Southern Hemispheric Features and their Teleconnection with Indian Summer Monsoon

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by

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CERTIFICATE

This is to certify that the thesis entitled ***Southern Hemispheric Features and their Teleconnection with Indian Summer Monsoon*** is an authentic record of the research work carried out by **Mr. Nithin Viswambharan** (Reg. No. 3227) under my supervision and guidance in the Department of Atmospheric Sciences, School of Marine Sciences, Cochin University of Science and Technology. I also certify that the subject matter of the thesis has not formed the basis for the award of any Degree or Diploma of any University or Institution.

I also certify that Mr. Nithin Viswambharan has passed the Ph. D qualifying examination conducted by the Department of Atmospheric Sciences, Cochin University of Science and Technology in March 2011.

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DECLARATION

I hereby declare that the thesis entitled ***Southern Hemispheric Features and their Teleconnection with Indian Summer Monsoon*** is an authentic record of the research carried out by me and no part of this thesis has been submitted to any University or Institution for the award of any Degree or Diploma.

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PREFACE

Southern hemisphere is entirely distinct from the northern hemisphere in many aspects, which is well reflected in atmospheric and oceanic properties. Moreover, it is interesting to note the variation after the documented climate shift of second half of the 20^{th} century. The variability in the upper stratosphere is also exciting, as it is known that the stratospheric anomalies can influence tropospheric features. The leading mode of variability in southern hemisphere is the Southern Annular Mode (SAM). This mode accounts of 10% variability in global scale and is one of the strong influential factors in many variables of southern and northern hemispheres. The Southern Annular Mode is known to influence the rainfall pattern over northern hemisphere. It is therefore, interest to note the teleconnection of southern hemispheric changes on tropical weather systems, especially the influence of SAM on the summer monsoon over India.

The weakening relationship of El Nino and summer monsoon also supports to find the prevalence of new teleconnecction. Both observational and modeling studies have supported the importance of cross equatorial flow over Indian Ocean during summer monsoon, indicating the influence of southern hemisphere. The tropical circulation changes associated with this southern extratropical mode give further insight of the circulation, which is important in the interpretation of physical and dynamical behaviour of atmosphere and ocean.

Not much study has been carried out extensively on the variations of southern hemisphere and its effect on tropical climate. Therefore it is an interesting area of investigation. A detailed study on the effect of Southern Annular Mode and the modulation of North Atlantic Oscillation on Indian summer monsoon has been carried out. The study is relevant main because the variability of monsoon and its predictability is integral in the socio-economic development of India. These variability and their effects were studied and reported in the thesis.

The thesis consists of eight chapters, in which the first chapter contains an overview of southern hemisphere. In this chapter, variability in southern hemisphere is described along with Indian summer monsoon and its teleconnection. The different types of data sets used and various methodologies adopted in the present thesis were described in Chapter 2. The period of climate shift and the magnitude of anomalies after the climate shift, which extended from troposphere to stratopause level, were investigated in detail and presented in chapter 3.

Chapter 4 depicts the recent trend and variability in southern stratosphere. The higher order variability during various months and the frequency of extremity is included in this chapter. It also describes the role of ENSO and stratosphere using the Nino indices. Climatology of divergence and convergence after the documented shift is reported in chapter 5. It also identifies the role of Southern Annular Mode in changing the tropical circulations like Walker and Hadley.

Southern extratropical connection to Indian summer monsoon through the modulation of SAM is presented in Chapter 6. The variability associated with phase change of SAM on summer monsoon parameters like wind, vertical velocity, humidity and rainfall is also described. Chapter 7 deals with the modulation of SAM-Monsoon link through North Atlantic Oscillation. This chapter gives insight over the Indian summer monsoon variability due to the simultaneous effect of SAM during June and NAO during April. The final chapter 8 summarises the outcome of the thesis work, and give future scope of the study. References in alphabetic order are included at the end of thesis. List of abbreviations used in the thesis are also presented.

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Chapter 6 Signature of extratropical influence on the summer monsoon over India

Chapter 1

Introduction

1.1 General features of southern hemisphere

Southern hemisphere is the segment of Earth that has more water bodies than landmass. About 80.9 % of the southern hemisphere is covered by water, including South Pacific, South Atlantic, Indian Ocean, Tasmania Sea and Weddell Sea, while the land comprises only 19.1%. The continents in this hemisphere are Antarctica, southern part of Africa, South America and Australia (fig 1.1). Only 10 % of the total world population lives in Southern hemisphere. This hemisphere is distinct with different varieties of flora and fauna. Amazon and Madagascar are areas located in southern hemisphere are rich in biodiversity than any part of the world. The Austral hemisphere shows diversity in many aspects that is well reflected in the climate and weather over this region. The coldest continent in the world is in this hemisphere, which receives less than five centimeters of precipitation annually. The cold winter season begins in June and persists till August and summer period is from December to February.

Based on climate, Austral hemisphere is divided in to different regions. Tropic is the area between equator and Tropic of Capricorn, this area has warm temperature and precipitation. Temperate zone is in between Tropic of Capricorn and Antarctic Circle at 66.5^0 S. The area has high precipitation, cold winters and warm summers. South of the temperate zone is the Antarctic Circle, the area covered by cold Antarctic land mass and are colder than the Arctic. Quite amazingly, it is found that this cold Antarctica has warm period that occurred about 15.7 million years ago and lasted for a few thousand years.

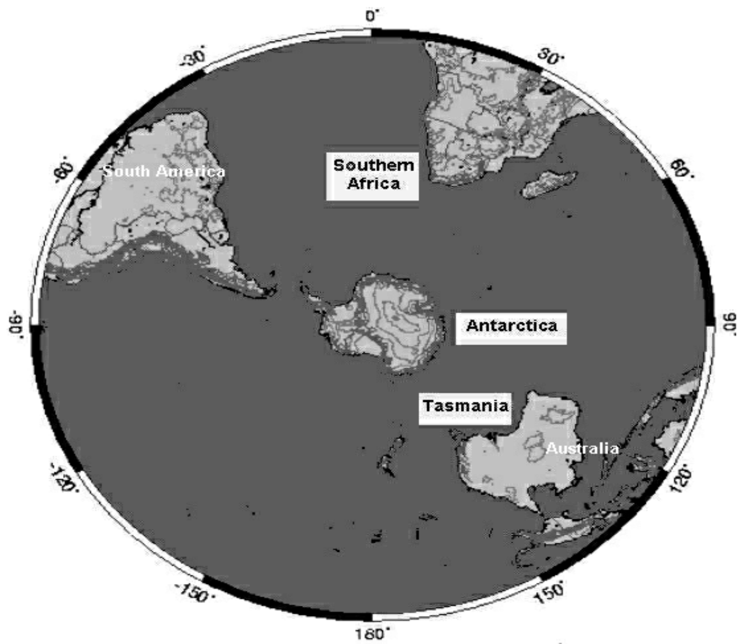

Figure 1.1. Schematic representation of Southern Hemisphere

Southern hemisphere identity is entirely different from the northern hemisphere in many aspects. The climate in the southern hemisphere does not vary like the northern hemisphere. This is due to the vast Ocean in this hemisphere, which heats and cools slowly as a result the effect of climate perturbation is not intense like in boreal hemisphere, where land-sea contrast is substantially high. Another noticeable thing is the deflection of an object caused by Coriolis force. The object moving away to the Antarctic Circle is deflected to the left. In southern hemisphere, clockwise rotation is related to low-pressure centre where it is anticlockwise in northern hemisphere.

The differences between the southern hemisphere (SH) and northern hemisphere (NH) are also observed in the stratosphere, where they differ in the dynamics, chemistry and radiation properties. Due to the low temperature in the southern polar cap region, the winter-time SH polar vortex is much stronger and is long lasting than its boreal counterpart. The austral winter-time stratospheric polar vortex appears a month earlier in autumn than its NH counterpart, and the vortex persists longer into the spring than the NH. The break-up of polar vortex is mainly through the planetary waves which deposit its momentum and energy to the stratosphere and thereby weaken the polar vortex, but the wave activity is less in southern hemisphere due to the absence of strong generating mechanism like in northern hemisphere.

Southern hemisphere has its own unique identity with respect to northern hemisphere that can be noticed in the climatology of wind and temperature pattern. During winter (July), the southern hemisphere troposphere and stratosphere are dominated by westerly winds (see fig 1.2). The westerly

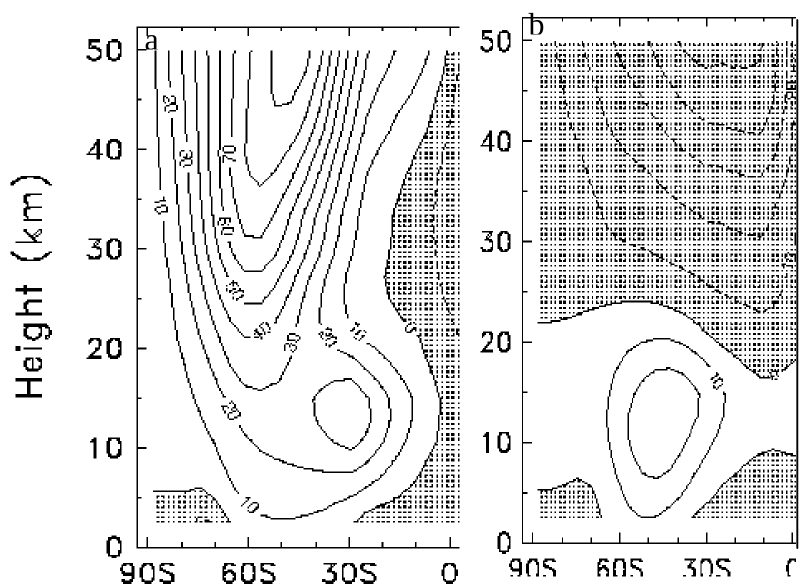

Figure 1.2. Zonal wind structure for a) July and b) January of Southern hemisphere (adapted from SPARC)

wind maxima in the upper troposphere level is 5m/s weaker and is 2-3^0 latitude nearer the equator than NH. Towards the polar region, the winter westerlies are quite stronger than the NH. The distribution of wind pattern in the two hemispheres is different during summer. During summer (January), the tropical easterlies in the middle and upper troposphere are weaker in SH while the westerlies in the subtropics are stronger than NH.

The summer hemisphere has cold tropopause and warm stratopause region. The mesopuase region is extremely cold in summer hemisphere. During winter, the midlatitude troposphere is warm and the polar region is extremely at very low temperature. During winter, Antarctic temperature are quite lower compare to Arctic winter. The winter polar vortex is quite stronger in SH than in NH. Southern hemisphere polar vortex is less disturbed compare to NH, so the event known as Sudden Straospheric Warming (SSW) is also less intense in SH

1.2 Major events in the southern hemisphere

Southern hemisphere is in the focus of attention due to its peculiar feature like Antarctic Ozone hole and Sudden Stratospheric Warming (SSW). These feature alter the existing dynamical and chemical behaviour in the Austral hemisphere and are important in global climate.

1.2.1 Sudden Stratospheric Warming

Sudden Stratospheric Warming is one of the dramatic phenomenon occurring in the stratosphere of high latitude. During cold winter, temperature in the stratosphere suddenly raised upto 70^0 C within a week, and on certain occasions the zonal wind change its direction from westerlies to easterlies. The anomalous temperature variation without overturning of zonal wind is known as *Minor warming* while the temperature perturbations with phase change of zonal wind is termed as *Major Warming*. Minor warmings are observed in Southern hemisphere during winter. But the Major warming event is unusual in the Southern hemisphere because the winter vortex is stronger and also due to the absence of large Planetary waves, in this part of Earth.

Due to the presence of strong vortex, Austral hemisphere is noted for minor warming, where temperature increases but the zonal wind does not change its direction. But on September 2002, the southern hemisphere experienced a major warming which was unusual for this hemisphere for a long period. The change in polar vortex occurred in a dramatic manner. On September 20 (fig 1.3), the polar vortex get elongated, and it got weakened in following

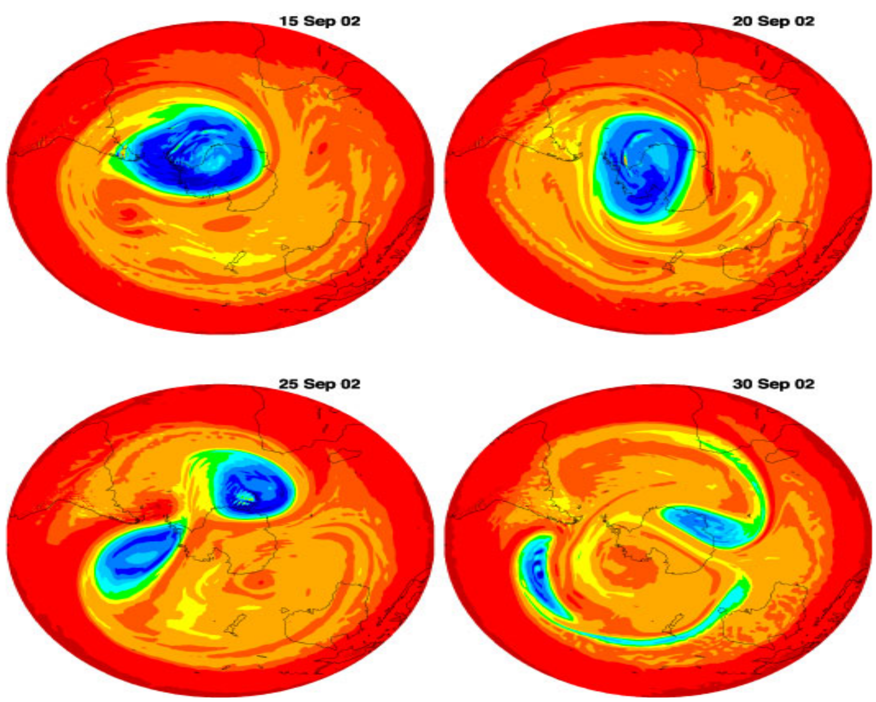

Figure 1.3. Polar vortex breakup on September 2002 (adapted from SPARC)

dates. The vortex split in to two on September 25 and resulted in the intrusion of warm air to the South Pole. After the split the vortices are weakened rapidly on September 30. It is noted that the zonal wave number 2 reached the peak when the polar vortex split in to two. The major warming also resulted in the split of ozone hole into two.

1.2.2 Antarctic Ozone Hole

Another important phenomenon that brought the southern hemisphere in focus during 1970 is due to the discovery of ozone hole. Antarctic ozone hole refers to the seasonal (spring) depletion of stratospheric ozone in a large area over Antarctica (Farman et al., 1985; Solomon, 1999; Staehelin et al., 2001). This discovery guide the world to rethink in the use of ozone-depleting substances (ODS). The seasonal thinning of ozone layer is shown in figure 1.4. Intense thinning of ozone layer occurs due to the emission of man-made halocarbon to the atmosphere and this chemical is transported to high and midlatitude of lower stratosphere. In the extreme lower temperature they find substrate in cold clouds and was dormant in winter.

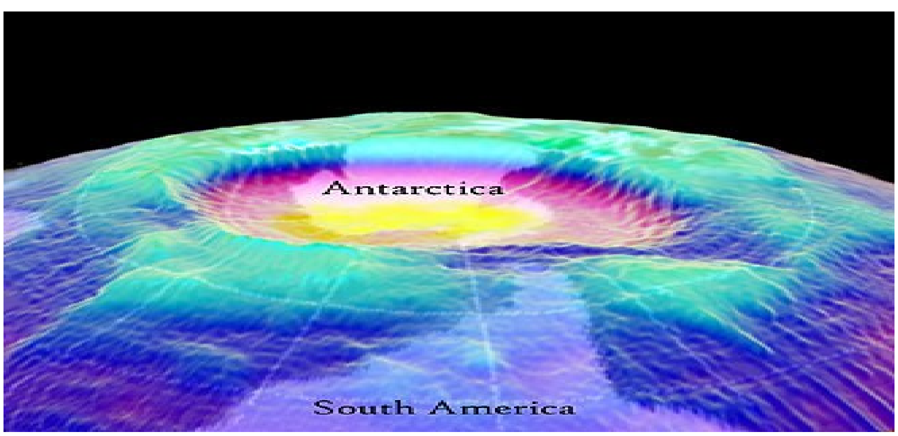

Figure 1.4. Seasonal thinning of ozone layer above Antarctic (adapted from NASA)

During spring the ice melt and the ozone destroying molecule get released and break apart the molecular bonds in UV radiation-absorbing ozone. As a result highly destructive rays penetrate to the lower atmosphere thereby affecting the living species. This event is noted mainly in Antarctica region in spring (September-October) due to the extreme low temperature, forming polar stratospheric clouds that will provide substrate to the ozone depleting compounds.

1.3 Southern Annular Mode

Southern Annular Mode is the normalised difference in the zonal mean sea-level pressure between mid and high latitudes. The sea level pressure pattern associated with SAM is a nearly annular pattern with a large low pressure anomaly centered on the south pole and high pressure anomalies at mid-latitudes and it is the principle mode of variability in Southern hemisphere (Thompson and Wallace, 2000; Visbeck and Hall, 2004). Southern Annular Mode is also known as Antarctic Oscillation (Gong and Wang, 1999) or Southern hemisphere Annular Mode (SHAM).

During the positive phase of SAM, pressure is lower than normal in high latitude and higher than normal in midlatitude. The strong westerlies undergo changes during the positive and negative phase of this high latitude mode and thereby affect the Oceanic circulation (Hall and Visbeck, 2002; Oke and England, 2004). The westerlies are stronger than normal over the southern oceans in the positive phase and weaker than normal during its negative phase. Pattern of the pressure variations associated with the positive phase of the SAM is depicted in figure 1.5.

In recent decades, SAM is showing positive trend (Marshall, 2003), and it is predicted that the upward trend will persist in future decades (Cai et al.,

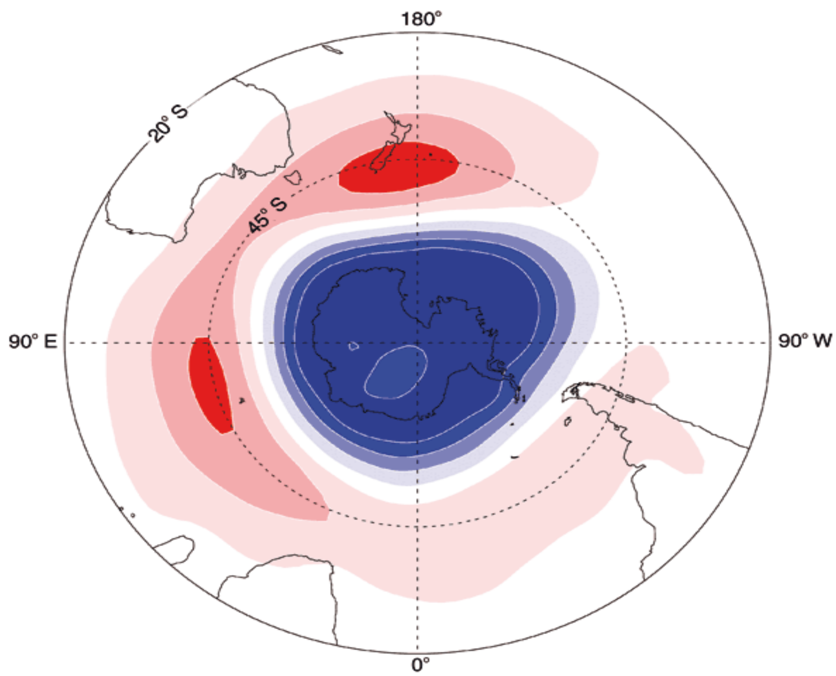

Figure 1.5. Pattern of the pressure variations associated with the positive phase of the SAM. Blue shading indicates below-average pressures and red shading indicates above-average pressures (adapted from *http://www.niwa.co.n*).

