CHAPTER 1

INTRODUCTION

1.1 Monsoon

The seasonally varying circulations associated with the annual contrast in heating of the Asian continent are the most important large-scale aspects of the general circulation of the atmosphere, which is called \textit{Asian Monsoon}. Figure 1.1 shows the region of global monsoon. The Indian summer monsoon affects the lives and the economies of many countries in Asia. The study of this phenomenon has been mainly limited to the Indian subcontinent whereas it takes place over the entire Indian Ocean. The main aspects of the monsoon are defined as follows. There should be a 120 degree shift in the direction of prevailing wind between January and July. The average frequency of prevailing wind must be greater than 40\% and the mean wind in either January (for winter monsoon) or July (for summer monsoon) must be greater than 3 m/s. It is believed that the differential heating of the land and ocean which leads to the formation of a pressure gradient gives the driving force for monsoon circulation. The Coriolis force deflecting the cross equatorial winds created by the pressure gradient causes the typical monsoon flow. The moist processes such as the release and absorption of latent heat determine the energetics of monsoon.

There are two components of monsoon moisture input; moisture due to actual evaporation from the Arabian Sea and that due to a water vapor flux from the Southern Hemisphere (cross equatorial transport). The greatest amount of moisture flux from the Southern Hemisphere is along the east African coast and is transported due to \textit{low-level Somali jet}, which is a typical feature of the
monsoon. This jet stream originates from a high-pressure system Mascarene anticyclone, formed over southeast Indian Ocean. From this high, large outflow of air takes place and it crosses the equator as cross-equatorial Somali jet and due to Coriolis force the wind becomes southwesterly. 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 stream is important in determining the moisture and rainfall over India. The Tibetan high-pressure system, which is an upper level anticyclone found above the surface monsoon trough located over north India is established over Tibetan High lands during July and is well developed at 200hPa level. It remains up to the end of summer season and then moves south-southeast direction with the movement of maximum heating to south. The outflow of air from the southern flank of Tibetan anticyclone gives rise to tropical easterly jet and it remains from June to September.

![Fig. 1.1 The geographical extend of monsoon. Average frequency of predominant surface wind directions shown in three categories. Unshaded areas are non monsoonal. (Adapted from Ramage 1971)](image)

For India, Monsoons (both southwest and northeast) are responsible for the majority of India’s annual rainfall. About 80% of the annual total rainfall is
received between the months of June and September. Orographic effects are important in determining the amount of rainfall at a particular station (Nair and Mohanakumar, 2009). Stations on the windward side of mountains receive more rainfall than those in the rain shadow region on the leeside. This factor plays a vital role over the west coast of Peninsular India. The rainfall over Central India is largely affected by monsoon depressions moving in from the Bay of Bengal and also by convergence due to the monsoon trough over central India. It is formed in the summer months as extension of Inter Tropical Convergent Zone (ITCZ) and is the region of low pressure and wind shear at the surface. South of the trough has southeasterly winds and north has northerly winds. Monsoon trough during south west monsoon exists parallel to about ~500 km south of the Himalayas. An inversion layer at 800hPa which hinders the instability would produce precipitation. The mean precipitation and winds over in Indian region during winter and summer are given in Fig 1.2

As the onset vortex moves to the north, westerlies increase sharply on the equatorward (convergent) side of the vortex. Stations on the west coast are influenced by these westerlies as well as the position of the onset vortex. But for the east coast of India, as monsoon depressions moves across, it moves into regions that may not have previously been affected by heavy rains. Stations in central India experience heavier rains than before because they are not as affected by the southwest flow as stations on the west coast. Another characteristic feature of the south west monsoon is the active and break phases. Breaks occur normally in the middle of August. During a break, monsoon trough moves north and the maximum rainfall shifts towards the foothills of the Himalayas. Low-level westerlies also move north to the Gangetic Plain and decrease in strength over the Peninsular India.
The monsoon usually starts withdrawing during the end of September from North India. Pressure gradient undergoes transition to a winter pattern (reverse conditions of what occurred in the summer pattern) and the easterlies in the upper troposphere weaken. Monsoon trough moves to the south and the westerly jet appears. The Inter tropical convergent zone moves back to the south.

