region and the
tropics and subtropics of northern hemisphere begin to warm. Weak vertical motions are present and the Hadley cell dominate with offshore flow of air. As the sun moves northward from May to June, the landmass gets heated up and strengthens the vertical motion. Also an increase in the moisture content is noticed as a result of the intensification of southern hemisphere Hadley cell and predominant wind direction is onshore. During this time the precipitation associated with ITCZ moves well to the north of the equator and causes the onset of monsoon over many areas.

During June-July, maximum surface sensible heat input is observed along with strong vertical motion, pressure gradient force and atmospheric moisture over the landmasses. The monsoon attains its maximum intensity with maximum precipitation and the air that crosses equator is subject to strong coriolis force. By September a pattern similar to that of April is attained and marks the end of wet season and the beginning of the dry season. By December, maximum heating and precipitation associated with ITCZ moves well south of the equator. Accompanied with this, the northern hemisphere
Hadley cell strengthens and zone of maximum heating attains its maximum southward position.

The monsoon exhibits large variations from the mean cycle described above. The weather disturbances, complex orography, tropical-extratropical interactions etc influence the monsoon. Changes in boundary conditions also affect the heat and moisture distribution of the atmosphere.

Three monsoon systems have been documented: Asian, African and Australian systems. The Asian monsoon system is divided into Indian and East Asian monsoon. The thesis focuses on the features of southwest monsoon. A brief description on the Indian monsoon systems is given below.

1.2.4 The Indian Summer Monsoon

The components of Indian summer monsoon as suggested in the work of Krishnamurthi and Bhalme (1976) is shown in Fig 1.3 and its descriptions follow:

1) **The monsoon trough over north India**: It is apart of equatorial trough of northern summer and extends from heat low over Pakistan to Gangetic west Bengal. Southwesterly winds to south and easterlies to the north of trough. It is a zone of horizontal wind shear and is associated with low pressure. Monsoon depressions move along the trough line.

2) **The Mascarene anticyclone system**: high pressure area at sea level of the equator in the Indian Ocean near Mascarene island, with its center located near 30S and 50E. There is a large airflow of air from this area. The intensification of this high strengthens the cross equator flow in the form of east African low level jet and becomes southwesterly wind reaching at maximum intensity in the summer months, it crosses the southern Arabian Sea and reach over central western and southern coast of India. Variation in the intensity of this jet is important in determining the rainfall over India.
3) **Low level cross-equatorial jet**- Strong low level currents located in the lowest 1-2 km of troposphere. They transport moisture and momentum. The core of the jet is at a height of 1.5km and speeds 12 to 15 m/s. Jet has strong horizontal and vertical shears. During summer the Jet splits into two branches at around 13°N, 60°E. These cross the south Arabian sea and reach the coasts of India. Variations in the intensity of this jet determine the rainfall amount over western India.

4) **The Tibetan high pressure system**- located more than 4500m above sea level. Tibetan anticyclone is a warm high located over Tibetan Plateau. By July it is well established at 200 hPa level Variation in the intensity and position of this high and its orientation are closely related to monsoon circulation over south Asia. East-west oriented Tibetan anticyclone at 500 hPa is related to the rainfall distribution over India.

5) **The Tropical Easterly Jet**- The outflow of air from the southern flank of Tibetan anticyclone gives rise to tropical easterly jet is a band of strong easterlies extending from southeast Asia across the Indian Ocean and Africa to the Atlantic, at a height of about 14 km. Jet axis found close to 14°N with core speed exceeding 40m/s but generally situated at around 9°N (Mokashi, 1974). Jet axis at 200-100 hPa presents over India peninsula from June to August and disappears by September.

Fig 1.3 Components of Indian summer monsoon system (Adapted from Krishnamurthi and Bhalme, 1976)
6) **Monsoon cloudiness** - Cloud cover is the manifestation of moist convective processes over the Indian subcontinent and it varies both in space and time. Maximum cloud cover formed in a belt running from western shore of Bay of Bengal to northern shore of Arabian Sea and minimum over foothills of Himalaya, south India. A reversal of the pattern is observed during the break condition.

7) **Rainfall**: Rainfall distribution closely follows the cloud distribution

After a brief overview of the main components of monsoon system, a description on the mean patterns of climatic elements over India region is given below.

### 1.3. REGIONAL SCALE FEATURES

#### 1.3.1 Pressure

![Fig 1.4. Spatial distribution of Sea level Pressure during June-September over India for the period 1951-2003. Unit is hPa.](image)

The Fig 1.4 shows a mean pressure pattern for the monsoon season. During the monsoon season, ridges establish over the Arabian sea off the west coast of India and a trough over the eastern peninsular India off the east coast. Over north India, a trough is established which extends from northwest India to Head Bay. This trough is known as
monsoon trough and monsoon depressions, which form over the head bay, move across the Gangetic plain along the trough line. The pressure values are in the range 1000-1002 over the monsoon trough regions and increases to the south with values of the range 1004-1008 over the peninsular India region. High pressure also can be noticed over the Tibetan region.

### 1.3.2 Temperature

#### (a) Surface Temperature

During summer monsoon season the temperature over the peninsular Indian region ranges from 28-30°C whereas it reaches to about 34°C over the Rajasthan region (Pant and Rupa Kumar, 1997). Usually the peak temperatures are expected in summer season but due to monsoon modification, the region experiences peak in the temperatures in the season prior to the monsoon

Due to the large spatial inhomogeneity, the temperature over the different homogenous region is presented below. The classification of the regions are presented in Fig 1.5. The mountainous region of western Himalayas are not included, as the temperature of the region differs highly from the inland regions of India.

---

Fig 1.5 The Homogenous regions of India
(b) Maximum temperature

The maximum temperatures observed over India region is shown for the period 1951-2003. The maximum of this quantity during the monsoon months are observed in the northwest and eastcentral region (Fig 1.6). The temperature ranges from 33-37°C and 32-35°C respectively. These regions receive lesser amount of rain compared to the other divisions of India. The temperature range is moderate for interior peninsular India with a range 30 -33°C. West central India receives more rainfall than any other region of the peninsula and the temperatures obtained are lesser compared to the other regions and it ranges from 28-31°C. This clearly implies the effect of rainfall on temperature, as the northeast India also experience low temperature, even though the lowest is over the westcentral region. The temperature ranges from 30 to 32°C.

(c) Minimum temperature

Eastcentral India and northwest India experienced higher temperature compared to the other regions of India: two regions that receive minimum rainfall during the summer season (Fig. 1.7). During the JJAS season the minimum temperature over Eastcentral and northwest India is in the range 25-26.5°C and 24-26°C respectively. The temperature difference between the maximum and minimum is around 7-8°C over these two regions. The westcentral India and interior peninsula has the same range of minimum temperatures with values 22-24°C. Northcentral experience a moderate
minimum and maximum temperature. Westcentral region has the lowest of maximum and minimum temperature. This is due to the cooling of the surface by the large amount of rainfall obtained during the season.

![Fig 1.7 Minimum temperature over different homogenous region of India during June-September for the period 1951-2003. Unit is °C.](image)

**1.3.3 Seasonal wind**

![Fig 1.8 Spatial distribution of Wind at A) 850 hPa and B) 200 hPa during June-September over India for the period 1951-2003. Unit is m/s.](image)

The wind pattern at 850 hPa (1.8 a) during the summer season (JJAS) shows a maximum along the latitudes of southern peninsular India. The Low Level Jet (LLJ) during the monsoon season is seen over the peninsular India region. The core is over the ocean with a magnitude of 14 m/s. The 200 hPa wind (Fig 1.8 b) shows strong easterlies of the Tropical Easterly Jet (TEJ) over Indian region that establish during the monsoon season.
1.3.4 Humidity

Fig 1.9 Spatial distribution of Vertically Integrated Moisture during June-September over India for the period 1951-2003. Unit is g/Kg

The maximum humidity is seen over the Tibetan region (Fig 1.9). The humidity is less over the Indian subcontinent compared to that over the Tibetan plateau. Central India region shows higher value than that over the peninsular India region.