2003; Kushner et al., 2001). The positive trend of this high latitude mode is as a result of stratospheric ozone depletion and emission of greenhouse gas (Hartmann et al., 2000; Thompson and Solomon, 2002; Gillett and Thompson, 2003; Marshall et al., 2004).

1.3.1 Effect of SAM on other meteorological parameters

On interannual timescales southern annular mode (SAM) contributes a significant proportion of SH mid-latitude circulation variability (Hartmann and Lo, 1998; Kidson, 1999; Baldwin, 2001). The southern annular mode is linked with variability of many parameters of northern and southern hemisphere. It has been noticed that the SH cyclone track undergo north-south migration during the alternative phase of SAM (Kidson and Sinclair, 1996). Model studies predict that the southern Ocean takes lesser amount of carbon dioxide during the positive phase of SAM (Lenton and Matear, 2007).

Extensive study has carried out by several scientists to understand the effect of SAM. One such study explored the impact of this extratropical mode on surface wind, sea surface temperature and chlorophyll concentration in the Southern Ocean. The study concludes that the enhanced westerlies during positive phase affect the Ekman transport and the SST anomalies. They even noted that the chlorophyll concentration is significantly correlated with phase of SAM (Lovenduski and Grube, 2005). Temperatures also get modified due to this mode, significant temperature anomalies are observed in Australia due to the effect of Southern Annular Mode (Hendon et al., 2007).

Sea ice content over the Antartic region has varied due to the impact of SAM, during its positive phase decrease in ice area in the Weddell Sea and Antarctic peninsula and increase in the Ross and Amundsen Sea is reported (Lefebvre and Goosse, 2005). This oscillation play enhanced role in the intraseasonal oscillation over tropic to extratropics of the Southern and Northern hemisphere (Carvalho et al., 2004). This mode is known to influence the rainfall over many regions. The effect of SAM on Australian monsoon is observed for a period from 1958 – 2002 and it has been noted that the SAM has in-phase relationship with north Australia and has inverse relationship with southern Australia. There is significant association between winter and spring rainfall over South America (Silvestri and Vera, 2003) and SAM, due to the impact of this mode, significant reduction in rainfall is observed in southwest Western Australia (Li et al., 2005). An inverse relationship between SAM and winter rainfall over South Africa is identified (Reason and Rauault, 2005).

1.3.2 Impact of SAM on northern hemisphere

It is well explicit that the Southern Annular Mode has high influence in the temperature, circulation and rainfall properties of southern hemisphere. But the question arises whether this mode can account for the northern hemisphere variability. It is reported that, this high latitude mode accounts for 10% of global variance (Trenberth et al., 2005), several studies also has connected the boreal hemisphere variability with extra tropical phenomenon of SH. It is observed that the global monthly variability has linked with SAM that is seen active in all months compared to the oscillation in northern counterpart, known as North Annular Mode. North Annualar Mode is the leading mode of variability in northern hemisphere which has strong teleconncetion with many Oceanic and atmospheric properties. The NAO modulates the tropical circulation during its phase change.

Summer rainfall over Yangtze river shows significant positive correlation with the boreal phase of SAM (Nan and Li, 2003). Positive polarity of SAM favours weakening of East Asian Summer Monsoon and westward expansion of western subtropical high. Further the variability play a key role in modulating the vertical velocity, specific humidity and water vapour flux and thereby precipitation. Winter monsoon over China has exhibited relation with this high latitude mode. During the positive phase of SAM in Autumn, anomalous change occur in Hadley cell associated with SST anomaly and influence the lower troposphere over China and weaken the monsoon pattern existing there (Zhiwei et al., 2009).

1.4 Climate change in southern hemisphere

Change in the mean state of variables for a decade or longer due to anthropogenic origin or natural forcing is termed as climate change. This climate change has noticed in the 20th century in many atmospheric and oceanic parameters. Compared to northern hemisphere the variation in southern hemisphere ocean is more intense in the second half of the 20th century. Several model and observation studies have suggested the climate change is will evident in the Southern ocean during recent decades (Arbic and Owens, 2001; Wong et al., 2001; Wainer et al., 2004; Sprintall, 2008).

Not only oceanic properties have varied in SH, the impact is well evident in troposphere and in stratosphere. Anomalous variation is observed in Antarctican tropopause due to phase change in SAM (Santer et al., 2003). It has been reported that the midlatitudes of SH is warmer than the equator while high latitude and Antarctic have cooled in recent decades (Parkinson 2006; Chapman and Walsh 2007). Strengthening of westerlies and its contraction towards South Pole is also observed (Marshall, 2007).

Stratosphere over Antarctica region has showing anomalous variation. The stratospheric anomalies are observed seriously by researchers due to its downward propagation to the troposphere and its known influence to circulation pattern and thereby playing prominent role in changing the climate. The lower Antartican stratospheric region has shown large trend during 1979 –2007 period due to ozone changes (Randel et al., 2009). Intense variation in temperature properties is observed in many southern hemispheric variables since 1970. Sea ice had experienced a rapid decrease during the 1970s, followed by a slow gain, indicating higher variability in the Antarctic sea ice during recent decades (Parkinson, 2006). Southern Annular mode had shown significant upward trend since 1970. This trend is related to anthropogenic activities due to CO_2 and change in ozone concentration (Sexton, 2001; Miller et al., 2006; Cai and Cowan, 2007). It is reported that the increases of greenhouse gases and ozone layer changes in future decade may have adverse consequence in Antarctican region by melting of sea ice (Shindell and Schmidt, 2004).

1.5 Meridional circulation in southern hemisphere

The meridional circulation is a response to the differences in insolation between low and high latitudes resulting in the transfer of energy. During the seasonal march of Sun the centre of convergence and divergence migrate to its respective hemisphere (fig 1.6). Meridional circulation in the southern hemisphere has three cells. viz., Hadley Ferrel and Polar cell. The annual mean tropical Hadley cell in the southern hemisphere is stronger than its counterpart in the northern hemisphere. During the solstice period, the area of rising and subsidence change according to the shift of mean position of Sun. Hadley cell has ascending in the equatorial region and sinking motion near 30^0 S. During Austral summer, the ascending motion of the Hadley cell occurs south of the equator and it subside to the south of 30^0 S. In winter the ascending branch is to the north of equator while subsidence occur to the

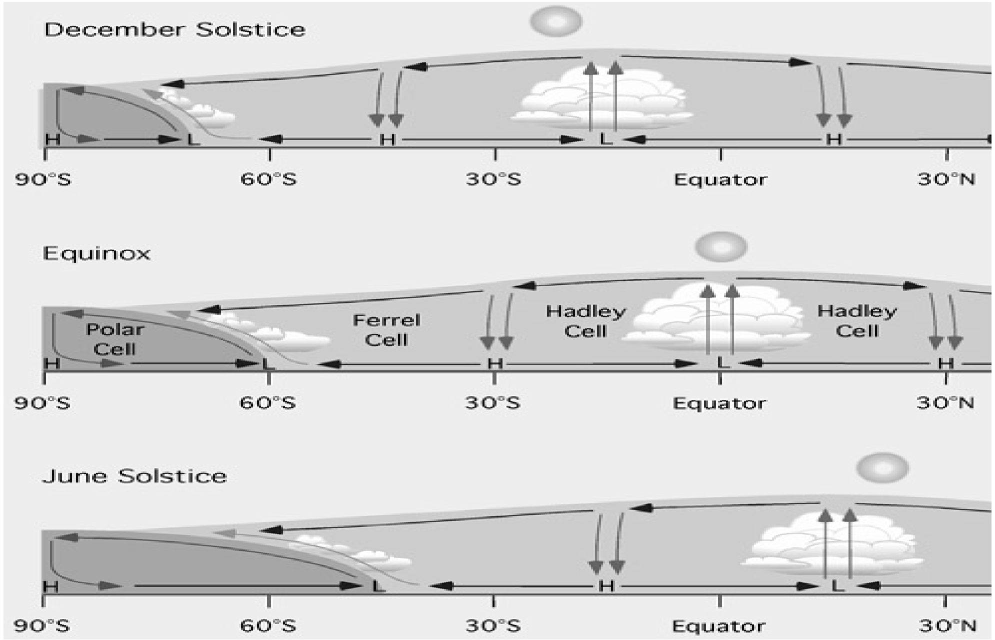

Figure 1.6. Schematic representation of Three- cell structure of meridional circulation in southern hemisphere (adapted from http://www.eoearth.org/ article /Tropical weather and hurricanes)

equatorward of 30^0 S. The change in Hadley circulation also shifts the mean position of other cells. The southern hemisphere winter Hadley cell is stronger than the northern hemispheric cell. During the equinoctial transition period Austral hemispheric cell are double intense than the boreal cell. The differences in the thermal properties give rise to region local Hadley circulation which is well evident in the Pacific and also during the summer monsoon period over India.

The Ferrel cell extend from 30^0 S to 60^0 S which is thermally an indirect cell, due to rising motion in the cold area and subsidence over the warm region. Another cell that exist in the Austral hemisphere is the Polar cell. The Polar cell extend from 60^0 S to polar region, this cell bring cold air equatorward. Both the Ferrel and Polar cells change their mean position, and are generally weak in intensity throughout the year. Recently, changes are observed in the Hadley cell circulation that has influenced the existing circulation pattern and modulated the rainfall characteristics to a great extent.

In addition to this meridional circulation, the southern hemisphere has another meridional circulation known as Brewer Dobson Circulation (BDC). This BDC consists of three parts (fig 1.7), rising motion from troposphere to

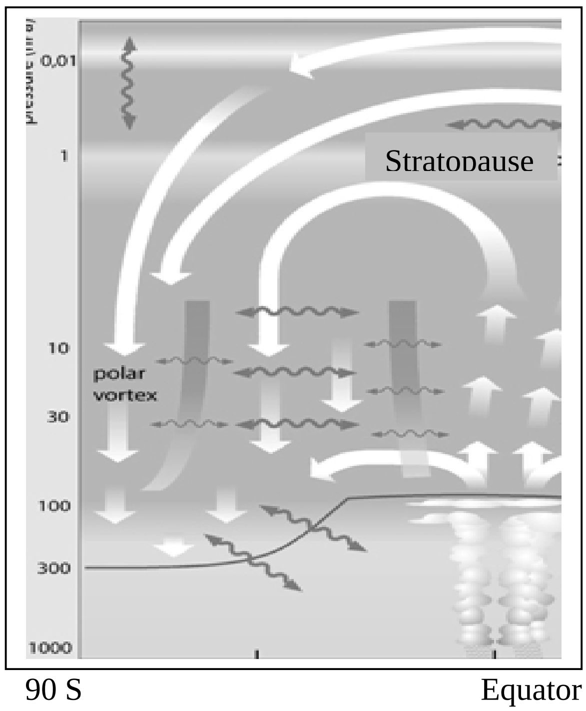

Figure 1.7: Schematic representation of Brewer Dobson Circulation (adapted from Frankfurt University)

stratosphere in the tropical region, secondly, poleward transport in the stratosphere and the third part is the descending motion in the midlatitude and polar stratospheric region. The midlatitude sinking part transport back in to the troposphere while the poleward descending air accumulates in the lower stratosphere. The BDC transport is weak in the SH due to the strong polar vortex and lack of planetary scale wave activites.

1.6 Walker Circulation

The pressure and temperature difference in the western and eastern tropical Pacific result in the Walker circulation. The warm west Pacific and cool east Pacific creates pressure gradient with flow from east to west along surface and west to east in upper atmosphere completing the Walker circulation. The convergence in the western Pacific is associated with rainfall and the divergent in the east Pacific causes dry weather. The circulation undergoes changes and lead to the evolution of La Nina in its strong phase and El Nino in its weak phase (fig 1.8). During El Nino, low level convergence is shifted to

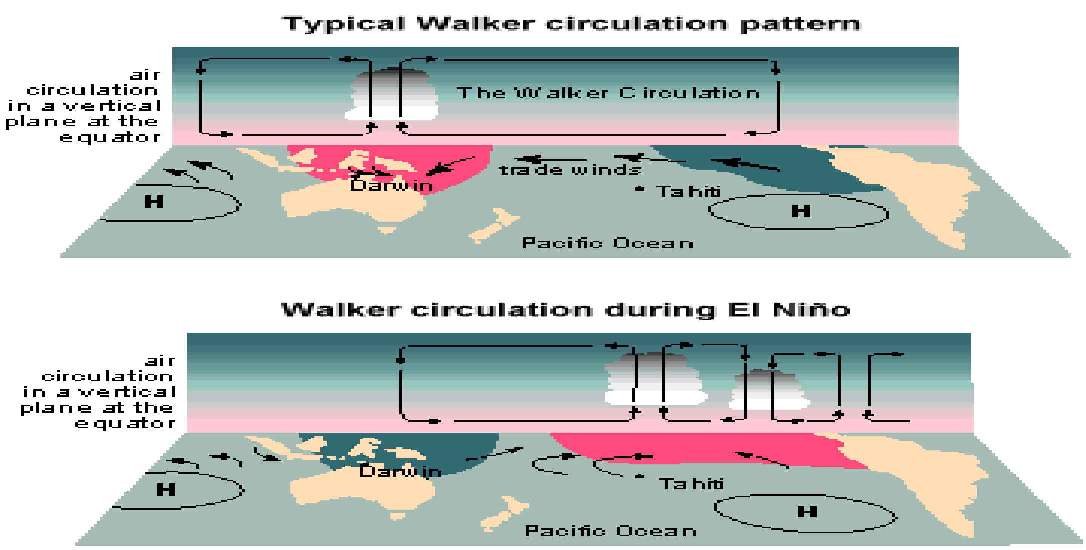

Figure 1.8. Schematic representation of Walker circulation during El Nino and La Nina phase (adapted from *http://www.bom.gov.au/lam/climate/levelthree/analclim/elnino.htm*).

the east Pacific and the divergence occur in the west Pacific region. During La Nina, the low level convergence occurs in the west Pacific and divergence is observed along the east Pacific. On the basis of Indo-Pacific sea level pressure it is concluded that the Walker circulation is undergoing a weakening trend and the models also have supported this observational finding. This weakening trend has modulated the thermal and circulation properties of tropical Pacific Ocean (Vecchi et al., 2006).

1.7 Asian summer monsoon

One of the large scale and complex circulation pattern is the Asian summer monsoon. This monsoon is one of the major system that decides the dry and wet spells of rainfall over the most populated regions. This monsoon mechanism arises due to the heat difference between continent and Ocean, Coriolis force by rotation of earth, moisture transport, and meridional temperature gradient and also due to various other phenomenon. One of the fundamental cause of monsoon cycle is the cross-equatorial pressure gradient due to the differential heating (Webster, 1987). During the northern summer, winds flow from the southern hemisphere, transporting moisture and thereby accelerating the precipitation amount over the south Asian region (fig. 1.9).

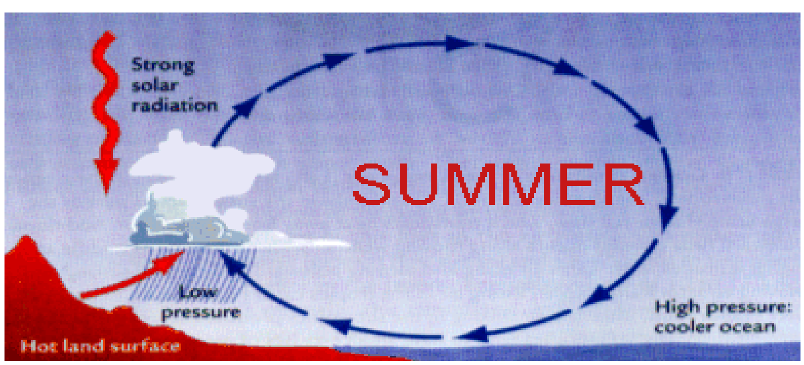

Figure 1.9: Summer circultation near the equatorial region (adapted from *http://www.ncclimate.ncsu.edu/secc_edu/images/monsoon.gif*)

The Asian summer monsoon can be divided in to two, the South Asian (Indian monsoon) and the East Asian monsoon system. Both of them are independent but interact each other through transfer of energy exchange, oscillation and moisture transport (Zhu et al., 1986). It is independepent because the East Asian monsoon is not an extension of Indian monsoon. Study has shown that the Asian summer monsoon can include the western north Pacific region. Thus the Asian-Pacific monsoon can be subdivided in to Indian Summer Monsoon (ISM), the western North Pacific summer monsoon (WNPSM) and the East Asian summer monsoon (EASM) which is schematically depicted in figure 1.10. The ISM and WNPSM are tropical monsoons in which the low level wind change its direction from easterlies to westerlies, and the EASM is a subtropical system where wind change its direction from northerlies to southerlies (Wang and Lin 2002).

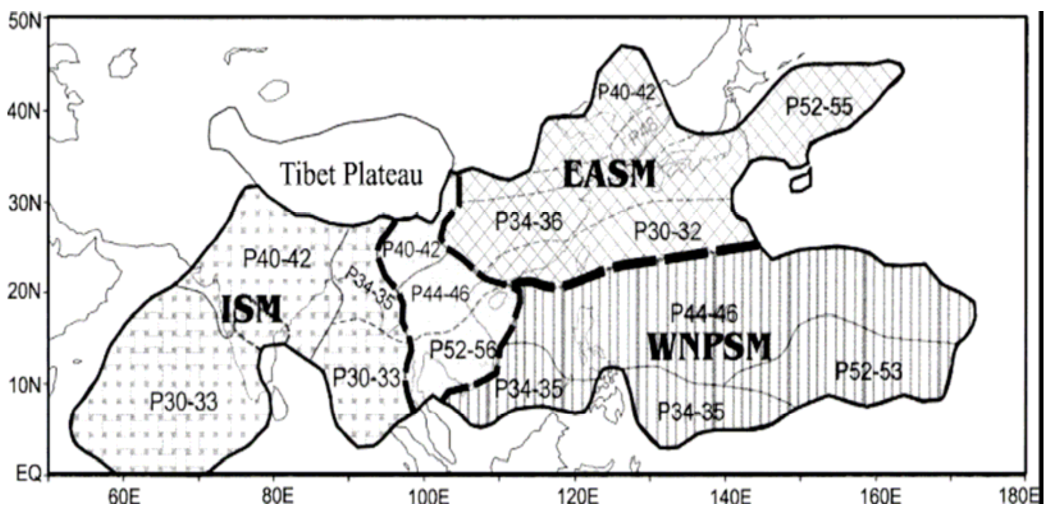

Figure 1.10. Schematic representation of Asian summer monsoon (adapted from Wang and Lin, 2002)

Onset phase of Asian monsoon is complex because it is regional dependent. Ding (2004) has split the monsoon onset dates in to four stages. During the first stage the earliest onset occur during late in April and in May in the central Indo-China peninsula but some times it may be in southern or western part of peninsula. On stage 2 (mid to late May), the monsoon advance northward to the Bay of Bengal region. During stage 3 (June), the monsoon onset begins on Indian region and also in the East Asian region. On stage 4, it will reach to the north China and the Korean peninsula that occurs during July. Fluctuation is noticed in the Asian summer monsoon, it basically arises due to meridional temperature gradient (Yanai etal., 1992; Meehl, 1994; Li and Yanai, 1996; Wu and Zhang, 1998).