![Climatological mean precipitation (unit: mm day\(^{-1}\)) during (a) boreal winter (DJF) and (d) summer (JJA). (b) and (c) are the same as (a) and (d) but for winds (unit: ms\(^{-1}\)) at 850hPa based on NCEP reanalysis. The contour interval for isotachs is 2 ms\(^{-1}\) with the minimum contour being 2. Adapted from Goswami (2005).](image)

The monsoon cloud cover is the manifestation of moist convective processes over the Indian subcontinent and it varies both in space and time. During active phase of monsoon, cloud cover is maximum in a belt running from western shore of Bay of Bengal to northern shore of Arabian Sea and minimum over foothills of Himalayas, south India and Sri Lanka. The pattern reverses during break phase and rainfall distribution closely follows the cloud distribution.

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Chapter 1

Introduction
Besides possessing the largest annual amplitude of any subtropical and tropical climate feature, the monsoons also possess considerable variability on a wide range of timescales. Within the annual cycle there are large-scale and high-amplitude variations of the monsoon. On timescales longer than the annual cycle the monsoon varies with biennial, interannual, and interdecadal rhythms. In the following sections an attempt is made to describe these variations.

1.1.1 Intraseasonal variability

Intraseasonal variability of the Indian summer monsoon, delineated by active periods of heavy rainfall interrupted by break periods, significantly modulates the seasonal mean monsoon fields (Krishnamurthy and Shukla 2000; Goswami and Ajayamohan 2001). These active and break spells of the monsoon are associated with fluctuations of the tropical convergence zone (Yasunari 1979, 1980, 1981; Sikka and Gadgil 1980). Tropical convergent zone over the Indian monsoon region represents the ascending branch of the regional Hadley circulation. The intraseasonal oscillations (ISOs) of the Indian summer monsoon represent a broadband spectrum with periods between 10 and 90 days but have two preferred bands of periods (Krishnamurti and Bhalme 1976; Krishnamurti and Arakane 1980; Yasunari 1980), one between 10 and 20 days and the other between 30 and 60 days.

Several recent modeling studies show that a significant fraction of the interannual variability of the Indian summer monsoon is governed by internal chaotic dynamics (Haugrallab and Sadourny 1995; Goswami 1998). Because the separation between the dominant ISO periods and the season is not large, the statistics of the ISOs could, in principle, influence the seasonal mean monsoon and its interannual variability.
Starting from around the mid to late 1970s, a number of studies began to identify intraseasonal fluctuations with periods around 30-60 days, associated with the Asian summer monsoon. These were well observed in both cloudiness (Murakami 1976; Yasunari 1979, 1980) and wind variability (Dakshinamurthi and Kesbavamurty 1976; Murakami 1977). The canonical space-time structures of a typical boreal summer ISO event have been well represented by Chen and Murakami (1988), Goswami et al (1998), Annamalai et al (1999), Annamalai and Slingo (2001), Lawrence and Webster (2002) and Hsu et al (2004).

In addition to the nominal time scale of 30-60 days, there also exists considerable intraseasonal variability at higher frequencies, at time scales around 10-20 days. Identification of these higher frequency fluctuations came from early studies on cloudiness and conventional synoptic observations (Krishnamurti and Bhalme 1976; Murakami 1976; Yasunari 1979; Krishnamurti and Ardanuy 1980). 10-20 day mode accounted for about a quarter of the subseasonal monsoon variability while 30-60 day accounted for about two third (Annamalai and Slingo 2001). In contrast to the low frequency ISO, variability of this mode is focused almost entirely over East Asia and the northwest Pacific region. Also 30-60 days mode originate in equatorial Indian Ocean and propagate northward/northeastward (Kripalani et al 1999; Singh and Kripalani 1985, 1986, 1990; Kripalani et al 1991) but 10-20 days mode originates in the equatorial western Pacific and propagates westward/northwestward in the form of Rossby waves at about 5 ms⁻¹. Another noteworthy feature of 10-20 days mode is that it possesses a relatively strong seasonal variation, in particular during the first half of the summer monsoon period (May-July). The 30-60 day variability prevails near the equator in particular over India, as compared to ISV of 10-20 days.
1.1.2 Interannual variability

Seasonal mean prediction and predictability experiments with General Circulation Models (GCM's) indicate that a significant fraction of observed interannual variability (approximately 50% or more) of the Indian summer monsoon may be due to internal low frequency variability (Goswami 1998). The internal low frequency variability acts as a background of unpredictable noise mixed with the predictable externally forced signal. Improvement in seasonal mean prediction would require successful extraction the signal from the background of noise of comparable magnitude. Goswami and Xavier (2005) carried out a numerical study using only annual cycle of solar external forcing and annual cycle of SST as boundary forcing. It is shown by their study that model simulates interannual variability of the Indian monsoon comparable in amplitude to the observed interannual variability. Any interannual variability of the simulated Indian summer monsoon (or any other climate system for that matter) is, therefore, of internal origin.