1.3.5 Rainfall

The mean rainfall obtained during the summer season is shown in Fig 1.10. The maximum rainfall for the season is observed over the west coastal and northeast region of India. The mean rainfall ranges from 350 mm over Tamil Nadu, Karnataka and Andhra Pradesh to about 3500 mm over west coast. As the focus of the study is on peninsular India the rainfall statistics over the region is detailed. The peninsular region receives about 665.7 mm of rain during the southwest monsoon season. The subdivisions in the coastal region receive more than 1800 mm of rainfall during the season. The seasonal rainfall obtained over each subdivision is given in Table 1.1.

<table>
<thead>
<tr>
<th>Subdivision</th>
<th>Rainfall (mm)</th>
<th>Standard deviation (mm)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Kerala</td>
<td>1870.0</td>
<td>347.78</td>
</tr>
<tr>
<td>Coastal Karnataka</td>
<td>2941.8</td>
<td>111.60</td>
</tr>
<tr>
<td>South interior Karnataka</td>
<td>505.2</td>
<td>80.97</td>
</tr>
</tbody>
</table>
Table 1.1 Average seasonal rainfall obtained over the subdivisions of peninsular India for the period 1951-2003.

<table>
<thead>
<tr>
<th>Region</th>
<th>Mean Rainfall</th>
<th>Contribution</th>
</tr>
</thead>
<tbody>
<tr>
<td>North interior Karnataka</td>
<td>607.4</td>
<td>64.50</td>
</tr>
<tr>
<td>Konkan and Goa</td>
<td>2468.2</td>
<td>436.30</td>
</tr>
<tr>
<td>Tamil Nadu</td>
<td>310.9</td>
<td>123.00</td>
</tr>
<tr>
<td>Rayalaseema</td>
<td>437.2</td>
<td>85.18</td>
</tr>
<tr>
<td>Telangana</td>
<td>734.4</td>
<td>68.72</td>
</tr>
<tr>
<td>Coastal Andhra Pradesh</td>
<td>531.1</td>
<td>143.33</td>
</tr>
</tbody>
</table>

As the rainfall from June-September season is the water predominant resources over the region, rainfall expressed as percentage of annual rainfall (Fig 1.11) helps us to compare the rainfall contribution from the season to the annual totals. Over the west coastal region, the region north of 12°N contributes more than 80% of the annual rainfall. The maximum contribution is observed over the central India region with a contribution of about 80-90% of the annual rainfall. Therefore any failure in the seasonal rainfall would affect the region in a dreadful manner. Over the northeastern region the contribution is about 60-80% of annual rainfall. But over Tamil Nadu and south Interior Karnataka, it is only about 40-60% of the annual total.

Fig 1.10 Spatial distribution of mean rainfall received during June-September over India for the period 1951-2003. Unit is mm.
Fig 1.11 Spatial distribution of seasonal rainfall (expressed as percentage of annual rainfall) during southwest monsoon season over India for the period 1951-2003.

Statistics regarding the contribution of rainfall from the individual months to the seasonal totals are also important. Figs 1.12 a-d show the percentage of monthly rainfall to the seasonal totals. During June, a maximum of 24% is observed over the Kerala subdivision, which is the gateway of southwest monsoon to the Asian continent. The least is observed over Tamil Nadu with 6-9% of rainfall. In the month of July, the west coastal region receives about 30-35% of seasonal rainfall and in August, the west coast receives about 20%, whereas the maximum shifts to the central India region with 30-35% of annual rainfall. The lee side of the Western Ghats gets about 10-15% of rain in August. The west coast contributes only 10-15% of rainfall to the annual total, as it in September the monsoon starts retreating. In September, the maximum is observed over the leeside of Western Ghats, over south interior Karnataka and Tamil Nadu with 20-25% of rainfall.

A maximum of 30mm/rainy day is obtained over the coastal regions and a minimum of 9mm/rainy day is obtained on the lee side of the Western Ghats (Fig 1.13). Subdivisions of east coastal region receive an average of 12mm/rainy days. The frequency of the rainy days (days with rainfall > 1mm) during the season ranges from 20-100 days (Fig 1.14). It varies from about 80-100 days over the west coast and northeast India to about 20-30 days over Tamil Nadu (the lee side of Western Ghats) and northwestern region of India. Over central India region the rainy days range from 40-80 days.
Fig 1.12 Spatial distribution of monthly rainfall (expressed as percentage of annual rainfall) for individual months A) June B) July, C) August and D) September over India for the period 1951-2003.

Fig 1.13 Spatial distribution of the mean rain rate during the southwest monsoon season over India for the period 1951-2003. Unit is mm/day.
As there is large variability present in the rainfall amount and rainy days over the country, a comprehension on the longterm variation of rainfall is adequate.

1.4 LONGTERM VARIABILITY

The rainfall pattern over a region is determined by the natural variability at interannual and intraseasonal scales. Other than these, human activities also change the rainfall pattern over a region. One important action is the change in land-cover resulting from their need to fulfill life’s requirement (Turner et al., 1994). A change in land cover would change the hydrological cycle (Henderson-Sellers et al., 1993; Zheng and Eltahir, 1998) by changing the evapotranspiration and thereby affecting rainfall and run off to rivers. The effect of deforestation on the rainfall has also been documented (Kanae et al., 2001; Chase et al., 1996). These lead to significant local variation in rainfall occurrence in humid tropical regions (Reichardt et al., 1995).

Numerical model simulations with enhanced greenhouse gases suggest an increase in atmospheric moisture, which results in increased frequency and intensity of extreme rainfall events (Trenberth et al., 2003; Meehl et al., 2000). At a regional scale the impact of global warming is more significant due to the changes in the hydrological cycle (Hewitson, 1997). In an increased greenhouse scenario, it has been shown that the range of
precipitation intensity broadens over the Southeast Asian monsoon region \textit{(Bhaskharan and Mitchell, 1998)}. Lal (2003) had also suggested a strong possibility for the central plains of India to receive intense monsoon showers. Similar results in the frequency and the magnitude of extreme rain events over central India was also obtained for the period 1951-2000 \textit{(Goswami et al., 2006)}. A decreasing trend in the frequency of moderate events was also noticed (Fig 1.15). Another study by Meehl et al. (2000) has shown an increase in the Indian monsoon variability and extremes and several other numerical studies have shown an increased greenhouse gases would increase moisture transport to Indian subcontinent resulting in the rise of extreme precipitation events \textit{(Bhaskaran et al., 1995; May, 2002)}.

![Fig 1.15 The Trend of (A) heavy (R≥100 mm/day, bold line) and moderate (5≤R<100mm/day, thin line) daily rain events and (B) very heavy events (R≥150 mm/day) during summer monsoon over Central India. (dashed line show statistically significance) (Adapted from Goswami et al 2006)](image)

Warming of the sea surface temperature \textit{(Trenberth et al., 2003)} affects the large-scale availability of moisture \textit{(Trenberth et al., 2005)}, which in turn affects the rainfall. Fig 1.16 shows an increase in the Indian Ocean SST that has been reported in \textit{Goswami et al. (2006)}. It is also known that the extreme events have the potential to indicate long-term seasonal climatic changes \textit{(Keim and Cruise, 1998)}, therefore a study on the variability of extreme events are highly significant. Moreover, such events are the result of small-scale convective instabilities in a moist atmosphere \textit{(Goswami et al., 2006)}. The trends of extreme events were analysed at different spatial scales, in order to understand the
longterm variations. *Easterling et al. (2000)* has studied the extremes in climate events around the globe, whereas *Zveryaev and Aleksandrova (2004)* analysed the trends in seasonal rainfall over south Asia.