1.8 Indian summer monsoon

One of the strongest monsoon system in northern hemisphere is the Indian summer monsoon (ISM). This system brings copious rainfall all over India during June to September. Around 80% (Selvaraju, 2003) of the annual rainfall occurs during southwest monsoon period. This system starts with cross equatorial flow and wind blowing southwest direction associated with heavy rainfall all over India (fig 1.11). The migration of Indian monsoon

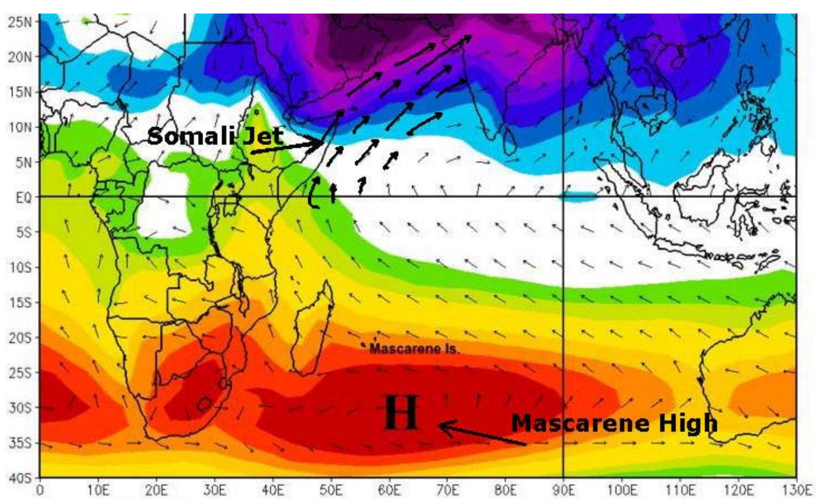

Figure 1.11. Pressure and wind pattern during Indian summer monsoon (adapted from *http://www.cdc.noaa.gov/Composites/Day/)*

system depends on the synoptic scale features. The western and central India receives more than 90% of annual precipitation, on the other hand the southern and northwestern India receives 50 – 70% of their annual rainfall during ISM period. The fluctuation in the monsoon system affects the production of agriculture and thereby it can alter the economic balance of India.

The failure of monsoon brings famine, while strong monsoon leads to flood. The delay in monsoon onset adversely affects the agriculture sector. Extreme precipitations do occur in India that also lead to economic and human loss. During summer monsoon, heavy rainfall occurs in the west coast of peninsula where the rainfall is related to the orography and over the northeastern region. High rainfall is also observed around 20^0 N, which is considered as core monsoon zone because the variation of ISM rainfall are highly correlated with variation of rainfall over this zone (Sikka and Gadgil, 1980). Rainfall over Indian region during the summer monsoon mostly occurs in association with development of convective systems and its propagation to the subcontinent. There are certain features that control the amount and spatial distribution of rainfall (fig. 1.12).

One among the parameter that control the rainfall pattern is the Low Level Jet (LLJ) at 850 hPa level (also known as Somali jet stream), this flow are vital in the distribution and intensity of rainfall by carrying moisture to the subcontinent. The flow pattern passing through India at 100 hPa level near 13.5^0 N is also important in the distribution of rainfall. This upper level flow is the Tropical Easterly Jet (TEJ) stream that runs from Vietnam to the west coast of Africa. The movement of this jet is associated with active and break spell of rainfall.

Another component of monsoon circulation is the monsoon trough extending from heat low over Pakistan to head Bay of Bengal. Monsoon trough is an east west oriented semi-permanent feature that maintain the precipitation activity in India. The trough in its normal position accounts for well-organized rainfall where its undulation to north results in break condition

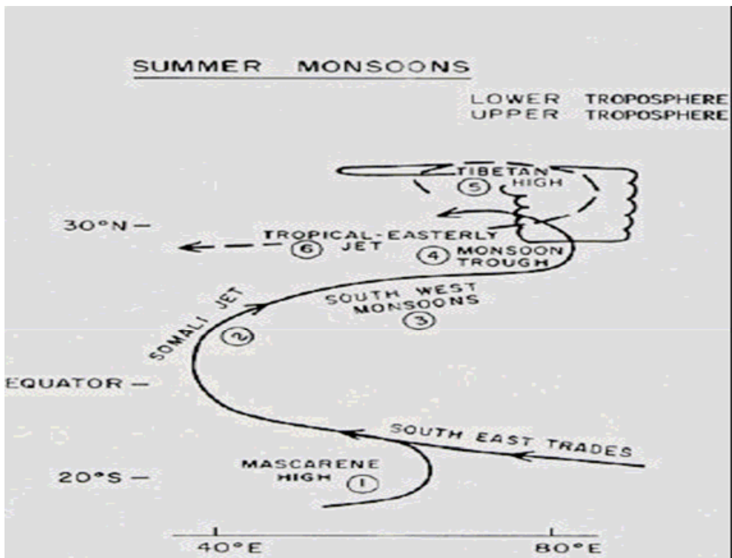

Figure 1.12 Schematic representation of Monsoon (*http://www.cdc.noaa.gov/Composites/Day/ circulation*)

over India, during that period Himalayan foothills experience heavy rainfall. At the same time, the monsoon low and depression embedded in trough enhances the rainfall. Dislocation of the monsoon trough can be identified with the presence of blocking high over the Caspian Sea and also through the meridional flow in upper troposphere of northeastern hemisphere.

Monsoon low and depressions developing in the Bay also contributes to the summer rainfall. The monsoon low and depression travel westward causing heavy rainfall in India. The formation of this system shifts during different months. Earlier advancement of southwest monsoon is favoured by the formation of low or depression in June. In July, they are formed north of 18^0 N in the northwest Bay while in September it from in southward of central Bay. Usually two depressions form each of the monsoon seasons.

Tibetan anticyclone, which is seen in middle and upper troposphere during summer monsoon period, has a strong hand in controlling precipitation. The shift of anticyclone to south adversely affects the Indian monsoon. The anticyclone also moves towards west from its climatological position and affects the monsoon properties. Presence of Mid-Tropospheric Cyclones (MTC) in the northern parts of west coast of India also influences the deep convection during monsoon.

1.8.1 Onset characteristics of Indian monsoon

Monsoon onset over India expected to begin in early June over the southwest India. The onset is dependent on the changes in the circulation features of the lower and upper troposphere (Pearce and Mohanty 1984, Ananthakrishnan et al., 1983, Joseph et al., 1994). The primary driver of the Indian monsoon is the pressure gradient between southern (Mascarene high) and northern hemisphere (heat low over Pakistan). Li and Yanai (1996) has concluded that the reversal of land–sea thermal contrast and large temperature increase over the Tibetan Plateau in May–June help for ISM onset. Goswami et al. (2006) suggested ISM onset index based on the reversal of the large-scale meridional temperature gradient in the upper troposphere. Strengthening of low-level wind over the low-latitude Indian region has been noticed to be a good indicator of the ISM onset (Taniguchi and Koike, 2006; Joseph et al., 2006). During the monsoon onset time a band of deep convection in the east-west direction passes through the southern tip of India (Sikka and Gadgil, 1980). After the onset of monsoon, moisture transport get well organized with cross equatorial flow in the western part of equatorial Indian Ocean thereby inducing rainfall mechanism in India.

1.8.2 Mean rainfall during summer months

The temporal and spatial distribution of rainfall during June to September has large scale variability which is evident from the subdivisional rainfall over India. This arises due to the fluctuation in synoptic scale activities. It is better to understand the monthly mean of rainfall distribution during each month. During June heavy rainfall greater 15 mm/day occurs in the west coast (fig. 1.13) and in the northeast regions. Rainfall less than 5mm/day

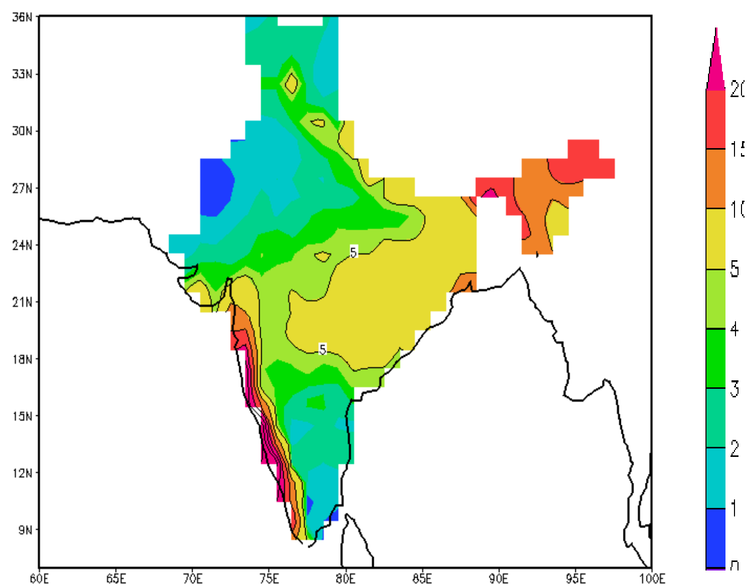

Figure 1.13. Mean rainfall (mm/day) over India during June

occurs in the southeast coast of India. Precipitation of about 2-4 mm/day occurs to the north of 20^0 N. In July, the precipitation enhances all over India. During this month widespread rainfall occurs all over India (fig 1.14). In the southwest coastal station and northeast India rainfall about 28mm/day is observed. About 10 mm/day of rainfall is seen around 20^0 N. The enhanced

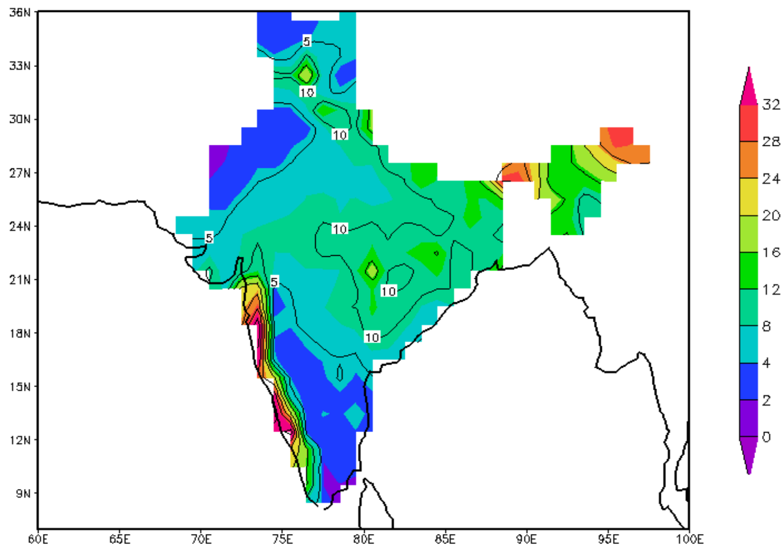

Figure 1.14. Mean rainfall (mm/day) over India during July

precipitation during July is well evident along southwest coast, northeast regions, north India and along 20^0 N of India.

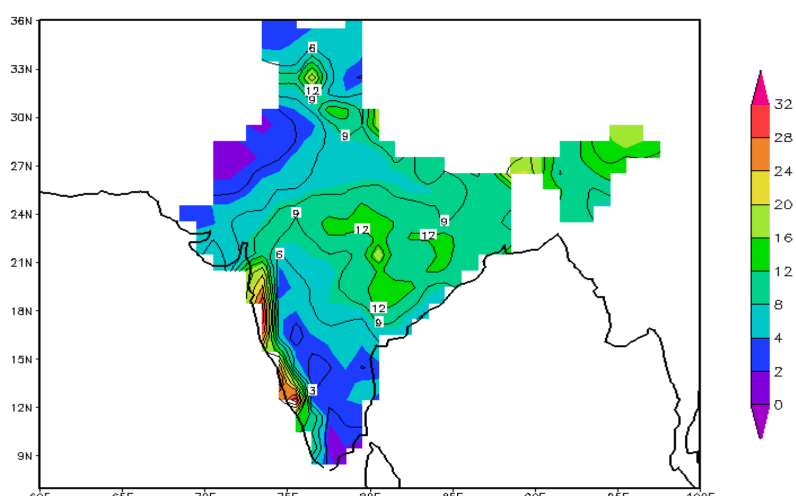

Figure 1.15. Mean rainfall (mm/day) over India during August

The precipitation during August seems to increase near 20^0 N than July precipitation by about 2mm (fig 1.15). Along the southwest coast in between 10^0 N to 21^0 N rainfall exceed to about 24mm/day. Precipitation greater the 12mm/day occurs in many parts of northeast India. Precipitation less than 2 mm/day is observed along northwest and southeast India.

As the monsoon recedes in the month of September the intensity of rainfall decrease which is evident in all over India. Most of the central, and south India shows precipitation less than 15 mm/day (fig 1.16). There are areas of northeast India were precipitation greater than 15 mm/day is observed. Northwest India shows precipitation less that 2 mm/day during the monsoon retrieval period.

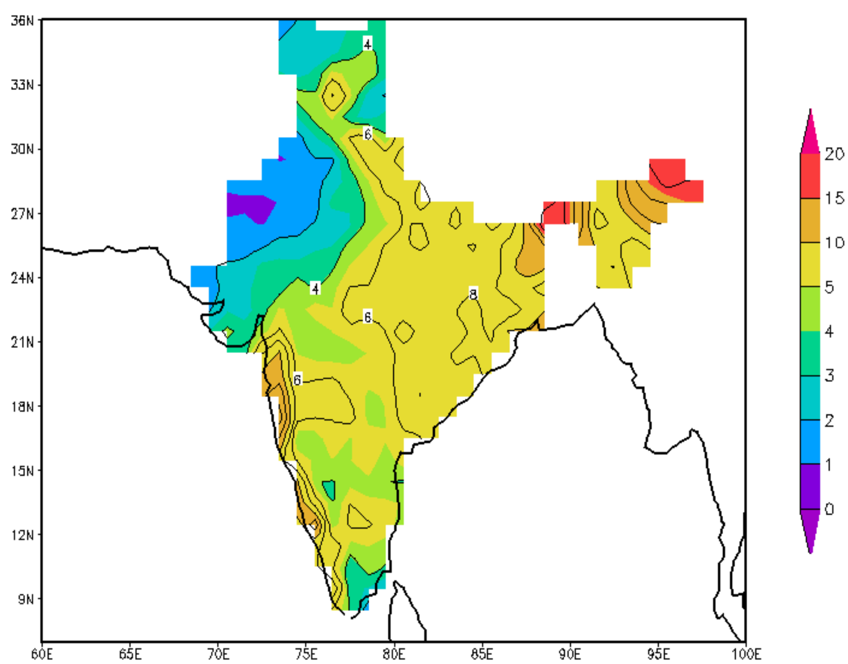

Figure 1.16. Mean rainfall (mm/day) over India during September

1.8.3 Interannual and Intraseasonal variability of Indian Summer Monsoon

Indian monsoon shows variability in both interannual and intraseasonal timescales. Interannual variability refers to unevenness within years, while intraseasonal is the variability within season. These changes are important to understand the scale of precipitation and its distribution over Indian domain.

1.8.3.1 Interannual Variability (IAV)

The year to year variation has high potential impact in the Indian economy. Interannual variability of the tropical climate is partially governed by internally generated low frequency (LF) oscillations in addition to being forced by slowly varying sea surface temperature, soil moisture, sea ice etc. Non-Linear interaction between high frequency oscillation, non–linear interaction of intraseasonal oscillation, interaction between flow pattern and topography lead to the generation of low frequency variability. Several studies were carried out to find the mechanism of the IAV of Indian summer monsoon. Extensive study of seasonal mean Indian rainfall had shown that the rainfall is sensitive to small changes in the initial condition rather than the slowly varying boundary condition (Brankovic and Palmer, 1997; Palmer and Anderson, 1994).

Several research work also supported that the interannual variability of Indian monsoon is controlled by internal dynamics (Goswami 1998; Hazrallah and Sadourny, 1995; Stern and Miyakoda, 1995). But these studies do not provide strong evidence of the origin of this internally generated mechanism. Studies with an atmospheric system models, (Goswami, 1997) has observed that change in the intra seasonal oscillation by the annual cycle could give rise to an internal quasi-biennial oscillation in the tropical atmosphere and influence the IAV of the Indian monsoon. The interannual variability in monsoon rainfall could cause severe droughts and floods (Kripalani et al., 2003) and will adversely affect the agriculture production. This seems that IAV of Indian summer monsoon is chaotic and complex.

1.8.3.2 Intra seasonal Variability (ISV)

Rainfall over Indian region varies in shorter time scales. The fluctuation in rainfall is associated with periodicities like 3-7 days, 10-20 days and 30-60 days. Monsoon trough undulations result in 3-7 day oscillation where the monsoon low/depression or the westward moving waves are associated with 10-20 days. The higher period of oscillation in the ISV is the 30-60 day known as Madden-Julian Oscillations (Madden and Julian, 1971). The Madden-Julian Oscillation propagates eastward slowly through warm centers of Indian and Pacific Oceans and help in the organised convection. These oscillations result in active and break spells of summer monsoon over Indian region. Rainfall is sporadic for several days during the peak monsoon months of July–August, have been called breaks in the monsoon by the Indian meteorologists for several decades (Krishnamurti and Bhalme, 1976; Sikka, 1980). This Intra seasonal variability is important in the evolution and character of Indian monsoon (Lau and Waliser, 2005, Zhang, 2005; Waliser, 2006). Several research works has carried out to understand the intraseasonal variability of Indian monsoon.

During active and break periods, the circulation and precipitation pattern have high variability. The northward migration of trough causes dry spell of rainfall like condition (Blanford, 1886). The intra seasonal variability is high during the peak monsoon months (July and August). The active-weak spells in rainfall are associated with fluctuations in the intensity of the continental tropical convergence zone (Sikka and Gadgil, 1980).

1.8.4 Extreme in monsoon rainfall events

Indian summer monsoon has experienced heavy rainfall during summer monsoon season. These extreme precipitation events destroy the production of agriculture and cause human loss. Several investigations have carried out to understand the summer monsoon rainfall extremity. It is noted that during a period from 1901 to 1980, significant increase occur in the extremes in the west coast north of 12^0 N and over certain locations of eastern part of Western Ghat where southern peninsula and lower reaches of Ganges exhibit a decreasing trend. Goswami et al. (2006) have analysed the extremes and reported that the frequency and magnitude of extremes has intensified for period from 1951 – 2003 at the same time the trend of moderate events decreased. Recent observation of Indian Meteorological Department has reported that there is decreasing trend of wet days in most part of the country and also added that the flood risk also has increased.

1.8.5 Strong and weak monsoon characteristics

Prolonged active and break condition in monsoon leads to strong and weak monsoon. The active and break situations are observed mainly in July and August. During this period rainfall over Indian region is well spread.

Maximum amount of precipitation occurs in month of July and August, but the amount of precipitation varies in strong and weak monsoon years. In association with the varying precipitation large scale circulation and SST also changes. One of the important parameter that affects the strength of Indian monsoon is the low level flow at 850 hPa (Somali jet stream). The composite difference of weak minus strong condition is depicted in figure 1.17. From

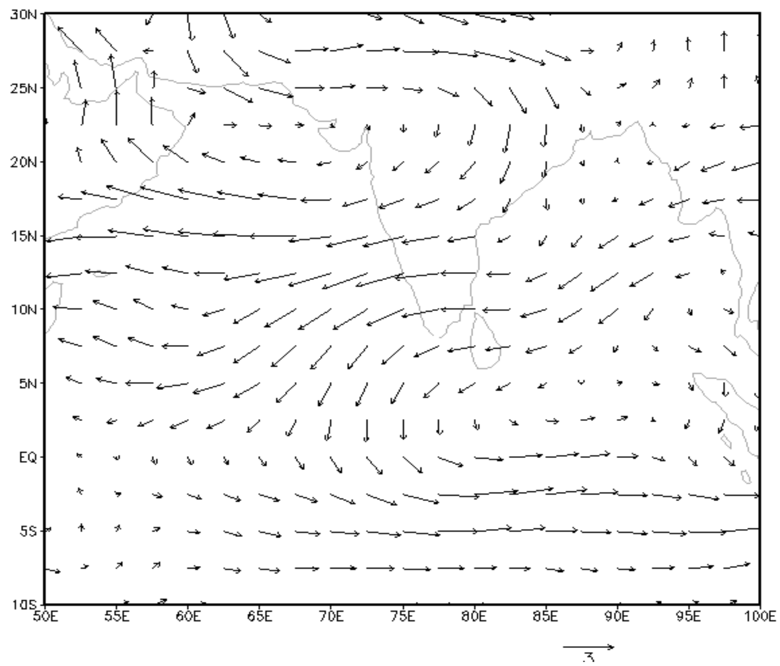

Figure 1.17. Composite of zonal wind at 850 hPa difference in July-August for weak minus strong monsoon.

the figure it is evident that during weak monsoon condition anomalous easterlies flow occur in the south Indian region. Anomalous cyclonic circulation is also observed along the northwest coast of India.