Observational studies have shown that there may be two time scales in the interannual variability of the monsoon-ocean-atmospheric system i.e. a quasi-biennial system with the tropical biennial oscillation (TBO) and a 4–6 year cycle associated with the El Nino Southern Oscillation (ENSO). These two scales are observed in a range of parameters like rainfall, surface pressure, wind and SST (Meehl 1987; Lau and Shen 1988; Rasmusson et al 1990; Barnett 1991; Meehl 1993). Meehl (1994) has proposed that the variability of the Asian monsoon is closely tied to the warm and cold phases of the TBO. Shen and Lau (1995) have suggested that the Asian monsoon as a whole may be more tightly coupled to the tropical ocean atmosphere on the time scales of TBO rather than ENSO. Long term climate simulations by most of the general circulation models suggest that approximately 30-40% of the variability in the Indian
summer monsoon precipitation is attributable to SST/Sea Ice variations, (Folland and Rowell 1995). In other words, a large part (60-70%) of the interannual variability of the Indian summer monsoon may be due to internal dynamics. While seasonal predictions of the mean rainfall in other parts of the tropics is not sensitive to small changes in the initial condition, simulation of seasonal mean Indian monsoon rainfall has been found to be sensitive to small changes (Palmer and Anderson 1994).

1.1.3 Long-range forecast of Indian Monsoon

Due to the intrinsic complexity, the predictability of day-to-day weather patterns in the tropics is restricted to 2-3 days. But the seasonal mean monsoon circulation in the tropics is potentially more predictable. This is due to the fact that the low-frequency component of the tropical variability is primarily forced by slowly varying boundary forcing such as sea surface temperature (SST), land surface temperature, soil moisture, snow cover which evolves on a slower time scale than that of the weather systems themselves. Observational studies have established that the Indian Summer Monsoon Rainfall is linked with several surface boundary conditions like east Pacific SST, Indian Ocean SST, Eurasian and Himalayan snow cover and land surface temperature. However data analysis and numerical modeling studies have suggested that an unyielding component indicating that the mean monsoon is also affected by internal dynamics.

1.2 Total column moisture

Moisture is critically important to life on Earth. It also plays an important role in the heat balance of Earth especially through the greenhouse effect of water vapour and, at the same time, by moderating surface temperature changes due to the removal of latent heat of evaporation. Reasons for the scanty
knowledge of moisture in the atmosphere and precipitation stem from the lack of observations (especially over the oceans) and their spatial as well as temporal variability. Rainfall and clouds usually occur on quite small temporal and spatial scales, so that an observation of moisture or precipitation event may not be representative of more than an area with dimensions of a few kilometers across or more than a small fraction of a day. Total precipitable water (TPW), which is defined as the total water vapor obtained by integrating the mass weighted specific humidity vertically from the surface to the top of an atmospheric column, varies essentially with hemispheric insolation changes. The spatiotemporal pattern of TPW over the Indian Ocean and nearby regions is closely associated with the variability of the Asian monsoon (Singh et al 2000). An increase in the Southern Hemispheric insolation results in an increase of TPW over the southern Indian Ocean. When the Southern Hemispheric insolation is enhanced from December to April, the largest increase in TPW occurs in the southern Indian Ocean in March-May. The patterns of TPW change are consistent with those of the SST change. Therefore, the increased SST is responsible, in large part, for the increase of TPW in the southern Indian Ocean (Liu et al 2006).

Theoretical studies on the moisture budget of the tropics suggest that the water vapour temperature relation in the tropics is height dependent. The observed interannual variation of water vapour and temperature give an opportunity to assess the water vapour temperature relation in the tropical troposphere. The interannual variability in the tropics is affected by the El Nino phenomenon.