Studies on the longterm variability of rainfall have been an area of interest from the earlier decades. Using the rainfall dataset of 30 stations for varying periods, southwest coastal and northeast India regions were found to exhibit a positive trend *(Pramanik and Jagannathan, 1953)*. The trend was examined for subdivisional regions during 1875-1955 and found to exhibit small decreasing trend in some subdivisions but a major change could not be noticed *(Rao and Jagannathan, 1963)*. The study conducted by *Parthasarathy and Dhar (1974)* showed a positive trend over central India, parts of the peninsula, small regions of northwest and northeast India whereas a decreasing trend were noticed over some parts of eastern India. This study utilized the subdivisional data from 1901-1960. *Alvi and Koteswaram (1985)* found increasing trends in rainfall over west–coastal stations from 1901-60 and a decreasing trend after the sixties.

Long-term observational data records were used to study the trends in the rainfall patterns, extremes events *(Rupa Kumar et al., 1992, Subbaramayya and Naidu, 1992)* and extreme events of 1 to 3 days’ duration over India *(Rakhecha and Soman, 1994)*. *Srivatsava et al. (1992)* have shown that the all India monsoon and annual rainfall has no trend but a decreasing trend over the hilly regions of the northeast India. *Chhabra et al. (1997)* noticed a decrease in the rainfall over the hilly stations and an increase in the precipitation in the

![Fig 1.16 The increase in summer tropical India ocean SST for June-September season during the period 1950-2000 (Adapted from Goswami et al 2006)](image-url)

Fig 1.16 The increase in summer tropical India ocean SST for June-September season during the period 1950-2000 (Adapted from Goswami et al 2006)
urbanized/industrialized cities. Sinha Ray and Srivastava (2000) studied the trend in extreme events over 151 stations over India during the period 1901 to 1990. Trends and periodicities over 29 subdivisions of India were analysed for 124 years (Naidu et al., 1999).

Recent studies on the rainfall trends at different spatial scales are supplemented here. Using a long-term data, Singh and Sontakke (2002) studied the trends in rainfall over the Indo Gangetic Plain region. The trend of seven extreme annual rainfall indices for 129 stations over India from 1910-2000 were examined and obtained an increasing trend for 61% of the data series (Roy and Balling, 2004). Longterm trend of rainfall (1901-2003) analysed at subdvisional scale showed that the rainfall contribution of June, July and September months has decreased over many subdivisions whereas it has increased for the month of August (Guhathakurta and Rajeevan, 2006). Joshi and Rajeevan (2006) found significant positive trend in extreme events over the west coast and northwestern parts of the peninsula during the period 1901-2000. They have also presented the trend of different rainfall indices of 46 stations over the country.

Dash et al. (2007) indicated a twentieth century climate change by examining the trends in the temperature, precipitation, and depressions. Significant inter-annual and inter-decadal variation in the frequency of extreme rainfall events for the period 1901-2004, was attributed to the increase in sea surface temperature and flux over tropical Indian Ocean (Rajeevan et al., 2008). Krishnakumar et al. (2009) studied the rainfall trends over Kerala at monthly, seasonal and annual scale for the period 1871-2005 using the subdivisional rainfall and showed the trend to be increasing in winter, autumn and spring seasons. But a significant decrease was noticed for the summer season. Pal and Al-Tabbaa (2009) has also studied the seasonal rainfall trend over small spatial grids over Kerala and obtained similar results.

1.5 INTERANNUAL VARIABILITY

The rainfall variability at a longer period is well studied and documented. Therefore awareness on other scales of variability is quite significant. The year to year variability of summer seasonal rainfall is so high over India (Fig 1.17.). As indicated
earlier, the summer seasonal rainfall is about 85cm with a standard deviation of 10% of its rainfall. But the standard deviation at different homogenous regions implies that the rainfall over these regions has large variability at an interannual scale (Table 1.2). Northeast India exhibits maximum standard deviation whereas; the least of this quantity is over Peninsular India.

<table>
<thead>
<tr>
<th>Homogenous region</th>
<th>Standard deviation of JJAS rainfall (mm)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Northwest India</td>
<td>126.04</td>
</tr>
<tr>
<td>Westcentral India</td>
<td>123.97</td>
</tr>
<tr>
<td>Central northeast India</td>
<td>113.66</td>
</tr>
<tr>
<td>Northeast India</td>
<td>141.30</td>
</tr>
<tr>
<td>Peninsular India</td>
<td>101.02</td>
</tr>
</tbody>
</table>

Table 1.2. Standard deviation of summer seasonal rainfall obtained over the homogenous regions of India for the period 1951-2003.

The efforts to understand the year-to-year variability began long back from the days of Blanford when he first attempted to forecast the monsoon rainfall. His analysis led to the discovery of the sea-saw oscillation (Southern oscillation; SO), which is still found to be the most influencing feature of the Indian summer monsoon at interannual scale. Many attempts were made to understand the interannual variability of rainfall and its relation to ENSO (El Nino- Southern Oscillation) (Webster and Yang, 1992;
Krishnakumar et al., 1995; Rajeevan et al., 2004). Fig 1.18 shows the wavelet analysis of the JJAS rainfall over India during the period 1951-2003. It is clear that the wavelet power significant at 90% level corresponds to the 2-8 year period, which implies the ENSO signal in the rainfall.

El Nino, the oceanic component of SO is found to have a negative relation with the Indian monsoon (Ropelewski and Halpert, 1987; Nageswara Rao, 1999; Pant and Rupa Kumar, 1997). Fig 1.19 shows the schematic representation of regions whose precipitation is influenced by ENSO. Kane (1999) also found that the El Nino events active during the SWM season are associated with droughts in India.

Other than ENSO, Indian Ocean Dipole (IOD) also influences the rainfall variability over India. Kulkarni et al. (2007) has shown that the Dipole Mode Index (DMI)
Indian Monsoon Rainfall (IMR) relationship has weakened in the recent four decades (1961-2002). During the period 1871-1960, 74% of the strong monsoons were followed by weak DMI, whereas during 1961-2002, this was reduced to 43%. Ashok et al. (2001) also have studied the interannual variability of the Indian Summer Monsoon Rainfall (ISMR) for the period 1958-1997 and found that IOD and ENSO have affected ISMR complementarily. Whenever the ENSO-ISMIR relation is weak, the IOD-ISMIR relation is strong and viceversa (Fig 1.20).

The Indian Ocean (IO) plays a significant role in the evolution of ENSO. A strong (weak) SWM is preceded by positive (negative) SST anomalies over the southeastern subtropical Indian Ocean during boreal winter, which is related to the ENSO phenomenon (Terray et al. 2003). This boreal winter SST over the southeast Indian Ocean (SEIO) has been identified as a unique precursor to the Indian-Australian Monsoon, IOD and maritime continental rainfall. Significant spectral peaks at 2 and 4–8 years time scales were found in SEIO index (Fig 1.21), similar to that of ENSO and Indian-Australian rainfall (Terray et al 2005). The interdecadal variations in the Pacific also influenced ENSO and its relationship to Indian monsoon. The shift of 1976-77 is marked by a warmer eastern Pacific and cooler western Pacific (Nitta and Yamada, 1989, Trenberth, 1990) along with a deepening of the Aleutian Low and strengthening of the midlatitude westerlies (Nakamura et al., 1997, Zhang et al., 1997). Such Interdecadal variations occurred in the background state has influenced the El Nino onset (Wang 1995). An and Wang (2000) found that the frequency of the ENSO cycle decreased from
2-4 yr during 1962-75 to 4-6yr during 1980-93 period, associated with a change in the structure of the ENSO mode.