The western Pacific and tropical Indian Oceans are the warmest water in the global ocean. This warm pool is sensitive to slight variations which will affect the atmospheric convection and thereby changes the precipitation pattern existing there. The sea surface temperature during the composite difference of weak minus strong monsoon years shows slight cooling in the western Arabian Sea. During weak monsoon years, the equatorial Indian Ocean and Bay of Bengal region experiences warming (fig 1.18). The warming is intense in the southern hemisphere (area near Mascarene high). Along 100^0 E and to

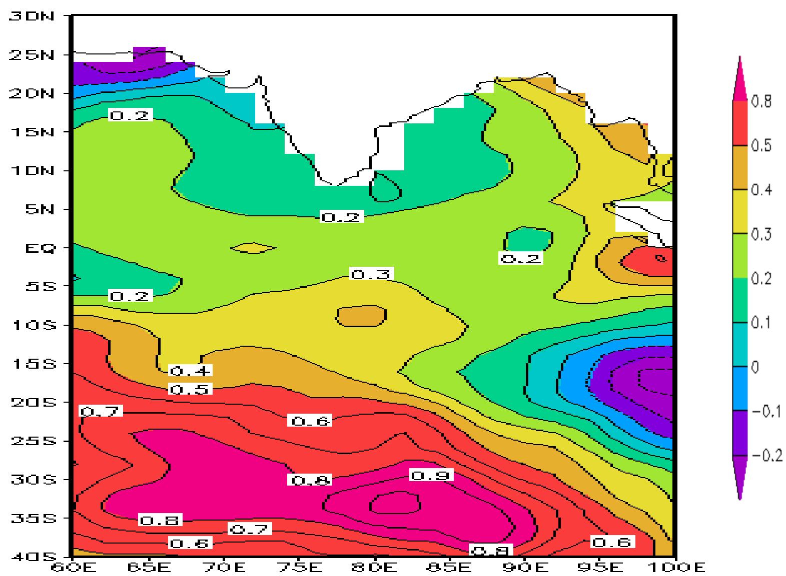

Figure 1.18. Composite of SST difference in July-August for weak minus strong monsoon.

the south of 10^0 S cooling occurs while to the north of 10^0 S warming intensifies during weak monsoon. In addition to this features, weak monsoon result in large scale subsidence over Indian domain with equatorial enhancement of rising motion (fig 1.19). Anomalous ascending motion also occurs near 27^0 N with descending motion near 30^0 N. During weak monsoon condition the moisture transport through the monsoon area also get reduced. These changes will adversely affect the precipitation amount during July-August.

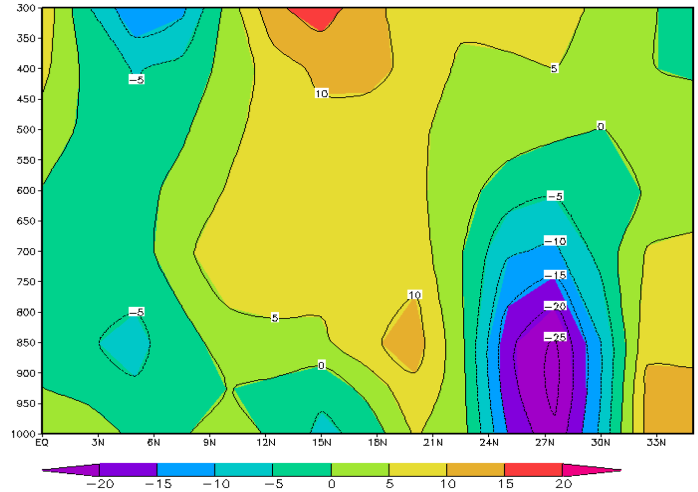

Figure 1.19. Composite of vertical velocity (10^{-3} Pas^{-1}) difference in July-August for weak minus strong

1.9 Teleconnection of Indian summer monsoon

Teleconnection is the link between weather changes of different regions of world. The global scale interaction of various parameters with El Nino and Annular modes is one of the best examples of teleconnection. Indian summer has teleconncetion with various parameters. Prediction of Indian monsoon depends upon the variability of nearby features. One among them is the El Nino, which plays major role in the success or failure of Indian summer monsoon. El Nino the tropical disturbance of Pacific region is associated with lower than normal rainfall due to the positive pressure anomalies over Indian region (Maity and Kumar, 2006; Rasmusson and Carpenter, 1983).

During El Nino, the Walker circulation gets altered with subsidence over Indian region. Warm phase of ENSO result in anomalous descending motion over Indian sub-continent while ascending motion is enhanced in near equatorial region (Krishnamurthy and Goswami, 2000). Another parameter that affect the Indian summer monsoon is the Pacific Decadal Oscillation (PDO), warm phase of this oscillation contribute to decrease in Indian summer monsoon rainfall index.

Teleconnection of Indian summer monsoon is not only confined to Pacific region but the link is also observed with Atlantic region. Atlantic Ocean weakens the monsoon circulation over India (Wang et al., 2009). Several studies have reported the existence of correlation between subtropical Atlantic sea surface temperature and Indian monsoon (Yadav, 2008; Rajeevan and Sridhar, 2008).

The snow over the different region exhibits relationship with Indian summer monsoon. The winter snow cover western Eurasia shows negative correlation with Indian summer rainfall (Bhanzai and Shukla, 1999) while the snow melt over western Himalayas is conducive for Indian monsoon rainfall (Kripalani et al., 2003). The study has further concluded that the snow melt influence the cross-equatorial flow and the heat low over northwest India.

The extra tropical oscillation known as North Atlantic Oscillation (NAO), the mass difference between mid and high latitude of northern hemisphere also play significant relation in controlling the summer precipitation over India and the relation are more effective over the sub-divisional rainfall (Kakade and Dugam, 2006). The midlaitude and ISM relation is also studied in detail, it is hypothesised that the easterly vertical shear and moist dynamic instability is related to the eastward propagation of wave train originating in Atlantic region (Ding and Wang, 2007). The pressure fields of other region also influence the ISM on intraseasonal time scales.

Fabio et al. 2003 has shown that the precipitation during June – July - August over India correlates negatively with sea level pressure field over the eastern

Mediterranean. Another influential factor to Indian summer monsoon is the Tibetan anticyclone, the temperature anomalies in the northeastern region is a useful predictor of rainfall along the monsoon trough region (Bansood et al., 2003). The southern hemispheric variation is interrelated to the summer monsoon. It is shown that the low temperature over Australia can enhance evaporation rate over the eastern tropical Indian Ocean and can induce rainfall over western India (Lee and Koh, 2012).

Relationship of Indian monsoon various parameters have altered their relationship during recent decades, in this context the search for other parameter is of great interest. Extensive work has carried out to understand the extra tropical influence on Indian summer monsoon. One such relation has accounted by the Atlantic Multidecadal Oscillation (AMO). The AMO oscillation result in the negative anomalies in Eurasia during late summer and autumn, results in early withdrawal and persistent decrease of summer monsoon rainfall. Strong North Atlantic Oscillation (NAO) also produces similar anomalies in Eurasia (Goswami et al., 2006) and thereby affects summer monsoon. At the same time it is observed that the April NAOI index is significantly related to Indian monsoon (Kakade and Dugam, 2000). During strong NAO anomalies the outgoing longwave radiation anomaly are positive and suppress the convection (Dugam, 2008) and it is concluded that the NAO index can be used as a predictor for summer monsoon. Southern hemisphere polar region has shown connection with Indian monsoon through the Antarctic sea ice content (Prabhu et al., 2009). Southeast Indian Ocean is linked with the variability of monsoon rainfall, which is a precursor of Indian summer monsoon.

1.10 Objectives of the study

The primary aim of the present doctoral thesis is to understand the variability in southern hemisphere and its relation to tropical climate. Primarily, the study focuses to find the period of climate shift in the troposphere and stratosphere of southern hemisphere. An attempt is made to understand the magnitude of temperature change in the troposphere and stratosphere of southern hemisphere after the well documented climate shift. The spatial and temporal variation of temperature is observed for specific vertical levels. Radiosonde observations are also used to understand the period and magnitude of climate shift.

Recent trends and extreme variability of meteorological parameters in southern stratosphere is another objective of the present study. The frequency of extremity is noticed to understand the magnitude of anomalous change in different stratospheric levels. Percentage dispersion of mean temperature at specific levels has been studied to understand the variability in different months. In addition to this, the El Nino teleconnection to stratosphere is observed to find the relation of troposphere perturbation to the stratospheric temperature. For the above analysis, different Nino indices are used to find the variability.

Second objective of the study is to understand the Southern Annular mode impact on the tropical circulations. The tropical circulation pattern includes Hadley and Walker circulation. Velocity Potential is used to find the area of divergence and convergence. In addition to Hadley and Walker circulations variability associated with SAM, the climatology of these circulations after the climate shift period has been analysed. Interannual variation of Hadley cell is studied to understand the year-to-year change in circulation.

The teleconnection of summer monsoon over India with the Southern extra tropical oscillation is the third objective of the present study. The study also focuses on the predictability of monsoon with Southern Annular Mode. Variability in rainfall, sea surface temperature, moisture transport and vertical velocity is investigated to find the teleconnetion of SAM to Indian summer monsoon. The influence of SAM on India summer monsoon is analysed using June and July-August indices. Further the study concentrate on the influence of North Atlantic Oscillation on the SAM-Monsoon relationship. Simultaneous effect of SAM and NAO are studied to understand the variability in various monsoon parameters like precipitation, sea surface temperature and moisture transport. By keeping the constant phase of North Atlantic Oscillation, the effect of June SAM to the Indian monsoon has been studied.

1.11 Justification of the Thesis work

The present study mainly concentrates on the change of southern hemisphere climate. Compared to northern hemisphere, the variability in the Southern hemisphere has not studied in detail. Moreover, it is interest to note the variation after the well documented climate shift. Recently it is known that the stratosphere is also a driver of climate change, so the anomalies and abrupt variation in the stratosphere is also a matter of concern. Moreover the finding of El Nino teleconnection to the stratosphere is important thing to be analysed. In addition to this, the present study focus on the influence of Southern Annular Mode in changing the large scale patterns like Hadley, Walker and Monsoon circulations.

As it is known that the Indian summer monsoon circulation arise due to differential heating of Mascarene high (located in Southern hemisphere) and heat low in the northwest Indian region. Both observational and modeling studies have reported the importance of cross equatorial flow over Indian Ocean and moisture flux from both Indian Ocean and Arabian Sea regions in Indian monsoon rainfall (Saha, 1974; Pisharoty, 1976; Washington et al., 1977; Cadet and Reverdin, 1981).

In this context it is important to analyse whether the Southern extra tropical mode can influence Indian summer monsoon, in the light of weakening relationship of monsoon and El Nino, and also with various other parameters. Further, the association of NAO to SAM- Monsoon link also boost the potential predictability of summer monsoon parameters. The SAM related work has not studied well by the scientific community, so the present study will help in better understanding the tropical variation associated with southern extra tropical mode.

Chapter 2

Data and Methodology

2.1 Data

Data and methodology used in this study is described in this chapter. This chapter consists of source of different data, period of study and techniques used for the analysis. The main reanalysis data sets used in the present study are NCEP/NCAR (Kalnay et al., 1996), ERA-40 (Uppala et al., 2005), and ERA-Interim (Simmons et al., 2006). Southern hemisphere variability has been analysed using Reanalysis and Radiosonde and other reconstructed data sets. The period of the study considered for the present work extends from 1950 - 2010. Both Interannual and intraseasonal variabilities have been studied using these data sets.

One of the main source of reanalysis data used in the present study was the National Centre for Environmental Prediction/National Centre for Atmospheric Research (NCEP/NCAR). The NCEP/NCAR provides reanalysis data of atmospheric and oceanic parameters for a period from 1948 to present. These data sets have horizontal resolution of $2.5^0 \times 2.5^0$ and consist of 17 vertical levels from 1000 hPa to 10 hPa. In the present study, zonal wind, specific humidity and vertical velocity products from this data centre have been utilized. This data sets provide reliable variables and is one of the best quality data set in reanalyses (Kalnay et al., 1996). NCEP used the climate model that was initialised with a wide variety of weather observations: ships, planes, RAOBS, station data, satellite observations etc. Using this reanalysis data, scientists have examined in detail on climate and weather statistics and also the dynamic processes.

Other sets of reanalyses data set used in the present study were the ERA-40 and ERA-Interim data from European Centre for Medium-Range Weather Forecasts (ECMWF). Both the data sets have vertical levels from 1000 hPa to 1 hPa level. Temperature data were used from the two data sources. The ERA-40 global reanalysis data provides for 45 - year period, from 1st September 1957 to 31st August 2002 and has horizontal resolution of $2.5^0 \times 2.5^0$. It was produced using a June 2001 version of the ECMWF Integrated Forecast Model (IFS Cy28r3). The spectral resolution is T159 (about 125 km) and there are 37 vertical levels from surface to 1hPa. Observations were assimilated using a 6-hourly 3D variational analysis (3D-VAR). Satellite data used include Vertical Temperature Profile Radiometer radiances starting in 1972, followed by TOVS (TIROS-N Operational Vertical Sounder), SSM/I (Special Sensor Microwave Imager), ERS (European Earth Resources Satellite) and ATOVS (Advanced TIROS Operational Vertical Sounder) data. Various data from past field experiments were used, such as the 1974 Atlantic Tropical Experiment of the Global Atmospheric Research Program (GATE), 1979, First GARP Global Experiment (FGGE), 1982 Alpine Experiment (ALPEX) and 1992-1993 TOGA-COARE.

ERA-Interim is the one of the latest global atmospheric reanalysis product provided from the European Centre for Medium-Range Weather Forecasts (ECMWF), the data sets have horizontal resolution of about $1.5^0 \times 1.5^0$. The ERA-Interim project was conducted in part to prepare for a new atmospheric reanalysis to replace ERA-40. Special emphasis is placed on various difficulties encountered in the production of ERA-40, including the representation of the hydrological cycle, the quality of the stratospheric circulation, and the consistency in time of the reanalysed fields. Temperature data for a period from 1989 – 2010 has procured from the reanalysis to understand the variability in stratosphere.

The summer monsoon rainfall over India has been studied using a high resolution ($1^0 \times 1^0$ lat/long) gridded daily rainfall, which was obtained from Indian Meteorological Department (Rajeevan et al., 2006). Using 1803 stations the data were constructed for a period from 1951–2008. Better interpolation method was used to produce this data set. The data has been compared with other global gridded rainfall datasets, and it is understood that the present rainfall analysis is better in accurate representation of spatial rainfall variation. Using this data, active and break periods were calculated during the Indian summer monsoon period. It has been observed that the result was comparable with those identified by earlier studies. This data has been extensively used for climate studies and numerical weather prediction models and is used for intraseasonal studies. The data can be obtained from National Climate Centre, Indian Meteorological Department, Pune. Radiosonde data used in the study were obtained from British Antarctic Survey (*www.antarctica.ac.uk /met/READER/upper_air/ uath.html*) and Stratospheric Processes And their Role in Climate (SPARC) (*ftp://atmossparcs unsyb.edu/pub/sparc/noaa trop/Timeser ies*).

The Sea surface temperature (SST) data used in the present study is the recent version of the Extended Reconstructed Sea Surface Temperature (ERSST. v3b) from National Climate Data Centre (NCDC)**.** The ERSST.v3b analysis is exactly as the ERSST.v3, with one difference, satellite SST data were not used in ERSST.v3b. ERSST.v3b is generated using *in situ* SST data and improved statistical methods that allow stable reconstruction using sparse data. The monthly analysis extends from January 1854 to the present, but because of sparse data in the early years, the analyzed signal is damped before 1880. After 1880, the strength of the signal is more consistent over time.

2.2 Methodology

Various techniques and statistical methods were adopted in the thesis to find the variability of the meteorological parameters used in the analysis. In the study of climate shift, Fast Fourier Transform (FFT) analysis was carried out to understand the high and low frequency variability. The FFT was used when the frequency of noise is higher than the true signal. Low pass filter analysis were carried out to eliminate high frequency undulations in the data sets. The climate shift in the southern hemisphere were identified from the Radiosonde data using regime shift indicator, program developed by Rodinov (2006). This is one of the effective methods in finding the period of shift, and it has been widely used in both atmospheric and oceanic parameters to find the climate shift. The proposed sequential program allows for early detection of a regime shift and monitoring of changes in its magnitude over time. The program can handle the incoming data regardless whether they are presented in the form of anomalies or absolute values.

The standard deviation is used to represent the deviation from mean. This statistic tool is used in understanding the stratospheric temperature dispersion during various months. The Pearson's correlation analyses were carried out to find the relation between two variables and it is tested using the table value at the desires level of significance (1% or 5%). Linear trends of temperature and its significance were noted over the global grid points. The significant differences of high and low phases of the variables were tested using *Student's t*- distribution. Moisture transport and velocity potential were calculated and verified with their climatological values and is described in the respective chapters. The methodology used in various chapters of the thesis is explained correspondingly.

Chapter 3

Impact of Climate shift over Southern Hemisphere

3.1 Introduction

Global climate has undergone many secular changes and that occur within a human life span is most alarming. These changes can occur as a result of gradual shift to a new state, a series of long term swing or a sequence of climate changes. The climate record over the last 100 years exhibits ample evidence for all these types of variations (Jones et al., 1986). Though many of these variations in climate are certainly natural, some components could be associated with increased concentrations of greenhouse gases or other anthropogenic effects. To advance our understanding of mankind's potential influence on climate, the study of various natural climate variations is also of paramount importance.

The recent climate regime shift was observed in the late 1970's and is a major scientific issue at present to be addressed. During the 1976-1977-winter season, the atmosphere –ocean climate system over the north Pacific Ocean was observed to shift its basic state abruptly (Graham, 1994). It has been suggested that the 1976-1977 shifts was caused by remote forcing from the tropical Pacific Ocean, via well known atmospheric teleconnection to the mid-latitudes (Graham, 1994). Allowing this climate shift in the tropical

Pacific, many properties of El Nino Southern Oscillation (ENSO), such as frequency, intensity and the direction of propagation, have changed (Trenberth, 1990; Wang, 1995). In addition, the ENSO period appears to have increased to 5 years since the 1980s. Clarke et al. (2000) put forward the idea that the 1976 shift influences the relationship between Indian Ocean SST and Indian summer monsoon. The long-recognized negative correlation between Indian monsoon rainfall and ENSO has weakened rapidly during recent decades.

Many previous studies have explored the possible reasons for the weakening of the ENSO–monsoon relationship (Webster and Palmer, 1997; Chang et al., 2001; Kinter et al., 2002). Kumar et al. (1999) noted the southeastward shift in the Walker circulation anomalies of ENSO after the 1976 climate shift and increased surface temperature over Eurasia in winter and spring as reasons for the weakening of the ENSO–monsoon relationship. The Indian Ocean experienced a sudden warming around 1976-1977 (Nitta and Yamada, 1989; Aoki et al., 2003). Pillai and Mohankumar (2010) showed that local Indian Ocean process gained upper hand over ENSO forcing from Pacific in biennial oscillation of Indian summer monsoon. Model study carried out by Knutson and Manabe (1998) pointed out that the late 20th century warming trend in the eastern tropical Pacific region is not likely to be solely attributable to internal (natural) climate variability. Instead, it is likely that a sustained thermal forcing, such as the increase of greenhouse gases in the atmosphere, has been at least partly responsible for the observed warming. According to a new set of climate simulations, increased emissions from fossil fuel burning set the stage for the climate shift in the 1960s, but natural variations delayed it until the 1970's (Mohankumar, 2008).