1.3 Solar activity and Earth's weather

Solar variability, which originates from the solar interior, has time scales from hours to billions of years. One aspect of solar variability is linked to solar
evolution driven by nuclear conditions in the core. This is a relatively slow process with changes on time-scales of the order of several million years and above for parameters such as mass, radius and luminosity. Many forms of solar activity, such as flares, solar radiation bursts and solar wind can cause radiation enhancement and plasma movement. All these affect global climatological changes directly or indirectly (Herman and Goldberg 1978).

Another aspect of solar variability is related to solar magnetic fields generated below the convective zone in the interior. The evolution of magnetic fields results in many manifestations, on time-scales in the range of hours to several hundred even thousand years, such as the well-known sunspot cycle and its longer-period modulations, solar wind structures, and coronal mass ejections. Spectral studies of solar activity unveil certain periods and frequencies. However, wavelet analyses have showed that the periods and frequencies change with time. A much more complex picture therefore appears, than variability on only the 11-year sunspot cycle. A third aspect of solar variability deals with periodicities due to orbital conditions such as solar rotation, changes in Earth's orbit, or inclination of rotation axis with respect to the ecliptic plane.

Many of these processes are also the mechanisms by which the Sun can influence Earth and the climate. The influences from the variation of the electromagnetic radiation are most often considered. However, a changing solar wind and energetic particles can also influence Earth either directly or indirectly through modulating the cosmic ray flux. In this scenario many research works have been done on Sun Climate relations and they have detected solar signals in climate records.

Many previous studies have shown that there is a regional to global factor caused by ENSO signal with periodicity about 3-6 years and a signal with 11
years periodicity indicating the solar influence on cloud cover over Indonesia, which is typically an equatorial region. This solar signal does not appear on all regions with different climate patterns or seasonal time ranges. The stronger solar signal appears over regions with monsoon activity and during dry seasons, when local or regional effects are the least. Though there are empirical evidences about solar effects on climate in many regions of the world, actually the physical mechanism behind this is not yet well understood. Cosmic rays mechanism plays a vital role. The cosmic rays interact in the upper atmosphere and produce secondary particles neutrons and muons. When these particles interact with the air molecules or water molecules, they get polarised and act as condensation nuclei for the formation of clouds. During the sunspot minimum, the intensity of the cosmic ray becomes maximum which in turn increases the coverage of clouds. This implies that solar radiation reaching Earth will be minimized. Conversely, during sunspot maximum, the intensity of cosmic ray reaching lower levels of the atmosphere reduces and the cloud cover decreases. Many studies have shown that monsoon could be sensitive to relatively small changes in forcing (0.25% change in solar output) (Overpeck et al 1996; Neff et al 2001; Fleitmann et al 2003)

While greenhouse gas forcing is more spatially uniform, solar forcing is more spatially heterogeneous (i.e., acting more strongly in areas where sunlight reaches the surface). Consequently, solar forcing induces feedbacks involving temperature gradient-driven circulation regimes that can alter clouds. Over relatively cloud-free oceanic regions of low-level moisture divergence in the subtropics, the enhanced solar forcing produces greater evaporation. Moisture then converges into the precipitation zones, intensifying the regional monsoon and Hadley/Walker circulations. This produces greater subsidence and less cloud over the subtropical ocean regions and this process get reinforced.
The solar signal in tropospheric zonally averaged winds has also been studied by many researchers (Gleisner and Thejll 2003, Haigh et al 2004, 2005, Crooks and Gray 2005). They show that the subtropical jets are weaker and further poleward at solar maximum than at solar minimum. The magnitudes of these signals are relatively small compared with the variations associated with the North Atlantic Oscillation and volcanic signals. Gleisner and Thejll (2003), also examined the solar signal in zonally averaged vertical velocities and the longitudinal distribution of the solar signal in vertical velocities at the equator, and found a stronger subsidence at mid-latitudes and a poleward movement of the peak subsidence. They found evidence for a latitudinal broadening of the equatorial upwelling region, in good accord with the results of Haigh (2003) and Crooks and Gray (2005) though they did not find a significant solar signal in the strength of the ascending branch of the Hadley circulation. The solar signal in tropospheric zonal averaged temperatures and circulation was studied using the NCEP reanalysis by Labitzke (2002), Haigh (2003), Gleisner and Thejll (2003) and in the ERA-40 reanalysis by Crooks and Gray (2003). The results are fairly complimenting each other. A positive solar response is seen in the tropical lower stratosphere i.e. warmer tropospheric temperatures in solar maximum than in solar minimum. This signal extends in vertical bands throughout the troposphere in both hemispheres at latitudes 20°-60° with maximum amplitude of 0.5 K around 40°-50°N where there are strong gradients. Hence substantial local changes can be produced by modest North-South shifts.