Gong and Ho (2002) and Huang (2001) suggested that the ENSO-like decadal warming in the tropical central-eastern Pacific to be the cause of the climate shift, while an interdecadal warming of Indian Ocean also found to play a key role in the climate shift (Fu et al., 2009). Over India, after the climate shift of 1976, a weakening ENSO relationship with SWM rainfall (Krishnakumar et al., 1999) and a strengthening relationship with NEM rainfall was found (Zubair and Ropelewski, 2006; Pankaj Kumar et al., 2007). Turner et al. (2007) and Kitoh (2007) point that the fluctuations in the amplitude of the ENSO-monsoon relations cannot be distinguished from internal variations and may not be related to climate change.

1.6 INTRASEASONAL VARIABILITY

The fluctuation between active and weak spells/breaks and northward propagations of the Tropical Convergence Zone (TCZ)/rainbelt at intervals of 2–6 weeks are the two features of the intraseasonal variation of the monsoon rainfall. Specific patterns are observed in different meteorological parameters associated with the active and break spells. Active spells are characterized by well-distributed rainfall (more than the mean over western and eastern parts) over the monsoon zone. The thresholds of 8 mm/day (12.5 mm/day) for the western (eastern) part of the monsoon zone were used to identify days of active spells (Gadgil and Joseph, 2003). During active spells, composites of OLR show negative anomalies over the monsoon zone and positive anomalies over the equatorial region (70–90°E) respectively (Fig 1.23a). A cyclonic vorticity above the boundary layer is observed during active spells (Sikka and Gadgil 1978) and active spells of the continental TCZ are associated with weak spells of the TCZ over the equatorial Indian Ocean (Sikka and Gadgil 1980). Also at 850 hPa over the Indian region, westerly anomaly occurs south of 20°N and strong easterly anomaly to the north. This implies a large cyclonic vorticity anomaly, which is also seen over the W. Pacific in the region of low OLR (Gadgil and Joseph 2003).
Ramamurthy (1969) defined break condition as one in which the surface trough is located close to the foothills. But India Meteorological Department (IMD) identify a break based on the low level pressure and wind patterns associated with such a rainfall anomaly, rather than the rainfall distribution itself (Rao, 1976). Thus breaks are characterized by the migration of monsoon trough to the foothills of Himalayas associated with an increased rainfall activity over the foothills of Himalayas and decrease of rainfall over the rest of the country. The OLR anomaly shows a north south dipole over the Indian region and another (opposite in phase) over the western Pacific (Gadgil and Joseph, 2003) (Fig 1.23 b). The westerlies are observed over the entire Indian region with no easterlies to the north of the normal position of the monsoon trough, from the surface up to 700 hPa. At 850 hPa, strong westerly anomalies are seen north of 20°N over the Indian region and the axis of the low level jet stream over the Indian longitudes (70-90°E) shifts southward to about 5°N, strengthens over the west Pacific and extends further eastward. The vorticity above the boundary layer becoming anticyclonic (Ramamurthy, 1969; Sikka and Gadgil, 1978) whereas the cyclonic vorticity in the equatorial region is enhanced by this shift of the axis of the low level jet. Over the west Pacific, the cyclonic vorticity north of 10° N is enhanced with the strengthening of westerlies (Gadgil and Joseph, 2003). The 850 hPa wind (Fig 1.24) comprises a quadrapole similar to that seen in the OLR pattern. As noted by (Raghavan, 1973), during long intense breaks, the surface temperature increases rapidly and a heat low type circulation gets established over the monsoon zone with subsidence over most of the troposphere, no disturbances are generated, and the prominent trough at 700 hPa (associated with large-scale monsoon rainfall) disappears.

Revival of the monsoon occurs with northward propagation of the equatorial TCZ or with westward propagation of synoptic-scale systems generated over the Bay of Bengal (Sikka and Gadgil, 1980) which forms as a result of the westward propagating disturbance from the western Pacific (Krishnamurti et al., 1977).
The active and break cycle of precipitation of south Asian monsoon is presented by Wang et al. (2005). The composites of rainfall for the May-September (2000-2002) season were analysed to understand the patterns of monsoon ISO (Fig 1.26).

Fig 1.23 The OLR anomaly pattern during (a) active and (b) break (adapted from Gadgil and Joseph, 2003)

Fig 1.24 The composite wind vectors at 850 hPa during (a) active and (b) break (Gadgil and Joseph, 2003)
The summer monsoon rainfall has found to vary at quasi-periods of 2–6 days, 10–20 days (Krishnamurti and Bhalme 1976, Krishnamurti and Ardunay 1980), and 30–60 days (Sikka and Gadgil 1980; Yasunari 1980; Hartmann and Michelson 1989). Variability on the timescale of 2–6 days is believed to be a reflection of local instabilities, whereas the 10–20-day oscillation is found to be associated with the life cycle of monsoon depressions (Krishnamurti and Bhalme, 1976; Goswami, 2005). The mode is associated with a westward propagation from western Pacific to about 70°E (Bin Wang, 2006). The phase speed of propagation is about 4.5 ms⁻¹ (Chen and Chen, 1993; Chatterjee and Goswami, 2004) and this mode over the south Asian (SA) monsoon region is closely linked with that over the east Asian (EA) monsoon region during northern summer and that westward propagation from EA monsoon region influence the 10-20 day variability over the SA monsoon region (Fukutomi and Yasunari, 1999; Annamalai and Slingo, 2001). Rainfall variations on the 30–60-day timescale occur under the influence of northward-moving cyclonic circulation anomalies associated with the regional tropical convergence zone (Goswami, 1994).

Fig 1.26 Composite life cycle of the South Asian summer monsoon 20–50 day oscillation. The contours of green (lavender) are positive (negative) precipitation anomalies starting from 2 mm/day (-2 mm/ day) with
a contour interval of 3 mm/day. The thick green contour outlines the major positive precipitation anomalies. The color shading represents SST anomalies in units of °C. The magnitude of anomalous rainfall rate exceeding 2 mm/day and anomalous SST greater than 0.1 °C are statistically significant at the 90% confidence level. (Adapted from Wang et al 2005)

They repeatedly propagate northward starting from south of the equator to the foothills of Himalayas during the summer monsoon season. The 10-20day and 30-60day oscillations are the two dominant periodicities of intraseasonal oscillations (ISO). The active-break cycles of precipitation are the manifestation of the ISO; the circulation associated with them covers the entire Indian region and influence the tropics and north Pacific Ocean (Webster et al., 1998). They are associated with enhanced (decreased) rainfall over central and western India and decreased (enhanced) rainfall over the southeastern peninsula and eastern India (Krishnamurthy and Shukla, 2000).