Crowley (2000) showed that temperature increase in 20th century is the effect of green house effect than natural change. Ozone variation and green house gases leads a major role in warming of the troposphere and cooling of stratosphere, which are noticeable for a period from 1979 – 1999 (Santer et al., 2003). The heat is mixed vertically and horizontally in troposphere due to physical and dynamical process in the atmosphere resulting marked seasonal and spatial variations. So any change in the temperature results drastic change in all atmospheric phenomenon. Slight variation in the observed oceanic parameter also has some contribution towards the temperature change in the troposphere levels. Moreover, the long duration temperature change in the stratosphere also affects the global circulation (Baldwin et al., 2003). An assessment of WMO reports that ozone amount decreased in between late 1970 and early 1990s (WMO, 2007). Variation in the westerly vortex is also noted in southern hemisphere after 1970s, which could change the meridional geopotential height gradient (van Loon et al., 1993; Chen and Yen, 1997).

According to recent reports from the Intergovernmental Panel on Climate Change (IPCC) the observed increase in globally averaged temperatures since the mid-20th century is due to the increase in anthropogenic greenhouse gas concentrations (IPCC, 2007). It is suggested that human activities have exerted a substantial net warming influence on climate since 1750. The anthropogenic aerosols produce a net negative radiative forcing (cooling influence) with a greater magnitude in the northern hemisphere than in the southern hemisphere.

Stine et al. (2009) estimated the terrestrial phase shift to be 1.7 days between 1954 and 2007. This shift, and the changes in amplitude, is highly anomalous when compared with the data between 1900 and 1953, associating human agency as the cause. Stine et al. (2009) also compared their observations with the results of a group of two dozen climate models used by the IPCC, and the results are dismaying. Some of these models reproduce the decrease in amplitude, first shown in 1980, but none predicts, or even reproduce, the change in phase.

Thus the temperature changes caused by the natural or human induced is a subject to be studied in detail. But the literature with both observations and modeling results does not address the Southern Hemisphere tropospheric and stratospheric temperature changes in different strata in relation to the 1976 climate shift. The present study analyses the different periodicities associated with tropospheric – stratospheric temperature and also looks at the possible impact of 1976 climate regime shift in southern hemispheric temperature using both the reanalysis and in situ data sets.

3.2 Data and Methodology

The present study used air temperature data sets at various levels in the troposphere and stratosphere (1000 hPa, 500 hPa, 100 hPa, 10hPa, 5hPa and 1hPa), both obtained from 40 year ECMWF reanalysis (1958-2001). After filtering annual cycle from the data set, anomalies are calculated for tropical belt (Equator – 10^0 S) mid latitudes (30^0 S – 40^0 S) and high latitudes (70^0 S-80^0 S) for southern hemispheres. Fast Fourier Transform (FFT) analysis is applied for all data sets to identify major frequency of oscillations. In order to study the decadal changes of global temperature the high frequency oscillations are removed using a standard low-pass filter. Both vertical and horizontal distribution of tropospheric temperature are analysed and major areas of warming and cooling are identified. The available long record of radiosonde data set at standard tropospheric level (500 hPa) is used to verify the period of shift. For this statistical approach is done to understand the time and duration of climate shift. A sequential algorithm developed for the detection of climate regime shifts from empirical data was used to understand significant shifts in the anomalies occurred (Rodinov, 2006). This method allows for the automatic detection of discontinuities in the time series. Another advantage of this method is that it can detect regime shifts toward the end of a time series, which is not the case for other automatic detection methods. Only regime shifts detected using a cutoff length of 20 years and having a probability level (p) of 0.05 were considered. A Huber parameter (Huber, 1964) of two was used to reduce the weighting of outliers that deviate by more than two standard deviations from the expected mean value of anew regime when calculating the regime shift. The Radiosonde data are obtained from *ftp//atmos.sparc.sunysb.edu/pub/sparc/noaa_trop/ TIMESERIES* and *www.Antarctica.ac .uk/met/READER/upper_air.*

3.3 Result

3.3.1 Fast Fourier Transform analysis

Fast Fourier Transform (FFT) analysis is carried out to find out the major periodicities of the tropospheric and stratospheric temperature anomalies at different latitudinal belts. The FFT pattern over the tropics, midlatitude and polar region is shown in figure 3.1.

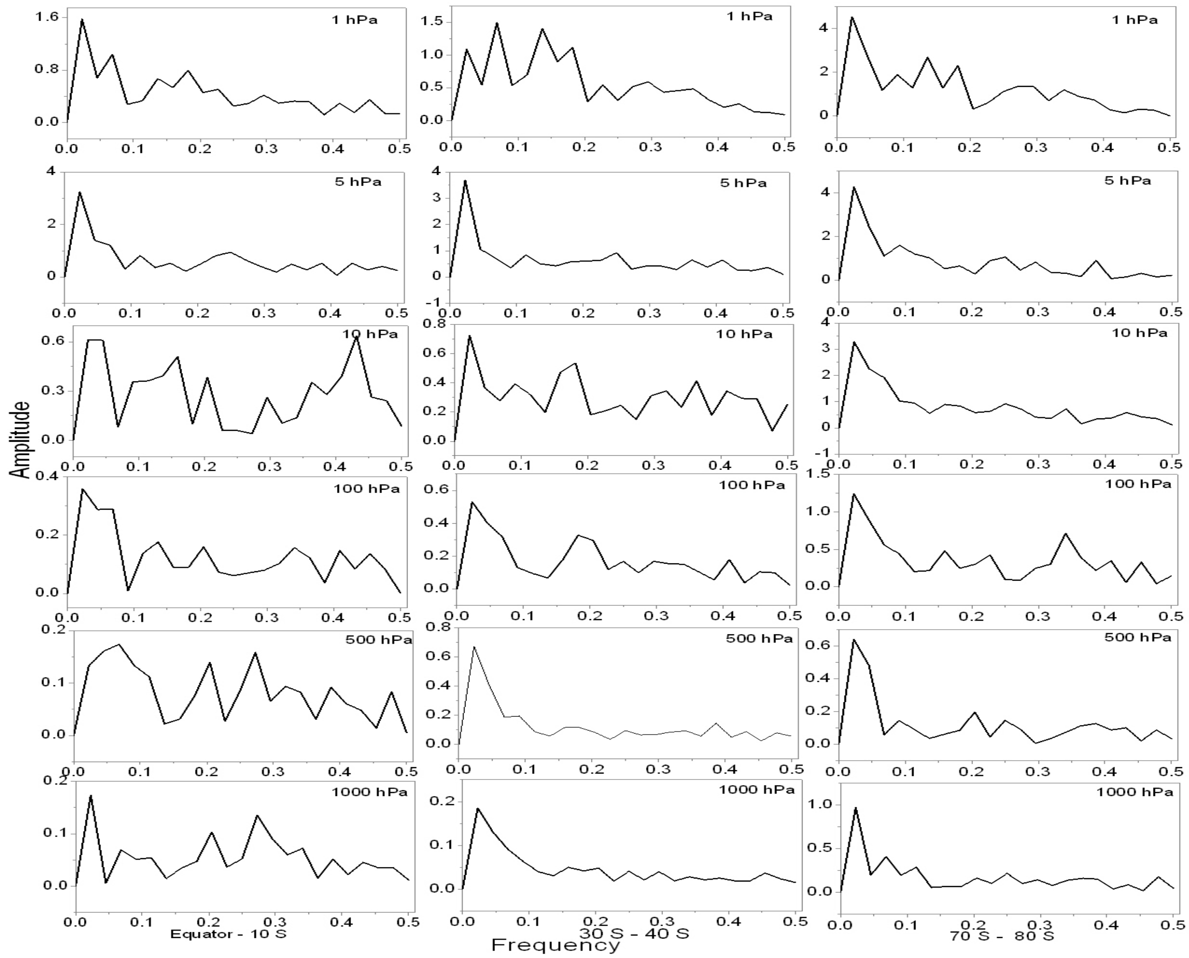

Figure 3.1. Fast Fourier Transform (FFT) of temperature anomaly (1958 - 2001) at 1000, 500, 100, 10, 5 and1 hPa level respectively for selected latitudes belt (10^0 S - Equator, 30^0 S – 40^0 S and 70^0 S – 80^0 S).

In the tropical region (0^0 - 10^0 S) the tropospheric temperature has both low and high frequency oscillations at all levels. Noticeable feature in tropics is that short periodicities (2 - 4) exhibit higher amplitude in temperature oscillations. Eleven year (solar cycle) periodicities are stronger in mid troposphere at 500 hPa. Long period oscillation is dominant in the tropopause level at 100 hPa level. The temperature anomalies near at stratopause (1hPa) level show lower frequencies. The magnitude of periodicities increased from troposphere to stratosphere. Twenty-two year cycle is observed at mid latitude in all levels of troposphere. Solar and other low frequencies oscillation are seen near the tropopause at 100 hPa height. The stratosphere is dominated by 5 –7 year periodicities with higher magnitude over the upper level (1 hPa). Southern polar belt (70^0 S - 80^0 S) is the region where shift has shown higher magnitude compared with all other region. Near the stratopause level (1 hPa), the temperature shows periodicities of 5 and 8 years.

Thus temperature has natural frequency of oscillation in all latitude zones of the southern hemisphere with low and high frequency ranges. So in order to study the interdecadal changes associated with climate shift the high frequency oscillations must be removed. Using a low pass filter, the frequencies below 15 years are removed from the entire data sets and are used for the climate shift studies.

3.3.2 Time series of low pass filtered tropospheric and stratospheric temperature

Time series of temperature from 1958 to 2001 showed an abrupt change during the 1976-1979 period. This sudden jump in the series is noted as climate shift. To understand the temperature variation before and after the climate shift period, the temperature anomaly before 1976 (1958-1975) is averaged represents PRE76 mean temperature and after 1980 (1980 - 2000) is averaged indicating POST76 mean temperature. Figure 3.2 indicates 15 year low – pass filter (solid line)) along with temperature anomaly (solid dark line), and averaged temperature POST76 – PRE76 (two point segment) at 1000 hPa, 500 hPa, 100 hPa, 10 hPa, 5 hPa and 1 hPa from 1958 – 2001 for selected latitudes.

In the southern tropics, the surface temperature increased after the post shift period. The mid troposphere showed only an abrupt jump during the shift period. The negative anomaly in the tropopause (100 hPa) region decreases after the post shift period reaching a positive maximum during 1990s. In the upper level at 10 hPa the temperature pattern has different character while comparing to that of the troposphere. Decreasing in temperature anomalies is seen during the post-shift period. Cooling of stratosphere layer is noted at 5 hPa level, the difference between the post and pre shift period seems to be greater than 4^0C. The upper stratosphere height at 1hPa level is noticed by a sudden variation in the anomaly during the shift period, thereafter the temperature decreased till late 1990s.

The surface of mid-latitude (30^0 S – 40^0 S) region exhibits an increase in temperature anomaly from the early 1960. The positive anomaly before the shift period (mid - 1970) decreased slightly for the post period. The mid-troposphere (500 hPa) exhibited the increase in temperature during the 1960s with a higher magnitude compared to the surface. The peak in temperature attaining during late 1970s and continued to be positive for the entire record. Compared to 100 hPa level with lower two tropospheric levels the variations is different. The positive temperature anomalies decreased form 1960 till late 1970s, followed by slight increase in anomalies till late 1980s. In the late 1990 onwards the temperature seems to be decreased. Gradual temperature decrease is noted for 10 hPa level for the two periods,

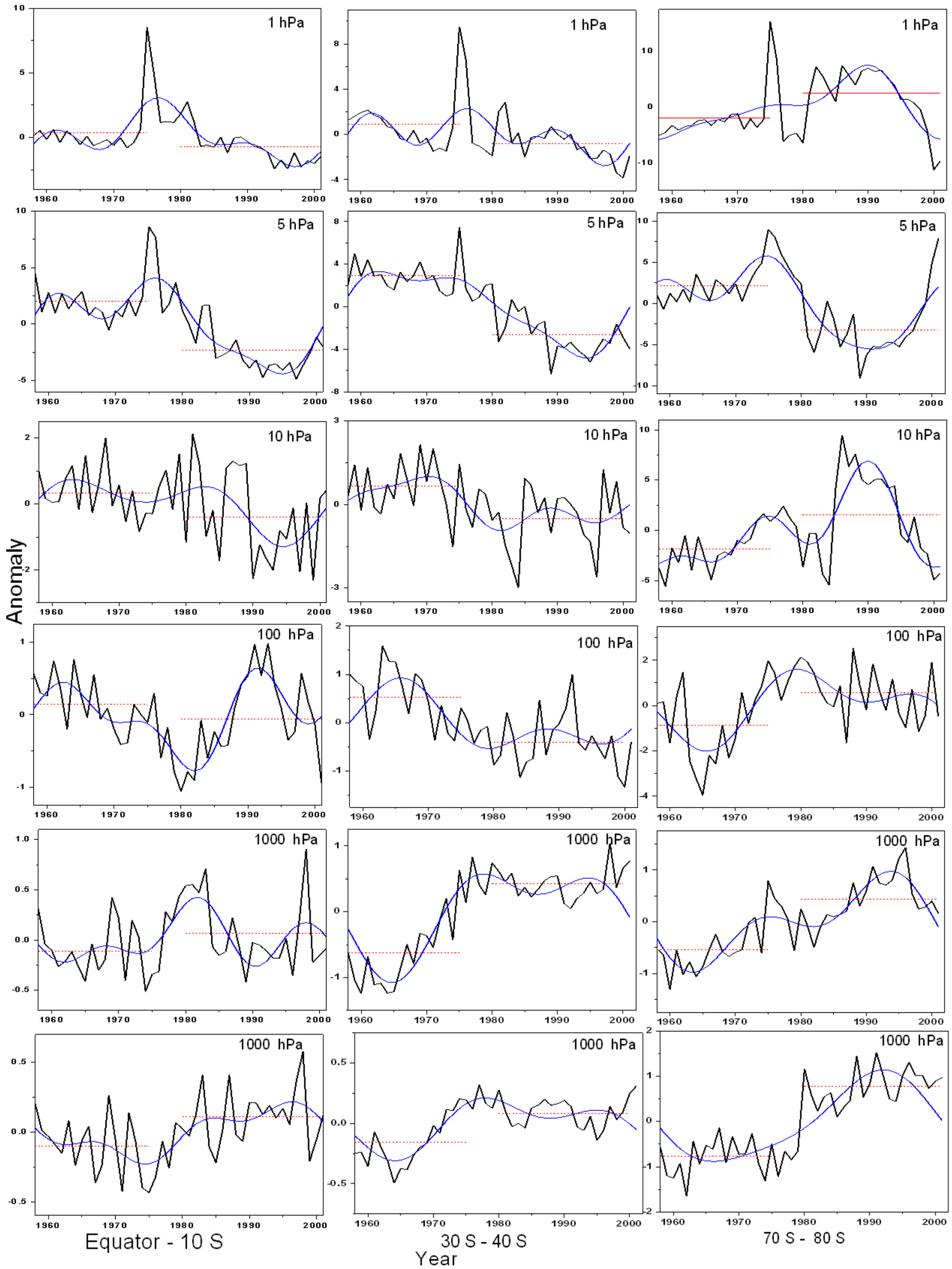

Figure 3.2. Temperature anomaly (solid dark line), 15 year low – pass filter (solid line) and averaged temperature POST76 – PRE76 (two point segment) at 1000, 500, 100, 10, 5 and1 hPa from 1958 – 2001 for selected latitudes

during late 1970s and in the late 1990s.The upper stratosphere at 5 hPa marked by negative anomaly after the post shit period. The mean difference in temperature during PRE76 and POST76 period is greater than 4^0 C at this level. In the upper stratosphere (1 hPa) the post shift period become negative but the magnitude of negative anomaly is less compared to the 5 hPa level. Southern hemisphere mid latitude exhibits large difference in mean temperature anomaly before and after the climate shift.

South polar region (70^0 S - 80^0 S) showed a strong biennial component over the surface after the shift of 1970s, with negative anomaly decreased form 1960 onwards reaching maximum positive temperature during late 1980s. Thereafter the positive anomaly decreased slightly. In the mid-troposphere region (500 hPa) the temperature anomaly showed minima in the late 1960s, and reached peak positive temperature anomaly during late 1980. In the polar lower stratosphere at 100 hPa level is marked by increase in temperature from late 1960s reaching a maximum during late 1970s. Thereafter the biennial oscillation is stronger. The middle stratospheric temperature (10 hPa) showed abrupt change during the post shift period, with an increase during the mid – 1980, and from the mid 1990s the positive anomaly decreased gradually. The upper stratosphere levels at 5 hPa and 1 hPa anomalies are in opposite phase after the POST76 period, the former exhibit a decrease in temperature and the latter showed increase in temperature till 1990s. Thereafter the temperature anomalies are in opposite phase. The difference of mean for the post and pre shift period is greater than 5^0 C.

3.3.3 Vertical structure of temperature variation

Figure 3.3 illustrates the latitudinal distribution of temperature difference for POST76 and PRE76 (for all vertical levels) from surface to 1 hPa. In polar latitudes, warming is noted over three regions, in tropospheric levels, another intense temperature variation in between 100 and 10 hPa (lower stratosphere) and the other near to 1hPa (upper stratosphere) level. Intense warming is noted near 20 hPa level (>8^0 C). Polar region is cooled at 200 hPa

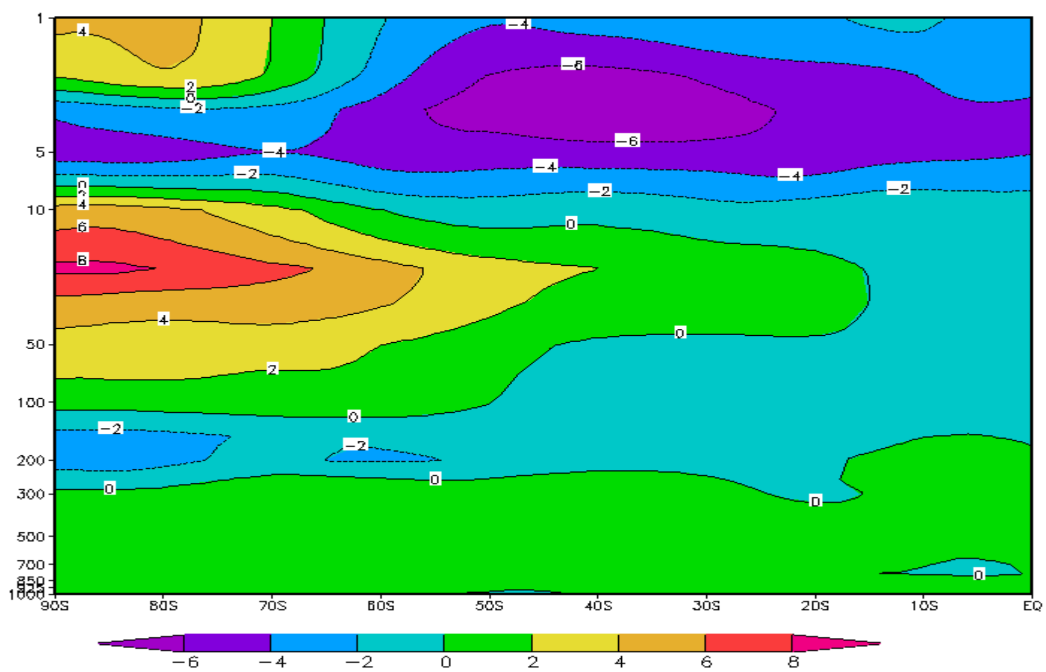

Figure 3.3. POST76 minus PRE76 latitude wise vertical tropospheric temperature profile.

and 5 hPa level, in which the latter is stronger. The upper level cooling (5 hPa) in the polar region broadened and intensified near mid – latitude region, further extending up in the stratosphere. The cooling in the stratosphere is also noted in the southern tropical region. Slight increasing in warming in all latitudinal belts is distinct for tropospheric levels from 1000 to 300 hPa. Latitudinal height cross section of the change in temperature due to climate shift indicates that the southern hemisphere has different response to climate shift.