1.4 Role of Indian Ocean on monsoon activity over India

Air sea interaction over the Arabian Sea and the Bay of Bengal are found to influence precipitation variability associated with summer monsoon over India. The Arabian Sea is a cardinal source of moisture fluxes across the west
coast of India resulting in more orographically forced rainfall over Western Ghats (Saha and Banakar 1977; Rakhecha and Pisharoti 1996). Vecchi and Harrison (2004) examined the role of SST anomalies in the northern Indian Ocean onto the interannual variability of precipitation in the Western Ghats region where summer monsoon rainfall shows local maximum in the seasonal mean and its interannual variability. For the Western Ghats region, they found a strong positive correlation with Sea Surface Temperature (SST) anomalies in the Arabian Sea at the onset of summer monsoon. The mean SST for winter and summer for Indian Ocean is given in Fig. 1.3. The active-break cycle of the Indian summer monsoon is associated with the northward propagation of clouds and convection bands from the equatorial region to around 30°N over the South Asian monsoon region, at a phase speed of about 1° latitude per day (Yasunari 1979; Srikka and Gadgil 1980; Yasunari 1980; Gadgil and Asha 1992; Goswami 2005). Many studies have attempted to develop a theoretical basis of atmospheric dynamics in the northward propagation of the cloud bands on this time scale (Murakami et al 1984; Jiang et al 2004; Yokoi and Satomura 2006).

Kemball-Cook and Wong (2001) examined the influence of latent heat flux on the northward migration of outgoing long wave radiation (OLR) anomalies over Indian Ocean. They suggest that the moisture flux from the Indian Ocean do favour convection by building up moist static energy. However, since their study is based on convective maximum over the equatorial Indian Ocean, the resultant composite signals get weaken towards higher latitudes and hence may not be effective in explaining the processes involving the regional precipitation over the Indian subcontinent. In addition, there is a meridional variation of the mean winds from the equator to the north Indian Ocean. Vecchi and Harrison (2002) examined the intraseasonal variability of OLR with respect to SST anomalies over the Bay of Bengal. They pointed out that positive (negative) SST anomalies on the intraseasonal scale in the Bay of Bengal
exhibit a strong statistical relation with the active (break) periods. The southern part of India lies between two seas: the intraseasonal precipitation variability may have regional characteristics between the east and west coast. Another feature of the Arabian Sea during the monsoon season, which influences the Indian subcontinent, is the Somali Jet. Halpern and Wozeshyn (1999) suggested that the Somali Jet in the Arabian Sea and the rainfall of west coast of India is connected because of simultaneous increases in surface wind convergence and integrated cloud liquid water in the eastern Arabian Sea.

1.4.1 Role of Sea Surface Temperature

Sabai et al. (2003) have shown that despite the weakening relationship of ENSO-Indian Summer Monsoon in recent years, the relationship between SST in some regions of the global oceans and the ISMR is consistent for more than a century and any small variation in this relationship is part of natural oscillations. In different decades, about 55–85% of the variance associated with the ISMR is explained by SST alone and only 15–45% is explained by other boundary forcings and internal dynamics.

Figure 1.3. Climatological mean SST (°C) during (a) boreal winter (DJF) and (b) summer (JJA), for a period 1998–2002. SST data is taken from TRMM.
Change in SST is important in the energy balance and the inter-hemispheric heat exchange over the Asian continent and Indian Ocean in a coupled ocean-atmosphere system (Wang et al. 2003; Clemens and Oglesby 1992). Tropical observations show convective activity increasing sharply above SST's of around 26°C and then decreasing as the SST exceeds 30°C and a maximum is observed when SST is about 32°C. Although some aspects of this relationship are reasonably well understood even though no theory has fully explained the decrease in convective activity above 30°C.

Active (break) phases of precipitation occur by the northward propagation of positive (negative) precipitation anomalies over the Arabian Sea and the Bay of Bengal. Over the windward side of the Western Ghats region during the active phase, the positive SST anomalies over the Arabian Sea formed by suppressed surface latent heat flux and increased downward shortwave radiation flux lead the positive precipitation anomalies. Surface air temperature anomalies follow the SST anomalies and then destabilize the lower atmosphere between 1000hPa and 700hPa. These results indicate that, in the northward propagating dynamical surface convergence, underlying SST anomalies tend to form a favorable condition for convective activity and may sustain enhanced precipitation over the convergence region. This gives enhanced precipitation anomalies over the Western Ghats region that move further northeastward and merge with the northward propagating precipitation anomalies from the Bay of Bengal.