1.7 DIURNAL VARIABILITY

Monsoon convection is highly organized in space and time, on diurnal, intraseasonal and seasonal time scales (Krishnamurthy and Shukla, 2007). Using 3-hourly in situ weather reports for the period 1975 to 1997, Dai (2001) was the first to study the diurnal cycle of global precipitation whereas Tropical Rainfall Measuring Mission (TRMM) data was later utilized to study the mechanism of the diurnal variability of rainfall over the tropics (Yang and Smith, 2006). Recent study on the diurnal variations of the global tropical precipitation using high resolution TRMM datasets (3B42 and 3G68) for the period 1998–2006 (Kikuchi and Wang, 2008) characterized the global precipitation into three regimes; oceanic, continental and coastal (Fig 1.27). The rainfall peaks between 0600-0900 for oceanic and 1500-1800 Local Solar Time (LST) for continental respectively. For coastal regimes, two subregimes - seaside and landside with peaks between 2100-1200 of next day and 1200-2100 LST respectively were also identified. Thus Indian Ocean and Indian subcontinent comes under oceanic and coastal regions respectively.
Station data and remotely sensed data were analysed over the tropical Asian region and a maximum convective activity was found to occur in the early evening over land and late night/early morning over coastal regions and windward sides of mountains (Ohsawa et al., 2001). The diurnal variability of the tropical rainfall has also been studied using TRMM and GOES datasets (Sorooshian et al., 2002). They identified that the convective storms over Calcutta have a maximum from local noon to 1500 and over the northeastern India and east Himalayan plateau the maximum is during 1500-1700. The climatological diurnal cycle of convective activity over South Asia was characterized by using the cloud and precipitation data from METEOSAT-5 and TRMM respectively. Three regions, the Himalayas, north India and Bay of Bengal were selected to study the convective activity. Analysis found that during the summer monsoon season cloud clusters move over Himalayas during the morning hours than during the afternoon hours (Bhat et al., 2010). Similar datasets were used to understand the continental mode of diurnal cycle during the Asian monsoon season (Krishnamurthi and Kishtawal, 2000). They showed that the afternoon and morning circulations are opposite to each other. The analysis also found that the rainfall distribution during the morning hours are located over the Bay of Bengal and the west coast of India whereas during evening hours this high precipitation region shifts to the land area, which was dry during morning hours.
Over the northwest side of BOB, convective activity with morning maximum in cloudiness during the summer monsoon season has been reported by Zuidema (2003) and found that the convection originated is due to the land-sea contrast. The importance of the diurnal variability of convection in the heavy-rainfall centers was emphasized by Xie et al. (2006). The study over the eastern equatorial Indian Ocean for the period October 2001 to 2003 showed a precipitation maximum in the afternoon hours and the diurnal variability was dominant for all the seasons (Ramesh Kumar et al., 2006).

Many studies have attempted to understand the diurnal variability over India Subcontinent. Pathan (1984) analysed the hourly rainfall over a 36 stations well distributed over India and found an enhanced rainfall activity during the midnight to noon hours over the coastal stations and an afternoon to evening maximum over inland regions. Over the stations Cochin and Mincoy, the hourly rainfall was analysed and found that Cochin receives 250 hours of rain and Minicoy receives 110 hours of rain, of which, half of the monsoon rainfall was contributed by heavy showers (Ananthakrishnan and Rajan, 1987). Recent study by Sahany et al. (2010) gave an idea on the climatological mode of peak rainfall over India (Fig 1.28). The study showed the time at which maximum rainfall events occur over India. It is clear that maximum rainfall occurs during the late afternoon-late evening hours over the Indian landmass whereas early morning peaks are observed over BOB close to land region.

Fig 1.28 spatial distribution of the “climatological” mode of the peak octet for June-September. Peak octet is estimated for 9 years (1999-2007) and pooled together into one frequency distribution and its mode is estimated (Adapted from Sahany et al 2010).
1.7.1 Mechanisms of diurnal variability

*Yang and Smith (2006)* used TRMM observations for 1998 to provide a detailed description of the possible mechanisms that govern the diurnal cycle of tropical rainfall. *Ramage (1964)* has described the role played by sea and land breezes in the diurnal variation of rainfall. He explained that the a breeze in the direction of the synoptic wind results in the low-level convergence that leads to increased convection and rainfall rate whereas a breeze opposite to the synoptic wind reduces the rainfall rate. Therefore during the summer monsoon the westerly winds along the west coast interacts with the land breeze to result in the maximum precipitation during the late night/early morning hours. Under favourable conditions, convective cells formed over sea also move inland leading to the maximum rainfall in the early morning hours. Over sea, the SST at night is higher than the temperature over land and enhanced sensible and latent heat fluxes into the atmospheric boundary layer lead to an increase in convective activity and the precipitation rate. Basu 2007 found a maximum in rainfall to coincide with the maximum afternoon temperature and get modified by mesoscale processes like katabatic winds or land–sea breezes that produce strong convergence and associated convection. Over Himalayas, and other hill ranges, the katabatic wind plays significant role in enhancing convection and thereby to result in maximum precipitation in early morning hours. Far in the sea, the hour of maximum rain is close to noon, which coincides with the time of maximum relative humidity. The early morning maximum over the sea regions along the west coast of India results from the convection produced by the low-level convergence produced of the synoptic-scale westerly wind with the opposing land breeze whereas the early morning peak over the east coast sea regions (south of 15°N) is due to the weak synoptic-scale westerlies and the diurnal variation over the region is dominated by the land–sea breeze. The stronger vertical is the basis for the heavy precipitation over the west coastal region. *Yang and Slingo (2001)* also found the afternoon/evening peak in continental convection is due to mountain–valley and land–sea breezes.

Monsoon is a result of the differential heating between land and surrounding oceans and the interaction of mountains over the region. The moisture laden southwesterly winds during the season are forced to rise over the Western Ghats which
leads to anabatic winds during day and katabatic winds during night which gives showers in coastal areas during the morning hours. The mechanism of rainfall over the inland region is explained as the result of daytime heating of the landmass, which results in convective showers (Roy and Balling, 2007). The effect of the mesoscale topography and the drift of thunderstorms have been noticed in the diurnal variation of precipitation over central India during the monsoon. The rainfall maximum in the afternoon/evening hours are found to be the result of land heating resulting in the convective showers (Halder et al., 1991). Prasad (1974) observed a maximum (minimum) in rainfall in the early morning (afternoon) hours along the foothills of the eastern Himalayas and indicate that an increase in convection is due to the low-level convergence produced by katabatic mountain winds.

**1.8 INFLUENCE OF OROGRAPHY AND LANDUSE ON THE CLIMATE**

Rising and descending atmospheric motions forced by topography affects the precipitation. Air impinging on a mountain can be forced up either mechanically/thermally. The forced ascent is the result of mechanical lift whereas the heated mountain slopes trigger buoyancy-driven circulations to lift parcel up by thermal processes. As air rises it gets expanded and results in cooling of the parcel. When air cools, the parcel gets saturated and the water vapor condenses into cloud droplets or forms cloud ice crystals. These droplets and crystals grow by various processes until they become large enough to fall as rain and snow (Roe, 2005). Thus orographic precipitation has basically three components. They are (1) moist, large-scale flow towards an obstacle (hill, mountain or mountain chain), (2) mesoscale orographically induced lifting of the large scale flow (which cools the air to saturation and induces condensation), and (3) conversion of the condensate to precipitable particles (by some combination of smaller scale convection, turbulent air motions, and cloud microphysics). But the growth processes depends on how the wind flows past the mountain. It may flow over or around the mountain. The processes differ for blocking and unblocked conditions. Grossman and Durran (1984) found that on the west coast of India, at a mesoscale distance upstream of the Western Ghats, blocking enhances precipitation. Using a “upslope” model (Smith, 1979; Smith,
found that the key controlling parameters of orographic precipitation are the moisture flux, which determines the vapor available for condensation, and the topographic slope in the direction of the airflow, which determines the rate of the forced vertical motion.

Orographic precipitation is basically a short-lived phenomenon and can be caused by the passage of a weather disturbance, resulting in increased rain rate during the storm with a change in the synoptic situation. Some of the dynamical effects described in section 1.8.1 can also play a significant role in contributing to the large amount of precipitation over the region. Even though these processes have a role in orographic precipitation it is not comparable to the storm-averaged precipitation pattern.