Figure 3.4 illustrates the vertical distribution of temperature difference occurred due to climate shift in low, middle and high latitude regions in

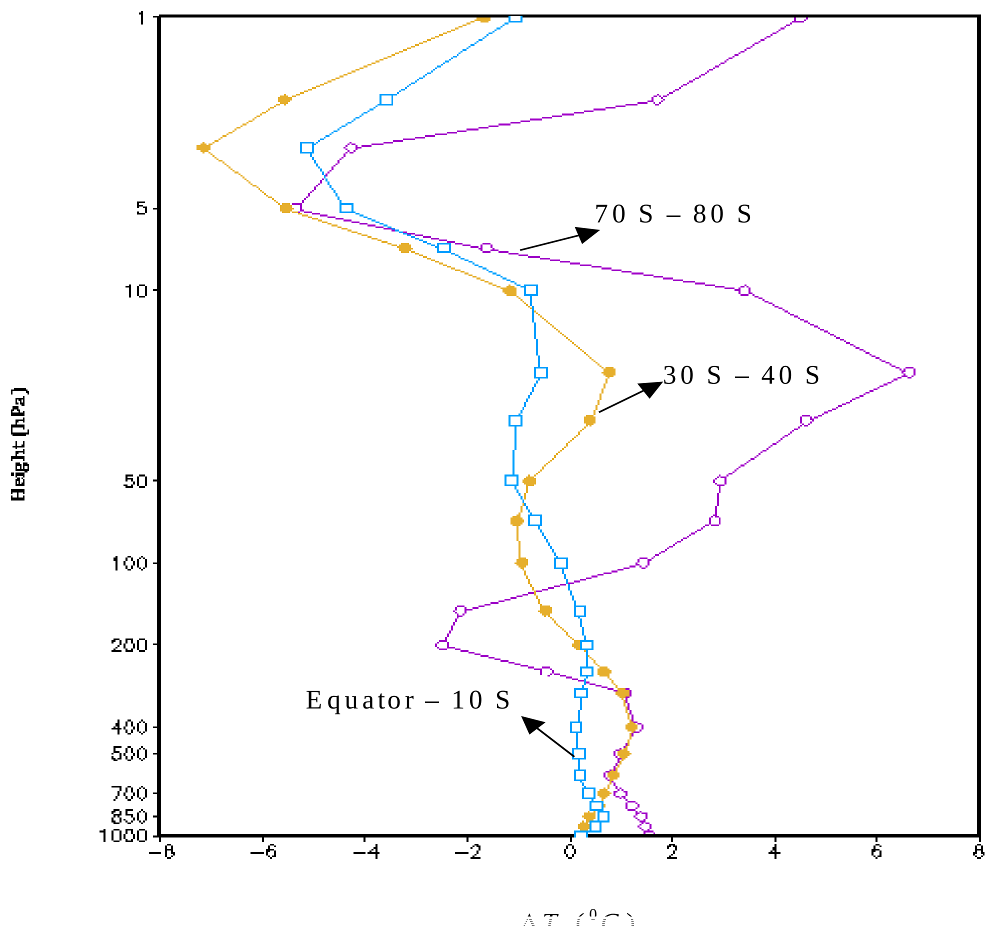

Figure 3.4. Vertical tropospheric temperature (POST76 – PRE76) for (Equator - 10^0 S, 30^0 S – 40^0 S and 70^0 S – 80^0 S) (Temperature difference - ΔT).

southern hemisphere. The low latitudes of southern hemisphere (Equator – 10^0 S) up to 10 hPa deviate around 1^0 C after the climate shift. From their onwards cooling increased reaching maximum at 5 hPa level and thereby increase the temperature at higher levels. The mid-latitudes of austral hemisphere exhibit a wavy nature of temperature change, increase at lower tropospheric levels and decreased below the tropopause region. While above the tropopause region temperature increased. In the mid – stratosphere cooling is intense around 5 hPa levels thereby decreasing at higher levels for all latitudes. Noticeable cooling is observed near 5 hPa level, where the mid-latitude variation are stronger. The polar region showed increase in temperature up to mid troposphere. Cooling of the order of -2.8^0 C is found at 200 hPa. The cooling further decreased at higher level below 10 hPa reaching a maximum increase of 7^0 C. Gradual cooling is noted after that, reaching minimum at 5 hPa level and thereafter the cooling decreased, and maximum of 5^0 C occurred at 1 hPa level. A sharp rise in temperature is observed in the lower polar stratosphere. Such a sudden increase in temperature is absent in tropics or mid latitude lower stratosphere. Upper stratosphere cooling is noted in all latitude zones in southern hemisphere, in which mid latitude region marks a cooling of 7^0 C.

3.3.4 Southern hemispheric temperature difference in the troposphere and stratosphere

Figure 3.5 describes the horizontal distribution of the temperature difference between PRE76 and POST76 period at 6 levels, viz., 1000, 500, 100, 10, 5 and 1 hPa. In the southern hemisphere, near the surface the temperature change in all longitudes are seen to be within 1.5^0 C. Cooling in the mid – latitudes (in and around Australian and south America) and warming in the polar region followed the shift. In the mid-troposphere (500 hPa) warming occurred after

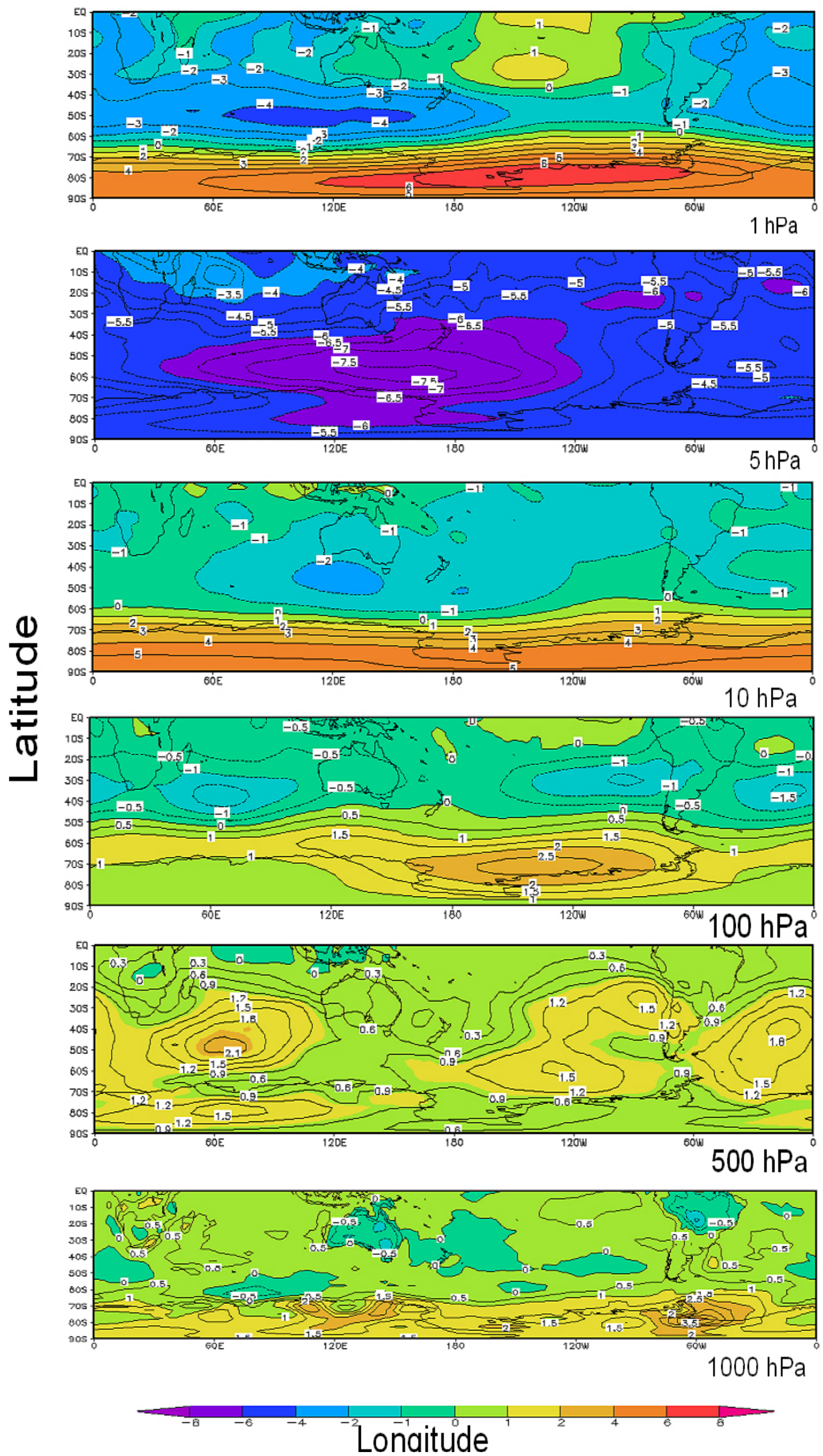

Figure 3.5. Distribution of temperature (POST76 – PRE76) for Southern hemisphere.

the shift period, mainly in midlatitudes (~2^0 C). At 100 hPa level, the temperature increased over the polar region, which peaks near 120^0 W. Mid latitude region resulted in slight cooling during the POST76 period. Stratosphere level at 10 hPa showed increase more than 4^0 C in the entire longitudinal belts of polar region, while slight cooling observed at mid – latitude region. The upper stratosphere at 5 hPa cooled in all longitudinal belts, which is intense over the eastern hemisphere (greater than 7^0 C). In the post shift period, warming is noticed in mid latitude upper stratosphere which is more intense than compared to that at 10 hPa level. The mid - latitudinal belt at 50^0 S experienced a cooling which is higher in eastern hemisphere. At the same time the temperature has increased south of 50^0 S region. In the upper levels (mainly in stratosphere), the mid – latitude shows marked decrease in temperature over the entire longitudinal belts. In Southern hemisphere the magnitude of temperature variations is high in upper stratosphere (1 hPa and 5 hPa).

3.3.5 Climate shift from in situ observations

Anomalous behavior of temperature for all levels in reanalysis data during the POST76 period is a matter of study in detail. In order to verify the climate shift time, we selected *in situ* observation of temperature at mid – troposphere (500 hPa) from following stations viz., Townswille (19.25^0 S, 146.77^0 E), Casey (66.3^0 S, 110.5^0 E), Amundsen_scott (90^0 S, 0^0 E), representing tropics, midlatitudes and polar regions in southern hemisphere, respectively. Anomalies are used to find the regime shift period, which is shown in figure 3.6. The cut-off length (l) determines the minimum length of

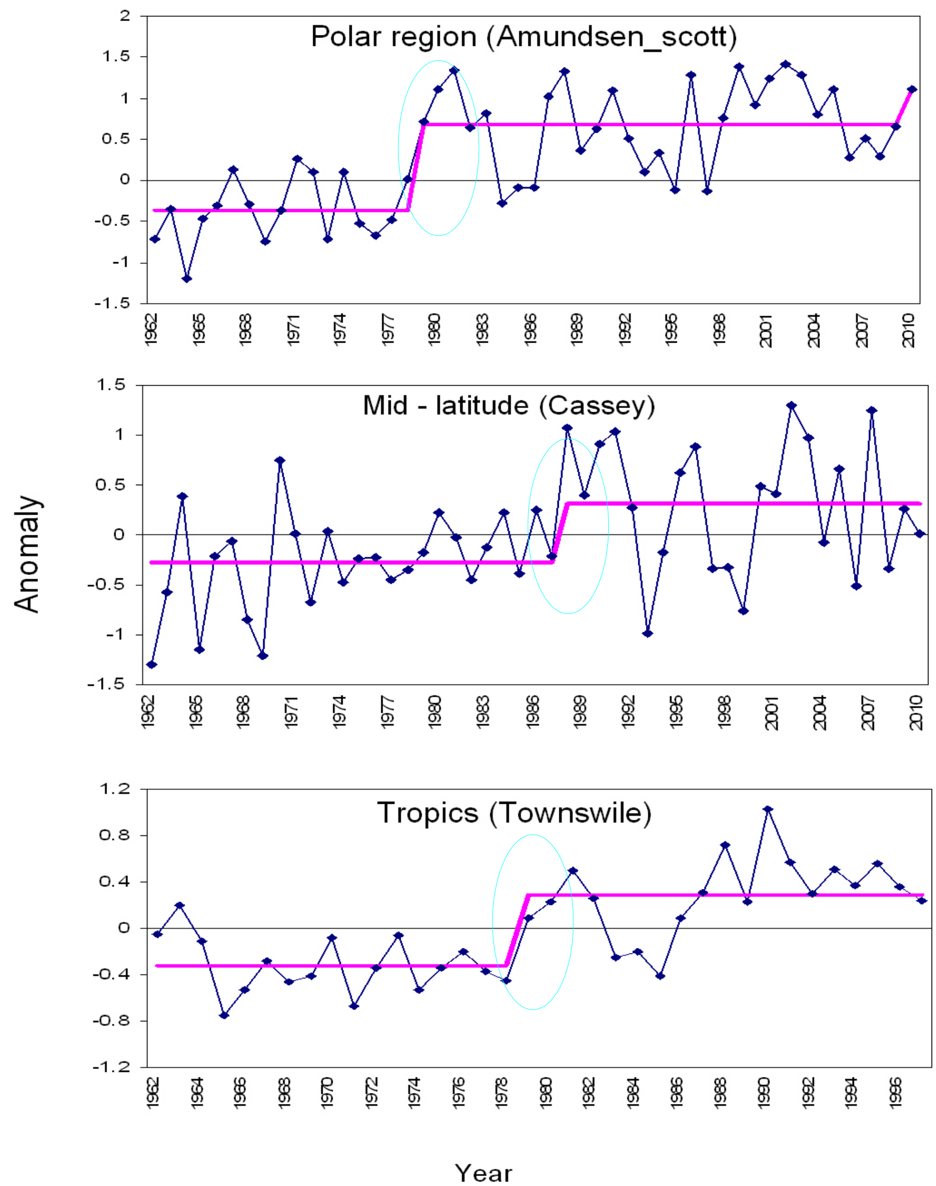

Figure 3.6. Temperature anomaly (dotted line) at 500 hPa level for Townswille (19.25^0 S, 146.77^0 E), Casey (66.3^0 S, 110.5^0 E), Amundsen_scott (90^0 S, 0^0 E). The thick pink line indicates the shift period. The bubbles indicate the time of shift.

the regimes, for which the magnitude of the shifts remains intact. The parameter l is similar to the cut-off point in low-pass filtering.

Applying the regime shifts algorithm a distinct shift is detected in the three time series of southern hemisphere representing polar middle and tropical region. The southern tropical region exhibited a climate shift in between 1978/79 periods. After the shift the tropical region consist of only three negative anomalies till 1997. For all cut of lengths, in between 10 and 25 years, the 1978/79 regime shift is detected. The mid latitude region did not pick up the late 1970s climate shift but indicates a shift during 1987/1988 with fewer negative anomalies. The polar region temperature clearly detected the late 1970s climate shift. The climate shift during the period 1978/79, is noted for all cut of length between 10 to 25 years. Only five low magnitude negative anomalies are noted till 2010. The mid latitude and polar regions are showing an increase in temperature till the late 2000s and also the tropical station till 1997.

The result from the *in situ* observations indicates a possible shift occurred at 500 hPa around late 1970s in the southern tropics and polar latitudes, and during late 1987/1988 in the mid latitudes, which altered the behaviour of temperature for a long time. This conformity indicates that the temperature changes have occurred in the atmosphere, during the second half of 20thcentury, mainly during the late 1970s.

3.4 Discussion

Reanalysis temperature pattern shows sharp change in the late-1970 (climate shift). This uncharacteristic change is quite large in the high latitude. The intensity of inhomogenities increased as we approach towards the stratosphere. Another noticeable feature is the presence of shorter period oscillations over stratosphere. Upper level stratosphere cooled over all longitudinal belts of mid latitudes with intense cooling over 5 hPa (see fig 5). At the same time, cooling is high over eastern hemisphere. In the polar region, the temperature shows an increase after the post shift period for all levels except 5 hPa. This polar warming is stronger at 1 hPa level. Vertical thermal structure shows warmer tropospheric latitudes. From equator to south polar region, temperature decreased at 5 hPa (> 6 ^{0} C). At the same time, stratosphere around 10 hPa warmed (> 8^0 C) form pole to mid latitudes in southern hemisphere (see fig 3). The polar warming is also noted at stratopause level after the abrupt change period. The different responses of temperature change over vertical levels are well evident in the time series of latitudinal belts after the post shift period. This gives further insight that troposphere and stratosphere is not warming and cooling respectively at all levels (see fig 4: 200 hPa, 5 hPa and 1 hPa), as the models (WMO, 2007) predicted. The surface as well as the lower stratosphere over polar region exhibits strong biennial oscillation in temperature after the shift period.

The shift in climate was noticed by Trenberth (1990) in the northern hemisphere during 1976-1977 period. Tsonis et al. (2007) also shown that climate shift has already occur around 1913, 1942, 1978 and will also occur in the years 2033, and 2072. Using satellite measurements Fu et al. (2006) has illustrated that the tropospheric warming trend is enhanced between 15^0 to 45^0 latitudes in both hemispheres relative to other latitudes during 1979 to 1997. Considering the change in temperature after the climate shift observational evidences indicate that the solar radiation reaching the Earth's surface decreased significantly (termed as global dimming) over northern hemispheric mid-latitude land areas from late 1950s to late 1980s (Wild et al., 2007; Ohring et al., 2008). Since then increase in solar radiation (global brightening) have been observed in many places in Europe and North America, but not in India and China. This slight variation in solar radiation may have influenced the changes in the reanalysis temperature during early 1990's, which we have also observed. Another study conducted by Soden et al*.* (2002) pointed that the water vapour feedback during the Mount Pinatubo in 1991 amplified the global cooling after the eruption. This indicates high volcanic eruptions, which contribute aerosol to the upper atmosphere and can sustain for a couple of years resulting in the changes in temperature of troposphere and stratosphere. A report of SST variation by Bob (2008) indicates that the north Atlantic region exhibits negative anomaly whereas the south Atlantic zone shows a warm anomaly after mid 1970's. Moreover the climate change during the late 1970s is noted recently in atmospheric, oceanic and even in the biological parameters is reported in Global Warming Science (2010)**.** This observational evidence suggest that climate shift of 1970s is an alarming one in a global point of view.

Model study highlights that the ozone and green house variation can play a vital role in late 20th warming (Santer et al., 2003). Another inhomogenities arouse due to expansion of radiosonde in the late 1950's and the satellite era in the late 1970's which has discussed in Greatbatch and Rong (2006) for northern hemisphere. Recent studies has pointed that the climate shift had started in 1960, but the transition occurred only during mid and late 1970s, due to the different impact of various forces (Meehl et al., 2009). The present study details the variability in the mid troposphere temperature using radiosonde data from different location in which the temperature change during the climate shift is well evident for the tropical and high latitudes of southern hemisphere. The transition period found in the tropics and polar region occurred is during 1978/1979 periods. The major shift (1978/1979) is noted for all cut of length between 10 and 25 years. While the mid-latitude does not catch the shift of late 1970s but a shift during 1987/1988 period is noted. The noticeable thing is that the 500 hPa temperature anomaly in *insitu* point towards an increase in temperature till 2010 in all the selected stations. As the reanalysis record is only upto 2001 there is some limitation in understanding the recent trend, but the radiosonde data points that the climate shift of late 1970 showed a phase change and the same phase seems to be prevailing upto 2010. Study carried out by Mohankumar et al. (2008) using reanalysis from NCEP/NCAR reanalysis also shows drastic temperature variation during mid 1970s in the stratosphere temperature. Ocean heat response, multi decadal oscillations, solar variability, volcanic eruptions and change in the chemical species may have contributed to the drastic variations during and after the climate shift.