1.4.2 Role of Evaporation over Arabian Sea on Monsoon

The role of oceanic divergence in determining the seasonal cycle in evaporation rate is very important in the viewpoint that the amount of rainfall over India during the southwest monsoon is a function of the amount of water evaporated over the pathways of monsoon as well as orographically
induced convective activity. Wajcowicz and Schopf (2001) in their analysis of Comprehensive Ocean–Atmosphere Dataset (COADS) shows that nearly 90% of the water vapour available to precipitate over India during the southwest monsoon results from the annual mean evaporation field. The seasonal change in direction of airflow, which opens up a pathway from the southern Indian Ocean to the Arabian Sea, rather than the change in evaporation rate is key to explaining the climatological cycle, though the change in latent heating due to seasonal variations is similar to that needed to account for observed interannual to interdecadal variability in monsoon rainfall.

1.4.3 Regional climate modeling of Indian summer monsoon

It has been demonstrated that for the study of the weather or climate features in greater detail, regional models are more suitable than the global models. Computationally it is affordable to increase the resolution of regional models so as to resolve regional climatic features reasonably well. Simulation of the Indian summer monsoon circulation features and the associated rainfall by a numerical model have been the most challenging problems so far. There have been some attempts to simulate monsoon features and extreme weather events over India by regional models. Bhaskaran et al (1996) simulated the Indian summer monsoon using a RCM with a horizontal resolution of 50 km nested with global atmospheric GCM. Their study showed that regional model derived precipitation exceeds 20% than GCM.

The regional climate modeling technique consists of using initial conditions, time-dependent lateral meteorological conditions and surface boundary conditions to drive high-resolution Regional Climatic Models (RCMs). The input data is derived from GCMs (or analyses of observations) and can include greenhouse gas and aerosol forcing. Another method of this technique is to
force the large-scale component of the RCM solution throughout the entire domain (Kida et al 1991; Cooke and LeRow 2000; von Storch et al 2000). Till now, this technique has been used only in one-way mode, i.e., with no feedback from the RCM simulation to the driving GCM. The basic strategy is, thus, to use the global model to simulate the response of the global circulation to large-scale forcings and the RCM to (a) account for sub-GCM grid scale forcings (e.g., complex topographical features and land cover inhomogeneity) in a physically-based way; and (b) enhance the simulation of atmospheric circulations and climatic variables at fine spatial scales. The nested regional modelling technique essentially originated from numerical weather prediction and the use of RCMs for climate application was pioneered by Dickinson et al (1989) and Giorgi (1990). They can provide high resolution (up to 10 to 20 km or less) and multi-decadal simulations and are capable of describing climate feedback mechanisms acting at the regional scale. A number of widely used limited area modelling systems have been adapted and developed for climate applications. More recently, RCMs have begun to couple atmospheric models with other climate process models, such as hydrology, ocean, sea-ice, chemistry/aerosol and land-biosphere models. Two main theoretical limitations of this technique are the effects of systematic errors in the driving fields provided by global models, and the lack of two-way interactions between regional and global climate. Practically, for a given application, consideration needs to be given to the choice of physics parameterizations, model domain size resolution, technique for assimilation of large-scale meteorological conditions, and internal variability due to non-linear dynamics not associated with the boundary forcing (Giorgi and Mearns 1999; Ji and Versekar 1997). Depending on the domain size and resolution, RCM simulations can be computationally demanding, which has limited the length of many experiments. GCM fields are not available at high temporal frequency (6-hourly or higher), as required for RCM boundary conditions, and thus
careful co-ordination between global and regional modelers is needed in order to perform better RCM experiments.

In this study, based on these background information, an attempt has been made to study the spatial and temporal characteristics of the weather and climate of peninsular India having influenced by external factors such as solar forcing, northern Indian ocean SST and moisture, air temperature and zonal wind. A regression model has been developed using parameters like SST, air temperature and zonal components of wind. A regional climate model RegCM3 has been validated and interpreted to study the influence of SST and moisture over this region. Detailed description and methods are discussed in respective chapters.