1.8.1 Mechanisms of orographic precipitation

The well known mechanism of orographic precipitation is the stable upslope ascent (Fig 1.29). Air forced upward by the mechanical lifting as it impinge on the windward side of a mountain leads to the cooling of air column, resulting in condensation and precipitation. But as the air descent on the lee side of the mountain, the air gets warming and dried up and precipitation is suppressed. Such processes are commonly observed over the large mountains of midlatitude with width 40Km and height 1.5Km (Smith et al. 2003). But the ascent is not possible when the atmosphere is stable or if the approaching wind is weak. In such cases, the air that gets blocked at a lower level may move around the mountain or stagnate. This blocked air can cause ascent further windward of the range and can also enhance the lifting (and hence the precipitation) that does occur. Blocking effects have been reported in many studies (Rotunno and Ferretti 2001; Houze et al. 2001; Jiang 2003). As the melting or evaporating precipitation moves through the air column, it cools the air and results in strong down-valley wind (Fig. 1.29c). Sometimes the airflow gets diverted and precipitates on the lee side of the mountain (Fig. 1.29d). Another mechanism is the triggering of the unstable convection (Banta 1990). If air is lifted above its level of free convection it will continue to rise (Figure 1.29e). As a result the condensation rate increases and more supercooled
water droplets are formed which leads to the formation of ice crystals. Thus an enhancement in precipitation occurs, which would otherwise a slow process. Another important mechanism in orographic precipitation is the resultant of solar heating on the windward slopes of the mountain, which helps the thermally forced ascent of the air flow. As a result the air gets expanded, cools and if enough moisture is present it gets saturated. This is responsible for the afternoon thundershowers in summer that occur over many mountains (Fig. 1.29f).

Fig 1.29 Schematic illustrations of different mechanisms of orographic precipitation. (a) stable upslope ascent, (b) partial blocking of the impinging air mass, (c) downvalley flow induced by evaporative cooling, (d) lee-side convergence, (e) convection triggered by solar heating, (f) convection owing to mechanical lifting above level of free convection, and (g) seeder-feeder mechanism. (Roe 2005)

Attempts of Sarker (1966, 67) to understand the orographic precipitation over Western Ghats was successful to some extent. He neglected the earth’s rotation and friction, while developing his two-dimensional model. But he obtained comparable results with observation (Fig 1.30). The model could account for the increases from coast to inland along the slope, the maximum before the crest of the mountain and sharp decrease leeside. The maximum on the windward side was found to occur at 10-12 km before the crest of the mountain. The effect of orography is felt about 30km from coast and the rainfall was found to extend to the lee side upto 40Km from crest. His study also found that the that rainfall on the windward side originates at low levels below 5 km and that on the lee side has a high-level origin above 5 km. Therefore the rainfall on the
windward side of western Ghats is due to the condensation-coalescence process that results from clouds below freezing level where as on lee side the Bergeron process through ice nucleation brings rainfall, as the clouds extend beyond freezing level.

De and Dutta (2005) studied the role of convective instability in producing heavy rainfall along the west coastal region of India. The events were associated with increase in CAPE, a non dimensional parameter.

Over Western Ghats, the study on the convective system during 23-25June 1979 helped to understand the basic feature present during heavy orographic rainfall (Ogura and Yoshizaki, 1988). The event produced precipitation rate of about 100–200 mm/day (Fig 1.31). The importance of the interaction between latent heating and the basic flow is described in Smith and Lin (1983). But the study of Ogura and Yoshizaki (1988) showed the presence of a strong low-level westerly jet, sheared environment, and latent and sensible heat fluxes from the ocean are essential, in addition to that found by Smith and Lin (1983). As the event selected occurs in the summer monsoon season, the region experiences strong westerly current of the Low Level Jet (LLJ), which carries high moisture, a requirement for heavy orographic precipitation. Other than these, their study found out a critical level at which the role played by the low-level westerly flow becomes easterly. A critical level is defined as the level at which the horizontal wind speed \( U \)
equals the propagation speed \( \dot{\phi} \) of the disturbance (wave); that is, \( U = c \). (Note that \( U = 0 \) for a stationary system.). Therefore for a convective system (quasi-stationary) the critical level coincides with the wind reversal level. The analysis made by Lin et al. (2001) hypothesized that the critical level helps to stagnate the convective system and thereby helps to maximize the vertical motion generated by the latent heating (Lin, 1987) to give rainfall on the windward side of the mountain.

![Fig 1.31 Precipitation intensity (mm day\(^{-1}\)) estimated from satellite and rain gauge data over India and the Arabian Sea on (a) 23, (b) 24, and (c) 25 Jun 1979. The maximum values in rainfall intensity are 114, 197, and 169 mm day\(^{-1}\) on 23, 24, and 25 Jun, respectively. (d) The vertical profile of the east–west component of the wind. The shaded areas are mean precipitation areas. (After Ogura and Yoshizaki (1988))](image)

Lin et al. (2001) based on the search for some common ingredients for heavy orographic rainfall by analyzing different cases across the globe, found the following: 1) high precipitation efficiency of the incoming airstream; 2) the presence of a moist, moderate to intense LLJ; 3) steep orography to help release the instability; 4) favorable (e.g., concave) mountain geometry and a confluent flow field; 5) strong environmentally forced upward vertical motion; 6) the presence of a high moisture flow upstream; 7) a preexisting large-scale convective system; 8) slow (impeded or retarded) movement of the convective system; and 9) a conditionally or potentially unstable upstream airflow. In addition, East Asian cases were found with higher CAPE values. Doswell et al. (1996) first
proposed a “ingredients-based methodology”, on which Lin et al. (2001) added more ingredients.

1.8.2 Landuse

Human-induced landscapes and human activities play a key role in altering the climate at a local and regional scale (Cotton and Pielke, 2007; Pielke et al., 2002) and their interrelationships have been extensively studied (Chase et al., 2000; Feddema et al., 2005a). In addition, the effects of deforestation and denudation of natural landscapes and their impacts on the surrounding environment have also been widely studied (Brown et al., 1991). Ramankutty and Foley (1999) found significant conversion of forest lands to croplands since the year 1700. He pointed out that over the Indo-Gangetic plain the cropland expansion has replaced the forests/woodlands. The possible cause for the reduction in rainfall during 1983–1992 over West Bengal to half of its amount during 1973–1982 has been attributed to the doubling of the paddy crops. The wetter soils along the coast also found to reduce the land-sea temperature contrast which weakens the sea-breeze circulation, reducing localized convective precipitation in the coastal area (Lohar and Pal 1995). About 25% of the increase in greenhouse gases are due to anthropogenic activities (Houghton, 1990), but beyond the radiative effects of greenhouse gases, land use change is important in climate policy concerns (Pielke et al., 2002). Gordon et al. (2005) indicate the increased risk of expanding irrigation over India as this would influence Asian monsoon, which in turn could affect food production over sub-Saharan Africa. But Bansil (2004) report that the irrigated agriculture in India to be the key component of economic development and poverty alleviation. But the dreadful effects of increased irrigation are dealt in Boucher et al. (2004). According to them, addition of water vapour over the dry regions will result in increased precipitable water, which would decrease the moisture availability over other regions through changes in convection.

Land cover changes and modification of the surface energy balance, and the resultant cooling over Asia has been recognized to be one reason behind the modification in summer monsoon over India (Feddema et al., 2005a). Several other studies
also indicate landuse modification to play a crucial role on the physical processes that determine the temperature trends over the region (Niyogi et al., 2002; Mahmood et al., 2004; Feddema et al., 2005b). Even agricultural conversions reported over southern Florida is associated with reduced precipitation, redistributed latent heat flux and atmospheric water vapor (Marshall et al., 2004a). The adverse effects of irrigation on regional climate is also reported over United States (Adegoke et al., 2003; Pielke, 2001).