3.5 Conclusion

Tropospheric and stratospheric temperature pattern over southern hemisphere is highly anomalous during the mid 1970's climate shift. Short period oscillations are noticed over upper stratosphere (1 hPa). During the climate shift, phase reversal of temperature is noticed in several areas with steep temperature gradient. It is interesting to see that after the documented shift in stratosphere, warming occurred at 1 hPa and 10 hPa level, while cooling occurred at 5 hPa level. This cooling is observed over entire southern hemisphere. The biennial components become stronger after the shift in polar surface and 100 hPa level. In the mid 1970 period, the inconsistent variation in temperature is high in southern hemisphere upper stratosphere.

The amplitude of difference in temperature increases with height in the stratosphere is noted after mid 1970's. Another minor phase change in temperature seems to be occurred during early 1990s. Radiosonde temperatures also indicate the climate shift (late 1970s) with steep gradient in southern tropical and polar location and the late 1980s phase change in the mid – latitude. After the climate shift, the mid troposphere (500 hPa) temperature anomalies indicates positive increase in temperature, in both reanalysis and in radiosonde. The two radiosonde observations from polar and mid latitude show an increasing trend of temperature till 2010 and in the tropics till 1997. In the late 20thcentury, climate has undergone dramatic variation in temperature that lead to unpredictable behavior of temperature that may have contributed change in other related variables.

Chapter 4

Extreme variability over Southern Hemisphere stratosphere and the possible role of ENSO

4.1 Introduction

In recent years, studies have been reporting the stratospheric variation and its influence over surface weather pattern, storms and precipitation (Shindel et al., 1999; Polvani and Waugh, 2004; Santer et al., 2005; Scaife et al., 2005; IPCC, 2007). It is noticed that the change of stratospheric condition can affect the tropospheric flow (Baldwin and Dunkerton, 2001; Thompson and Solomon, 2002; Thompson et al., 2002). Several studies have observed the fingerprints of stratospheric circulation in tropospheric weather systems (Baldwin et al., 2003; Ramaswamy et al., 2006). Detailed understanding of the stratospheric variability is needed to incorporate the changes of this radiatively sensitive layer so that it can be incorporated for future modeling studies.

After the second half of 20th century, the drastic variation of stratospheric temperature trends has been reported. World Meteorological Organization (WMO) in its recent assessment (WMO, 2007b) has shown a cooling of 0.5^0 K/decade in the middle stratosphere and more than 2^0 K/decade in the upper stratosphere. At the same time, the cooling in high latitude stated by the early assessment of WMO (2003) has been modified due to the increasing variability in polar vortex that result in decrease in cooling trend. Stratospheric temperature variations occur rapidly during seasons, which depend on the chemical and dynamical rearrangement of the atmosphere. The chemical species over the stratosphere can vary winter/spring temperature over high latitude (Randel and Wu, 1999; Zhou et al., 2001). The addition of green house gases also changes the stratospheric characteristics (Santer et al., 2003).

Sudden Stratosphere Warming (SSW), the events in the stratosphere of high latitudes, can vary the temperature to more than 30 – 40^0 C within a week, which depends on the dynamical activity of planetary waves (Sherhag, 1952; Matsuno, 1971). But the influence of the perturbants varies in both southern and northern hemisphere in which the latter is continuously disturbed. Indication of increase in the number of southern hemisphere warming events due to high wave activity during spring season is noted recently (Hu and Fu, 2009).

Southern hemisphere (SH) stratosphere variation has been continuously monitored after the discovery of Antarctic ozone hole (Farman et al., 1985). From September 2002, southern hemisphere is again in the focus of attention after a major sudden stratospheric warming, which was not reported till then. The present study mainly aims over the temperature change in austral stratospheric layers. Latitude and altitudinal temperature variations over SH have been reported during winter and spring seasons. Such changes have been observed by the spring time cooling around 50^0 S at an altitude of about 40 km (Barnett, 1974). After the 1970s climate shift, literatures have pointed out marked cooling over southern stratosphere during spring and summer (Solomon, 1999; Thompson and Solomon, 2002).

The influence of external forcing like El Nino, QBO (Quasi biennial Oscillation) and solar activity in the polar stratosphere is described by Camp and Tung 2007(a,b). El Nino Southern Oscillation (ENSO), the signal from the surface that has an impact over the global precipitation, is also known to modulate the temperature in the stratosphere through wave activity (Calvo et al., 2008). Recently it is identified that the El Nino, which warms up the Pacific sea surface temperatures has its signature over the stratosphere over Antarctic and added that the El Nino indices have varying responses (Hurwitz et al., 2011). The disturbed stratospheric polar vortex is defined during warm phase of El Nino (van Loon and Labitzke, 1987; Labitzke and van Loon, 1989). Noticeable intensification of stratosphere circulation, known as Brewer Dobson circulation is reported by model studies (Butchart et al., 2006; Li et al., 2008). This circulation has a major role in the distribution of the thermal and chemical characteristics of mid to high latitudes. Enhanced temperature variability since late 1990s compared to previous decades is a noticeable feature in the southern hemisphere (WMO, 2003).

The above studies show that there is high variability in temperature over southern stratosphere and the external force like ENSO may have impact on these layers. The purpose of the present study is to understand the characteristics of southern hemisphere stratospheric temperature**.** The study over this region is relevant due to its direct or indirect influence over surface features and recently noted influence in the climate change. Moreover, the dynamics of ozone and other chemical species depends on them are of primary importance. Main objective of this study is to find the variation of temperature in southern hemisphere latitudes over different layers, and to identify the region of higher variability. The selected layers have its own importance as it can represent the temperature variation in the region of polar vortex and ozone rich area. In order to understand the persistence of extremes, the number of days of warming/cooling during different seasons are noted. Stratospheric response over ENSO events were analysed to understand the variability in temperature, and to search whether this surface feature can play significant role in the variation of temperature above the tropopause. The study period is from 1989 – 2010, a record of more than 20 years, avoiding data taken during the pre-satellite era. Outcome of the present study provides the latest information in the southern hemisphere temperature variation in stratosphere, the time of global warming is of great concern.

4.2 Data and Methodology

The post – satellite era reanalysis data is used to study the variability in the stratospheric region. Model study suggested that the ERA-Interim (1.5×1.5) reanalysis data shows better agreement with the observations from 1990 onwards than ERA- 40 (Dhomse et al., 2011). The data also has better resolution compared to ERA- 40. So, ERA - Interim (Simmons et al., 2006) stratospheric temperature record is procured from 1989 to 2010. It is also indicated that the departure of radiosonde and ERA - Interim in stratosphere is reduced in magnitude and in vertical, the improved quality of stratospheric variable in magnitude is also reported (Dee and Upalla, 2009). The ERA - Interim records provide more than 20 years of southern hemisphere stratospheric temperature profile from 50 hPa to 1 hPa level, which is helpful for a better understanding of the climatological study. The present study focuses on the temperature trends over the southern hemisphere. Linear trends were calculated during annual and seasons from 50 hPa to 1 hPa level. The confidence level is assessed using the Student's t-test. The four seasons are represented as Autumn ~ March, April and May (MAM), Winter ~ June, July and August (JJA), Spring ~ September, October and November (SON) and Summer ~ December, January and February (DJF).

Based on the statistical significance in temperature trends, we further represented three different levels of stratosphere (1 hPa ~ 50 km (stratopause), 5 hPa ~ 40 km (upper stratosphere) and 25 hPa ~ 30km (middle stratosphere). To analyse the latitudinal variation of temperature distribution, Standard deviation (SD), which articulate the deviation from the mean. From the zonal averaged monthly climatology, SD of temperature with (n – 1) degrees of freedom is plotted for three layers. In order to retain the higher variability we have shown the deviation above 2^0 C.

$$\sigma = \sqrt{\frac{\sum_{k=1}^{n}\left(X_K - \bar{X}\right)^2}{n-1}} \qquad \text{...2.1}$$

where X is the individual value, $\bar{X}$ is the mean of the sample and σ is the standard deviation.

To understand the seasonal extremity of cooling/warming days, we plotted the number of days above/below 1 σ (σ ~ standard deviation) depending on the trends over the region. The extremity is found over three different latitudes, viz., Tropical region ~ 15^0 S - 25^0 S, Midlatitude ~ 45^0 S – 55^0 S and in Polar region ~ 75^0 S – 85^0 S. In this study, the influence of tropospheric variability over the stratospheric layers has been analysed, by understanding the temperature difference during ENSO and anti – ENSO (La Nina) years in the four seasons. Warm (El Nino) and cold (La Nina) episodes are based on a threshold of +/- 0.5^{o}C for the Oceanic Nino Indices (ONI) which is computed by taking 3 month running mean of SST anomalies in the Nino 3.4 region (5^{0} N-5^{0} S, 120^{0} W –170^{0} W) and Nino 4.0 (5^{0} N-5^{0} S, 150^{0} W-160^{0} E). The data used in the study is obtained from Climate Prediction Centre (CPC) (http/www. cpc.ncep.noaa.gov/data/indices). The significant temperature difference in the ENSO and anti ENSO (La Nina) years has been obtained by "Students" t – distribution. Two-tailed test is used to find the significance, where the hypothesis can reject the sufficiently large and small values at a time, which is not possible in the one-tailed test.

$$t = \left(\frac{\bar{x}_1 - \bar{x}_2}{S} \right) \times \sqrt{\frac{n_1 n_2}{n_1 + n_2}} \qquad \text{...2.2}$$

where $\bar{x}_1$ is the mean during El **N**ino years and $\bar{x}_2$ is the mean during La Nina years, n_1 and n_2 are the number of years. Combined standard deviation, S is computed as,

$$S = \sqrt{\frac{\sum (x_1 - \bar{x})^2 + \sum (x_2 - \bar{x})^2}{(n_1 + n_2) - 2}} \qquad \text{...2.3}$$

4.3 Results

4.3.1 Temperature

Figure 4.1 shows the altitude-latitude cross section of the seasonal and annular trends in temperature of southern hemisphere obtained for the period 1989 – 2010. Significance of trends at 95 % level are indicated in dotted lines as given in figure 4.1. Annual trend in the polar region shows distinct trends in the stratospheric layers. An increase in temperature trend near the stratopause at 1 hPa, whereas a decrease in temperature trend is seen near the upper stratosphere around 5 and 7 hPa stratums. The polar cooling near to 5 hPa extends up to tropical belts with core near 55^0 S and extends up to the stratosphere in the tropical region (> 2.5^0 C/decade). Latitudinal variation in warming is observed at 1 hPa level, which is intense over tropics. The annual trend shows statistically significant trend, but the dynamical and chemical effect for the temperature change with latitudes during the four seasons is not well explicit.

Vertical levels of temperature trend during autumn (Mar - May) shows that the latitudinal temperature deviates around 1^0 C/decade in between 50 hPa and 20 hPa. The strata inbetween 10 hPa and 2 hPa show cooling, which is intense over polar region (> 3^0 C/decade) near 7 hPa. Magnitude of polar cooling seems to decreases towards tropics. In contrast, in the upper level around 1hPa, the area of positive trends increases from pole to tropics, and become intense (> 2^0 C/decade) down the stratosphere at tropics. The winter (June – August) over SH shows that the polar temperature trend near 50 hPa is slightly increasing, while the mid latitude to tropical region temperature is slightly decreasing. The layer in between 10 hPa and 2 hPa shows decreasing trends over polar latitudes, intense especially around 7 hPa. At the same time the observed cooling extend upto tropical belt, with lesser magnitude and extend up to the stratosphere. The level at 1 hPa is marked by gradual increase in temperature from pole to tropics.

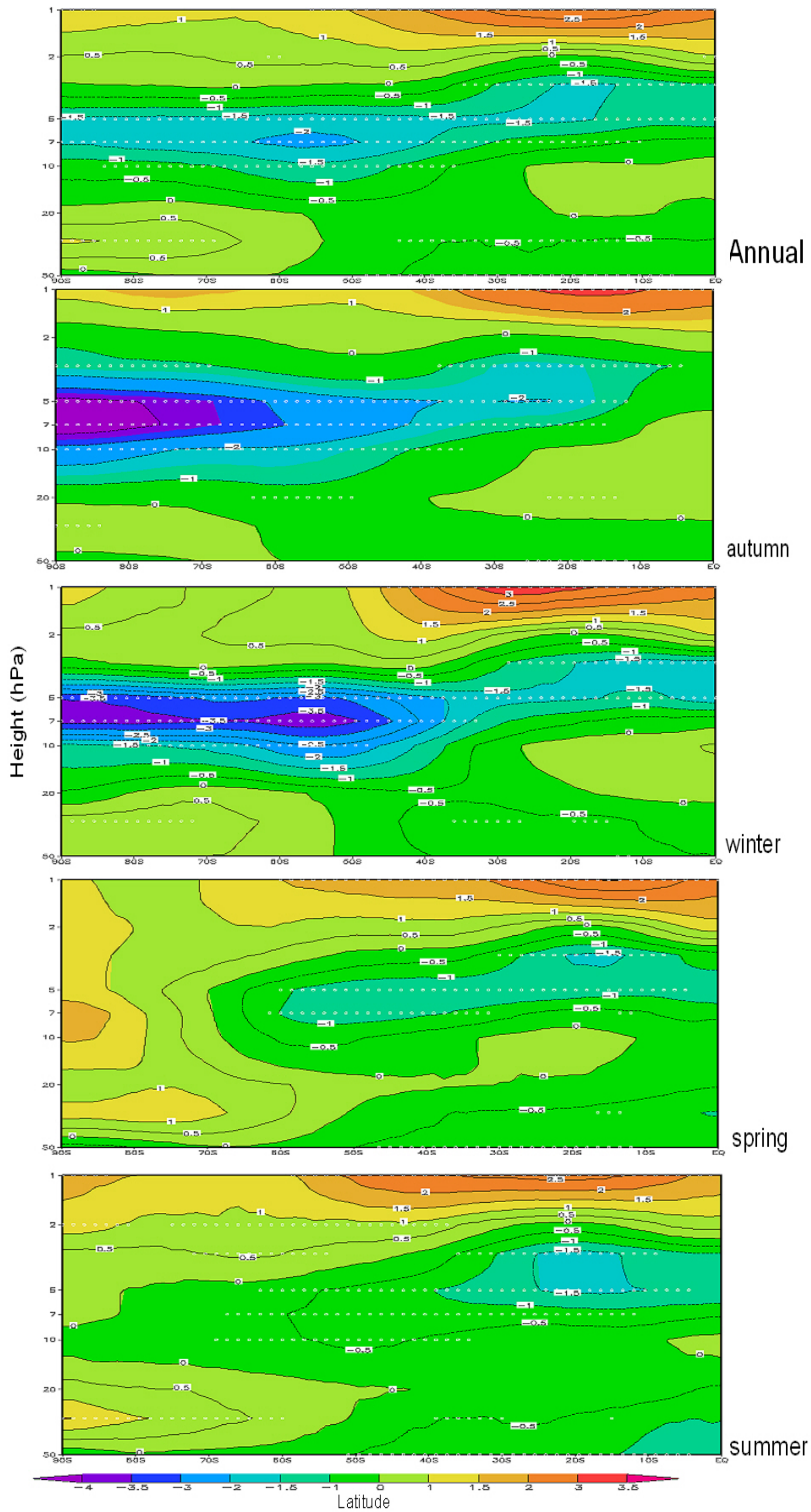

Figure 4.1. Annual and seasonal trends (^{0}C /decade) for a period from 1989 – 2010. The dotted line are significant at 95

At all vertical levels in polar region, warming is noticed during spring (Sep - Nov) with core warming is seen around 7 hPa. Such warming is absent during other seasons. The mid latitude has dissimilar trends, slight cooling around 50 hPa, with increasing magnitude of cooling around 7 hPa and an upper layer warming of about 1.5^0 C/decade. The tropical latitude belt cools up to 2 hPa and the warming increased further near 15^0 S. During summer (Dec - Feb) cooling trend is observed from 20 hPa to 2 hPa which is intense over tropics. At 1 hPa level, a gradual, increase of warming is noted from tropics to polar region.

Annual and seasonal trends show statistically significant temperature variation around the levels, 25 hPa (middle stratosphere), 5 hPa (upper stratosphere) and 1 hPa (stratopause). So in order to understand the monthly latitudinal variations of temperature over these layers Standard deviation (SD) are used, so that the area of maximum variation from the mean during each months can be noted.

Nearer to the stratopause level, the standard deviation is more to the south of 20^0 S (fig 4.2) during autumn to spring period. By the end of September, slope of maximum variation in temperature occur towards the polar region (> 8^0 C). In 5 hPa level, the deviation in mean temperature are stronger in winter (July) and spring (October) to the south of 20^0 S latitude. Around 6^0 – 8^0 C variation is noticed in the region between 50^0 and 55^0 S. The winter variations are stronger near to mid latitudes, while the variations are intense over polar belts in spring. During transition from autumn to spring at 25 hPa, the observed narrow belt of variation observed in the higher levels of the mid latitudes seems to be decreased. As the season progresses the variation broadened towards the polar region. October is observed as the month in which marked by maximum variation over polar region with a core noted around 70^0 S ($>12^0$ C). In all levels, from January to mid April the departure is less than 3^0 C.

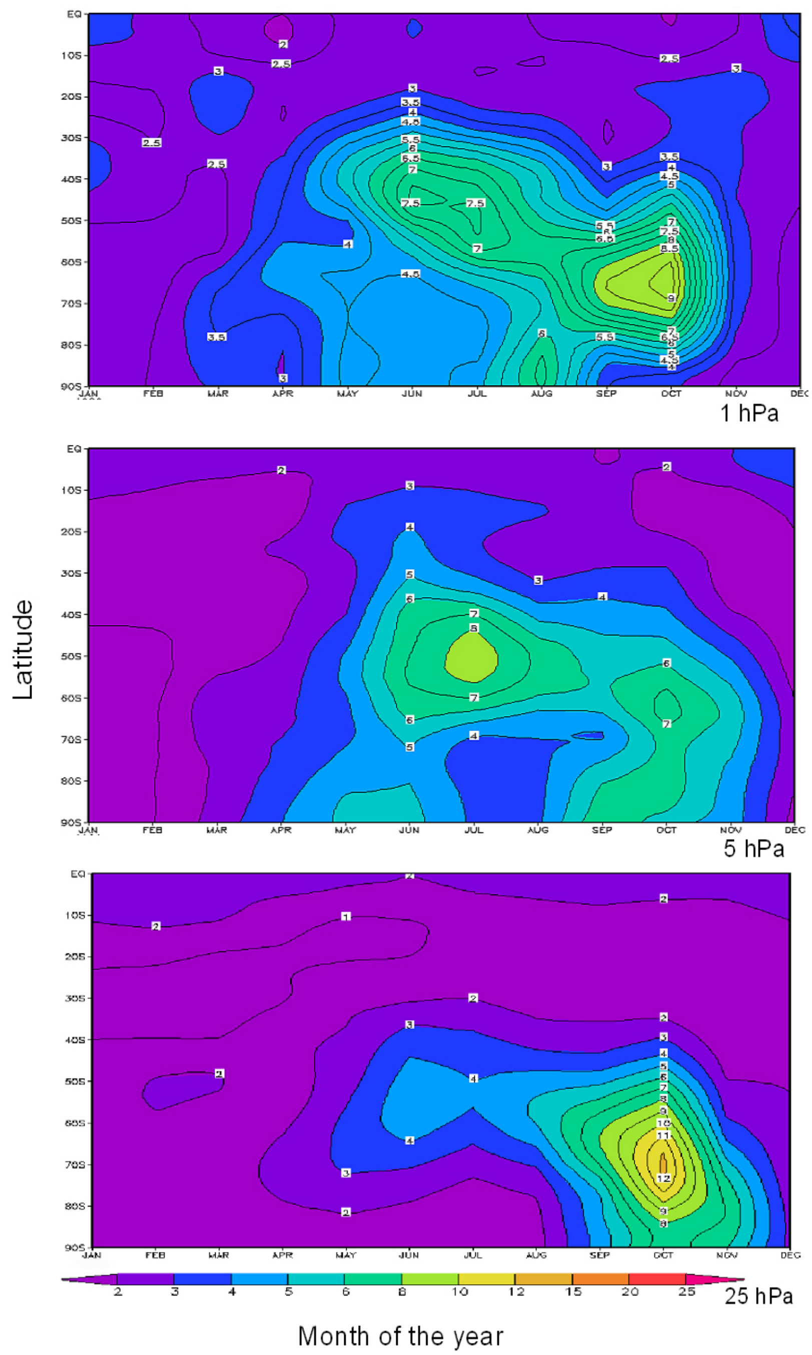

Figure 4.2. Latitudinal distribution of standard deviation in temperature for a period from 1989 - 2010.