The effects of tropical deforestation has been studied extensively over southeast Asia using regional climate models (Baidya Roy and Avissar, 2002; Sen et al., 2004). Deforestation results in higher surface albedo (Myhre and Myhre, 2003; Hales et al., 2004) and lower surface roughness (Sen et al., 2004). de Rosnay et al. (2003) reported a 9.5% increase in latent heat fluxes due to irrigated agriculture in India. Sarkar and Kafatos (2004) showed that Indian monsoon rainfall and land surface temperature over the Indian subcontinent have a significant impact on the spatial and temporal distribution of vegetation. Study by Roy et al., (2007) also showed that over India, agriculture and irrigation can substantially reduce the air temperature over different regions during the growing season. Land cover change over the Indian subcontinent during pre-monsoon season (March, April, and May) affects the early Indian summer monsoon rainfall and July precipitation for the period of 1982–2003. Decreased July surface temperature in the Indian subcontinent (an expected result of increased evapotranspiration due to irrigation and increased vegetation) leads to a reduced land–sea thermal contrast, which is one of the factors driving the monsoon, and therefore weakens the monsoon circulation. A weak early ISM appears to be at least partially a result of irrigation and the resultant increased vegetation and crop activity prior to the monsoon (Lee et al., 2009). Few studies have assessed the effect of urbanisation rates around cities using demographic (Mukherji, 2001), and spatial data (Bhagat and Mohanthy, 2008). As Davis (2001) state, humans prosper and agriculture flourish during wet phases of the summer monsoon while famine and population migration be in the lead during periods of weak summer monsoons.
1.9 EMPIRICAL FORECASTING OF MONSOON RAINFALL

The first attempt to forecast the monsoon was done by Blanford (1884) based on his hypothesis that the ‘varying extent and thickness of the Himalayan snow exercise a great and prolonged influence on the climate conditions and weather of the plains of northwest India’. And the operational long-range forecast (LRF) of monsoon rainfall covering the whole of India and Burma was started in 1886, followed by his cautious forecasts during 1882-85. Sir John Elliot who succeeded Blanford in 1895 utilised the weather conditions over the whole of India and surrounding regions. His used the (i) Himalayan snow cover, October to May, (ii) 'local peculiarities' of pre-monsoon weather in India and (iii) 'local peculiarities' over Indian Ocean and Australia (Thapliyal, 1987). Later, Sir Gilbert Walker (1923, 1924) conducted extensive study on the variations of pressure, temperature, rainfall etc. and identified two pressure patterns; North Atlantic Oscillation, NAO and North Pacific Oscillation, NPO in Northern Hemisphere with a regional nature and a global scale phenomenon; Southern Oscillation, SO in Southern hemisphere. The SO was later linked to the oceanic phenomenon of El Nino in the east-equatorial Pacific characterized by warming of the sea surface along the Peru coast. North Atlantic Oscillation (NAO), North Pacific Oscillation (NPO) and Southern Oscillation (SO). He was the first to introduce an objective method in forecasting by using the correlation analysis for finding the predictors in LRF.

Statistical or empirical techniques were commonly used in the studies of LRF. The suitable predictors are identified through correlation analysis, which brings out the relationships between rainfall and regional/global fields of several surface/upper-air parameters. The LRF predictors are broadly classified into 4 groups representing the (i) Regional conditions, (ii) El nino/Southern Oscillation (ENSO) indicators, (iii) Cross-equatorial flow and (iv) Global/hemispheric conditions.

The techniques commonly used for the prediction of AISMR include simple and multiple linear regression analysis. A large number of regression models (simple as well as multiple) have been proposed so far (Hastenrath, 1991). Sadhuram and Murthy (2008) used a simple multiple regression (MR) model for LRF of Indian Summer Monsoon
Rainfall (ISMR). They claim that their model is as good as the IMD’s 8 and 10 parameter operational models.

Auto-regressive integrated moving average (ARIMA) models used to forecast AISMR and monsoon rainfall over northwest and peninsular India have slightly higher accuracy compared to MR models (Thapliyal, 1990), but the autocorrelation for the AISMR were not significant for lags 1-5, during the period 1871-1990 (Parthasarathy et al., 1994). Dynamic Stochastic model (DST) model was used to predict AISMR and rainfall over Northwestern and Peninsular India (Thapliyal, 1987). The model uses the 500 hPa ridge position as the predictor coupled with a stochastic transfer system represented by ARIMA process with rainfall as output (Thapliyal, 1987). This single predictor input model has more accuracy compared to ARIMA and multiple regression models (Thapliyal, 1990), but it needs to be improved to accommodate multiple predictors.

1.9.5 Quantitative Forecasts

1.9.5.1 Operational Models of IMD

Parametric and multiple Power regression developed by (Gowariker et al., 1989) used 15 predictors for LRF of AISMR which was updated later to include one more parameter (Gowariker et al., 1991). IMD’s operational forecasts during the period 1988-2002 were based on these models. The parametric model is qualitative and indicated the likelihood that the monsoon rainfall will be excess or deficient depending upon the proportion of favourable and unfavourable parameters. The power regression model has the following mathematical form

\[ R = C_0 + \sum_{i=1}^{n} C_i X_i^P \]  

where \( R \) is the rainfall, \( X_i \) are the predictors, and \( C_i \) and \( P_i \) are constants. where \( R \) is the dependent variable, \( X \) the independent parameters, \( C_i \) and \( P_i \) denote the model constants. The values for \( C_i \) and \( P_i \) are determined by iteration along with the least square fitting method. The iterations are carried out at very small increments. The best model is developed by including suitable parameters in the decreasing order of their correlation.
coefficient. The parameters are also checked for the consistency in the relationship by using sliding correlation. The parameters are standardized before including into the model; otherwise certain scaling constants need to be used. The value of the power is a measure of the nonlinearity, greater the value, greater the nonlinearity. This model claims to account for the non-linear interactions of different climatic forcing with the Indian monsoon system. However, the mean forecast error during this period was more than the mean error of the forecasts based on climatology alone and the model failed to predict the severe drought of 2002.

A critical evaluation of the 16-parameter power regression and parametric models was made and the analysis revealed the weak correlations of six April–May parameters and four winter–spring parameters. These ten predictors were removed from the model and extensive data analysis was carried out to find more stable and physically related predictors for use in the LRF model. Thus four new predictors were found (Table 1.3) out and two new models were framed (8 and 10 parameter models) for the operational work.

Thus the forecasts were disseminated at two stages, first forecast is issued in mid April and an update or second stage forecast is issued by the end of June (Rajeevan et al., 2004). The new 8-parameter and 10-parameter models were developed using data of 38 years (1958–95). Another seven years’ data (1996–2002) were used for independent verification of the models. The 2003 operational forecasts for the southwest monsoon rainfall based on these new models were proved correct (Fig 1.32). However, in 2004, the forecast for a normal monsoon did not materialize, due to unexpected poor performance of monsoon.

Since 1999, IMD has been issuing long-range forecasts for three broad homogeneous regions of India, viz. northwest India, northeast India and the peninsula. For northeast and Northwest India, two new predictors were included, whereas only one new predictor has been added into the model for Peninsular India. A detailed study regarding the various empirical techniques for rainfall forecasting can be found in Nayagam et al., 2012.
1.10 REGIONAL MODELLING STUDIES

General Circulation Models (GCMs) are useful to simulate the large-scale features across the globe, but it fails to perform well over complex orographic regions due to the poor simulation of regional processes (Gates, 1992, Gao et al., 2002). Leung et al. (2003), Lee and Sub (2000) showed that the cumulus parameterization schemes of regional climate models do not work well over Asian Monsoon region. The presence of Tibetan plateau, complex orography (Himalayas and western Ghats), ocean-continent contrast and sea-air interactions makes the scale interactions are extremely complex in the Asian monsoon regions (Holland 1995). Giorgi and Mearns, 1991, 1999 was partially successful in using limited area model driven by GCM outputs. The idea behind regional climate models with a suitably high-resolution is that, given a large-scale atmospheric circulation, then the model resolves the complex topography, land-sea contrast, land use and physical processes and gives a reasonable simulation in both space and time.
Numerous attempts were made to simulate the Asian monsoon using Regional climate models. The extreme weather events and the features of Indian monsoon were simulated by Bhaskaran et al. (1996) using a regional climate model nested with global atmospheric GCM. The results showed that the model derived precipitation is larger by 20\% than GCM. Some were successful in simulating the monsoon precipitation in selected years (Galle’e et al. 2004; Messager et al. 2004). Ji and Vernekar (1997) nested Eta model with COLA GCM to simulate the contrasting summer monsoons of 1987 and 1988. Prasad (1997) used IMD’s model to study the role of moisture in cyclone tracking. MM5 was made use to simulate features such as sea level pressure, 925 hPa temperature, low level wind and precipitation of the Indian summer monsoon Rao et al. (2004).

Recent studies by Dash et al. (2006) and Shekhar and Dash (2005) showed success in simulating the salient features of Indian summer monsoon using RegCM3. The simulations of monsoon circulations over East Asia has been made by Singh et al. (2006)
and showed promising performance of this regional model in simulating important characteristics of monsoon. He used Grell convective precipitation scheme with Arakawa-Schubert (AS) closure. Dimri and Ganju (2007) utilized this (RegCM3) model to simulate the intraseasonal variability of surface climate over the Himalayas during October 1999–March 2000 with 60 km resolution. The improvement in capturing the migration and Position of African Easterly Jet using a Community Land Model (CLM3) with Regional Climate Model (RegCM3) is showed in Steiner et al., (2009). Thus Regional Climate Model (RegCM) has been used widely for various mesoscale studies (Pal et al., 2000; Giorgi et al., 2003). But many factors are found to influence the monsoon such as sea surface temperatures (SSTs) in the Atlantic, Indian, and Pacific Ocean basins (Messager et al. 2004), current and antecedent soil moisture conditions (Philippon and Fontaine 2002; Fontaine et al. 2007; Douville et al. 2007), inflow air mass characteristics (Chou and Neelin 2003), and vegetation feedbacks (Charney et al. 1977; Zheng and Eltahir 1998).

The modifications of the biosphere in the Indian subcontinent influences monsoon circulation and rainfall by changing the surface energy balance, the planetary boundary layer motion fields, moisture convergence and the hydrological cycle. These numerical studies also suggested an increase in surface albedo and a reduction in surface roughness that weakened the Indian monsoon (Sud and Smith, 1985). Recent modeling study by de Rarnay et al. (2003) shown that irrigation alters water and energy fluxes over Indian subcontinent. His simulations with and without irrigation for the years 1987 and 1988 indicate irrigation as a major component of the regional hydrological cycle, affecting in particular the partitioning of net radiation into sensible and latent heat fluxes. Another study on the differences in moisture and energy fluxes between an agricultural and a contemporary agricultural land cover showed an increase in soil moisture and evapotranspiration due to irrigation, which in turn hinders the development of sea breeze circulations and thereby results in diminished pre-monsoon rainfall Douglas et al. (2006). A numerical study by Douglas et al., 2007 showed that due to increased irrigation, northwestern India experienced a decrease in sensible heat flux (100 Wm⁻²) and air temperature (1 to 2 K). Using the Regional Atmospheric Modeling System (RAMS), the effects of landuse modification on the Indian rainfall has been studied. The agricultural
Chapter 1

Introduction

Intensification and irrigation modify the surface moisture and energy distribution, which alters the boundary layer and regional convergence, mesoscale convection, and precipitation patterns over the Indian monsoon region. The effect of irrigation has considerable effect than the conversion of vegetation to agricultural land. Two-thirds of this increase in vapour fluxes has been attributed to irrigation, of which groundwater-based irrigation contributes about 14% and 35% of the vapor fluxes in the wet and dry seasons, respectively (Douglas et al 2009). Many studies have been made on the effect of landuse changes/deforestation using RegCM3 over Asia and Africa (Abiodun et al. 2010, Nogherotto et al 2010, Chow et al, 2009) but attempts over India are not well documented.

1.11 RELEVANCE AND AIMS OF THE STUDY

Many attempts have been made to understand the temporal or spatial variability of rainfall over India. As shown in the different sections of this Chapter the variability of rainfall at different temporal scales have noticeable impact on the total rainfall that varies from region to region. Therefore, the longterm rainfall variability over India as a whole cannot account for the regional differences in rainfall. So studies at different spatial scales - subdivisional, homogenous regions etc have been carried out. Since studies on smaller spatial grids over homogenous regions have not been made, such studies would be beneficial to understand the longterm variability of rainfall over homogenous regions.

The rainfall variability at smaller temporal scales is also equally important because its consequence (say, floods and drought conditions) can be felt on shorter timescales. The year-to-year variability in rainfall in terms of floods and droughts has a significant influence on society. As India’s economy and human lives are dependent on agriculture, a failure of monsoon would have a negative impact. The studies on interannual variability found the dominant periodicity of rainfall at 2-8 year scale, which is related to the ENSO variations. However, the climate shift of 1976 has changed the periodicity of ENSO from higher to a lower frequency and therefore, the influence of this aspect on Indian rainfall need to be studied.
Chapter 1

Introduction

The Indian monsoon at the intraseasonal scale is dominated by the active and break phases of the monsoon. Many theories have been put forward to explain the mechanism behind the active-break cycle, though very few have tried to find any regional differences in the occurrence of these cycles. India being a land of large inhomogeneity, analysis on such aspects would be valuable.

Studies on a diurnal scale are important but are restricted by the unavailability of data. Recent studies on diurnal rainfall variability have been made with station data available for a longer period of time, though its spatial resolution and uniformity is a question of debate. This has been resolved by utilizing satellite observations even though they are available for a shorter duration of time. The freely available data has a maximum resolution of 0.25x 0.25 in latitude and longitude grids. However these resolutions are quite insufficient to explain the mechanisms that cause the diurnal variability in rainfall. So some efforts have been made to numerically simulate such rainfall patterns to explain the mechanism. Such experiments will be useful and would substitute the data requirement at higher spatio-temporal scales.

The regional differences in maximum rainfall amounts is largely due to the topography and landuse of the region. The presence of complex orography over India and surrounding regions is responsible for the channeling of monsoon current and the accumulation of precipitation over the west coastal and northeastern regions. The forests over the Western Ghats and northeastern states of India also help the condensation over these regions. Numerical simulations are the best solution to study the influence of landuse and topography.

Many meteorological and oceanic parameters influence the rainfall over Peninsular India. But the variability of monsoon rainfall at different subdivisions comprising the region shows very high variability. Of the nine subdivisions of Peninsular India, Kerala and Konkan and Goa are found to have maximum variability, of which a maximum standard deviation is observed over Konkan and Goa. In addition to the large interannual rainfall variability over Kerala, the subdivision receives both the southwest and northeast monsoon rainfall.
The present study attempts to bring out the variability of summer monsoon rainfall at different temporal and spatial scales. The study has simulated the monsoon for a ten-year period from 1997-2006, which helps understand the diurnal variability and the role of orography and landuse on the climate of peninsular India. The aims of the study are to:

- understand the longterm variability of rainfall
- study the influence of climate shift of 196 on the ENSO variability in rainfall
- analyse the regional differences in the active-break conditions
- find the effect of orography and landuse on the diurnal variability of rainfall
- delineate the control of Western Ghats orography on the rainfall
- study the modulation of monsoon by the declining mountain forest
- present the empirical relationship of atmospheric and oceanic parameters with Kerala rainfall