The SD indicates that maximum variation in monthly climatological temperature occurred in between mid latitude and polar region. These variations are mainly during mid of April to the end of November. At 1 hPa and 25 hPa levels the maximum variation occur in polar belts during October. But in the 5 hPa region the maximum variation occurred at mid latitude during July.

4.3.2 Extreme temperature variability in stratosphere

Since intense temperature variations are observed in different latitudes, and months, the present analysis is extended to understand the duration of seasonal cooling/warming days. The number of days of cooling above/below 1 σ is shown in fig 4.3. The criteria chosen for the warming/cooling depends on the identified trend over the specified region, which is already given earlier (see fig 4.1).

During the austral autumn (Mar - May), increase in the number of warming days is noticed in upper stratosphere (fig 4.3a). The frequency of warming days increased in the tropical belt from 1999 (> 20 days) onwards which continued upto 2010. In mid latitudes, cooling days increased during the entire period, an abrupt increase is noticed during 1997 to 2001.

Sudden increase in the warming days over the polar region begins from 1998 onwards and remained greater than 20 days above 1 σ from 1999 onwards (except in 2003). Cooling days are observed over the entire southern latitudinal belt at 5 hPa. Tropics shows extreme cooling days from 1998 and the frequency is greater than 30 days from 2007 onwards. The interannual fluctuation in the frequency of days seems to be stronger at 25 hPa level, where the mid latitude and tropics shows warming while the polar region has cooling days. Prior to the major SSW (September 2002) increase in warming days (1 hPa) and decrease in the cooling at 5 and 25 hPa are observed over mid latitudes.

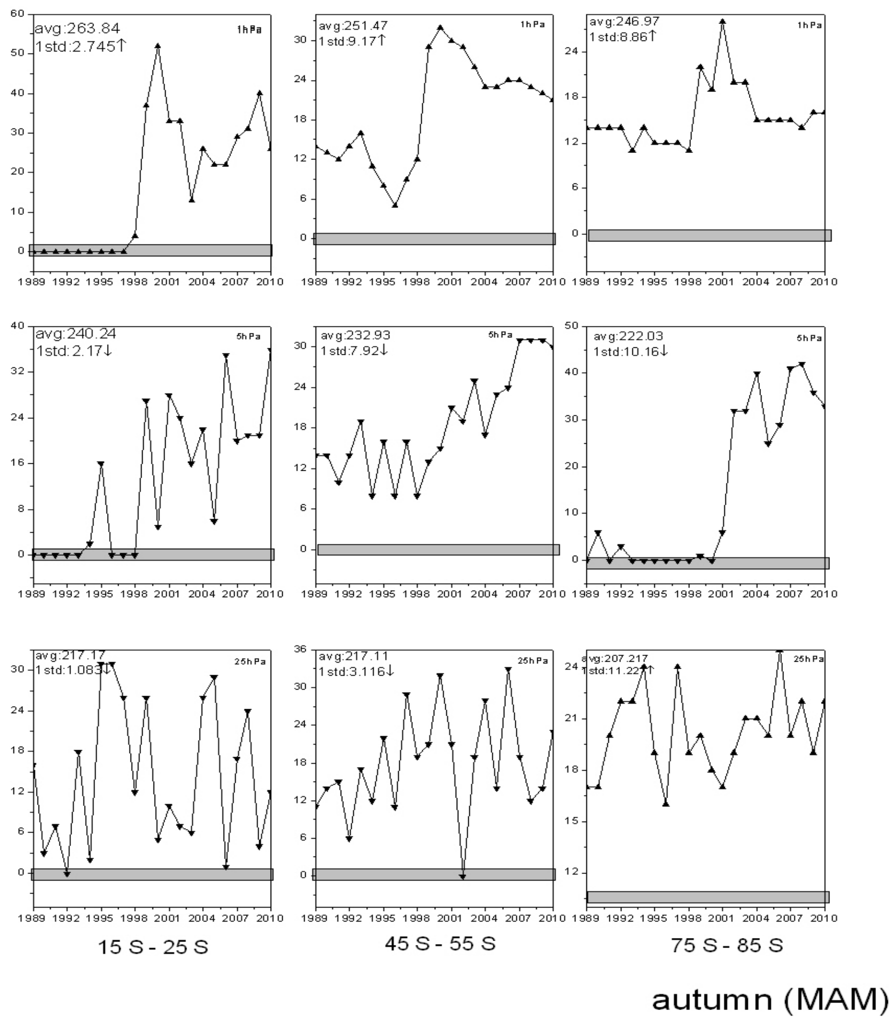

Figure 4.3a. Frequency of warming/cooling days above/below 1 σ for the autumn. The horizontal shaded region indicates the year in which the temperature is not above/below the 1 σ level during warm/cool period. Average (avg) and one standard deviation (std) is shown and the arrows represent the trend over the specific region.

During winter (JJA) the upper stratosphere in tropical belt (fig 4.3b) is marked by the increase in the frequency of warming days from 1998 onwards with a peak reached during 2000 (51 days). This warming is continuing till 2010. Mid latitude region also has similar increase in the

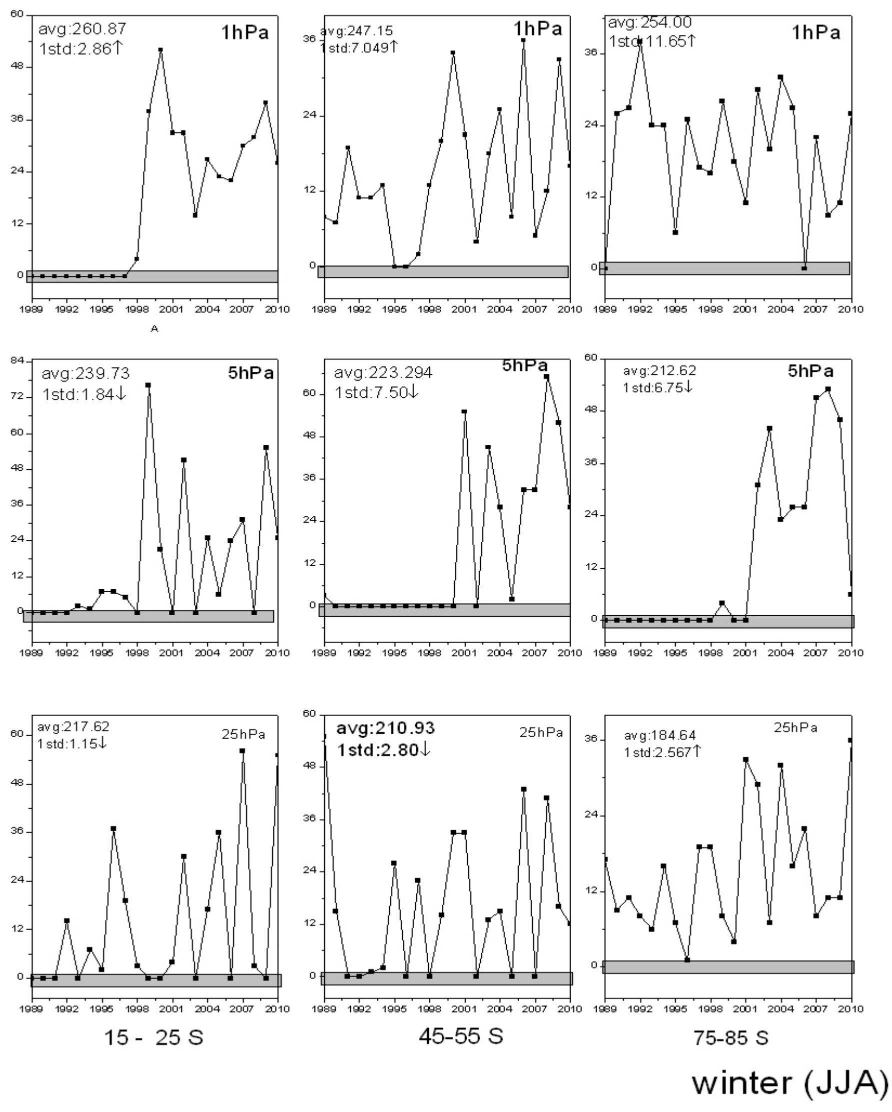

Figure 4.3b. Frequency of warming/cooling days above/below 1 σ for the winter. The horizontal shaded region indicates the year in which the temperature is not above/below the 1 σ level during warm/cool period. Average (avg) and one standard deviation (std) is shown and the arrows represent the trend over the specific region.

number of warming days where a sudden jump in the frequency is noticed from 1998 to 2001. Thereafter, the interannual fluctuations are stronger in the frequency of days. The polar belt shows a continuous pulse of warming frequency (except in 2006). Tropical latitude at 5 hPa experienced cooling days, increased from 1999 onwards (> 75 days).

At the same time cooling days enhanced in the mid latitude from 2001, except in 2002. The polar belt also experienced cooling from 2002 onwards. At 25 hPa level, strong interannual fluctuation in the extreme days are noticed for the entire period, with warming over the polar belt and cooling in the mid latitude and tropical regions. In mid-latitude, cooling extremes is not observed during 2002 (5 and 25 hPa), while the upper stratosphere warming pulses get decreased.

In the tropical and mid latitude upper stratosphere (fig 4.3c), the days of warming above 1 σ occurred from 1998 onwards. Whereas in the polar region, the frequency of sudden warming increased (warming) from early 2000 onwards during the spring. In the 5 hPa level, interannual fluctuation in the frequency of days are noticed. The polar belt experiences a warming, where 2006 has the maximum number of days of warming (> 30 days). In recent years, the mid latitude and tropical region shows increase in the frequency of cooling days. In polar region at 25 hPa level, the cooling days increased over the mid and tropical latitudes, with strong interannual fluctuation. In polar belt, the frequency of days of warming seems to decrease in the recent times but the interannual fluctuation is stronger.

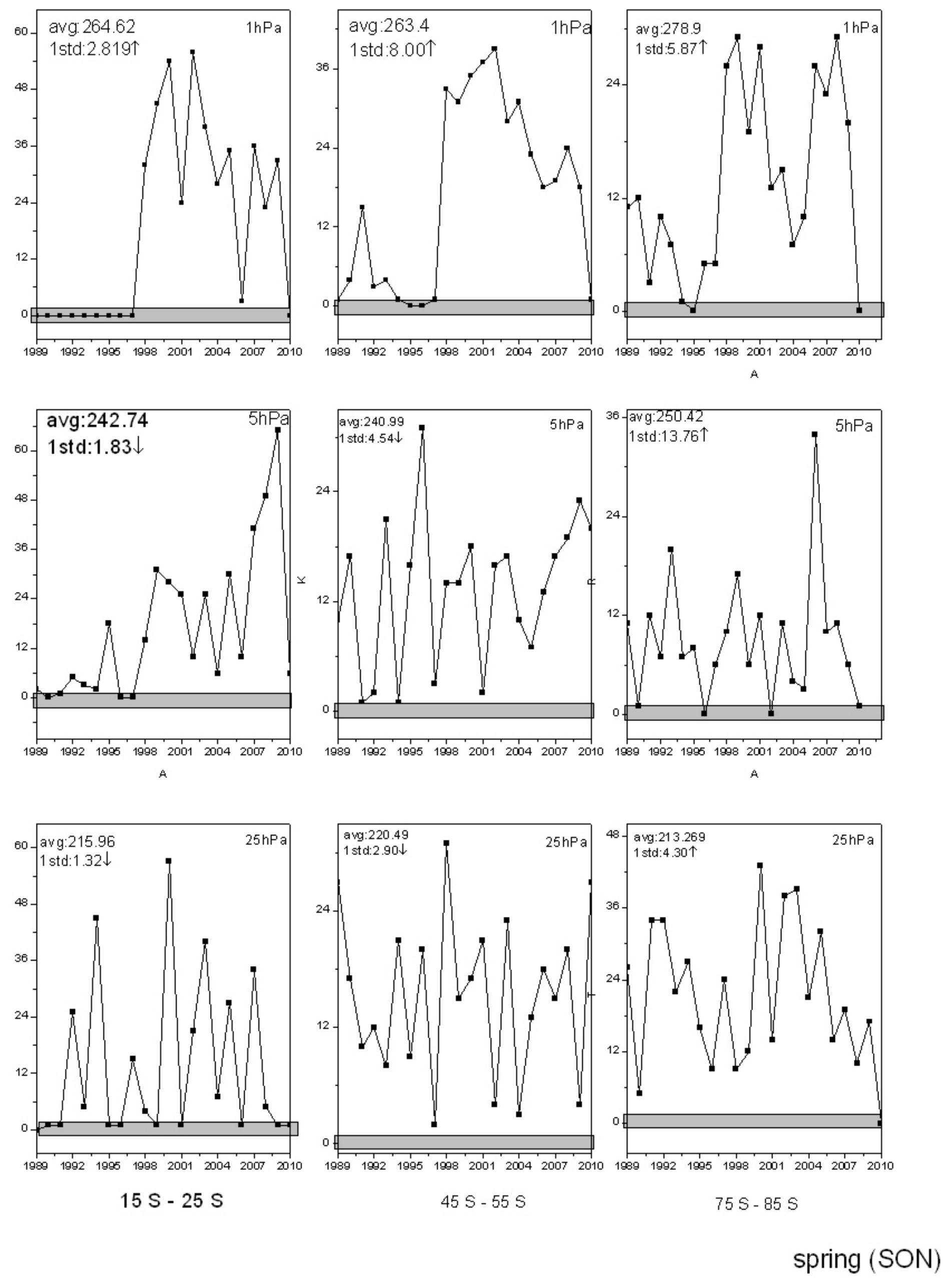

Figure 4.3c. Frequency of warming/cooling days above/below 1 σ for the spring. The horizontal shaded region indicates the year in which the temperature is not above/below the 1 σ level during warm/cool period. Average (avg) and one standard deviation (std) is shown and the arrows represents the trend over the specific region.

In summer tropical upper stratosphere (fig 4.3d) shows an increase in the frequency of warming days, which started during the austral summer of 1999. During 2001 to 2003, the frequency of warming remained above fifty days from the threshold.

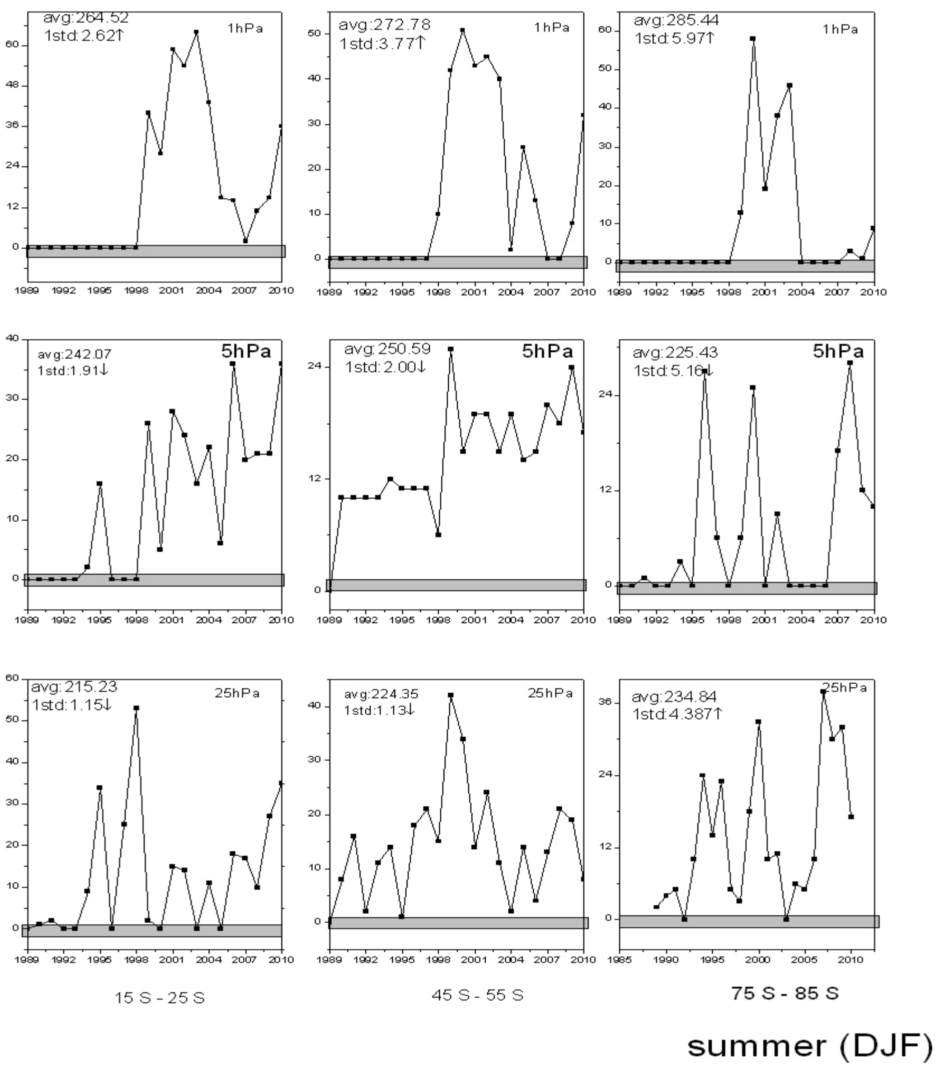

Figure 4.3d. Frequency of warming/cooling days above/below 1 σ for the spring. The horizontal shaded region indicates the year in which the temperature is not above/below the 1 σ level during warm/cool period. Average (avg) and one standard deviation (std) is shown and the arrows represent the trend over the specific region.

4.3.3 ENSO effect on the stratosphere

As intense temperature variation over the stratospheric layers is detected, the present study has focused over the recent findings of ENSO – Stratosphere relationship. Further analysis has been carried out to understand the tropical disturbance and its role in the southern stratospheric layers. For this purpose oceanic indices such as Nino3.4 and Nino 4 are used. The regions of Nino 3.4 and Nino 4 are shown in figure 4.4.

The Nino 3.4 is closer to Nino 3 region and has large variability on El Nino time scales, and this index is important for shifting the rainfall region in the far western Pacific. The Nino 4 region has a threshold temperature change that is important in producing rainfall.

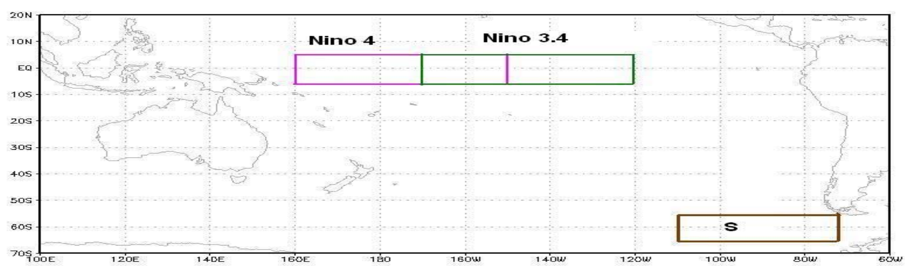

Figure 4.4. Graphical representation of Nino 3.4, Nino 4 region and S region.

The interannual variations of oceanic indices during different seasons are shown in fig 4.5. In the time series of Nino 3.4 and Nino4, it is well evident that an abrupt change in the anomaly of both indices are occurred in the four seasons from 1998 to 2002 (vertical shaded region), and thereafter weak El Nino to moderate La Nina are noted. The change in anomaly during the period is stronger in summer. Another drift in the anomaly is noted during 2007 to 2010 period.

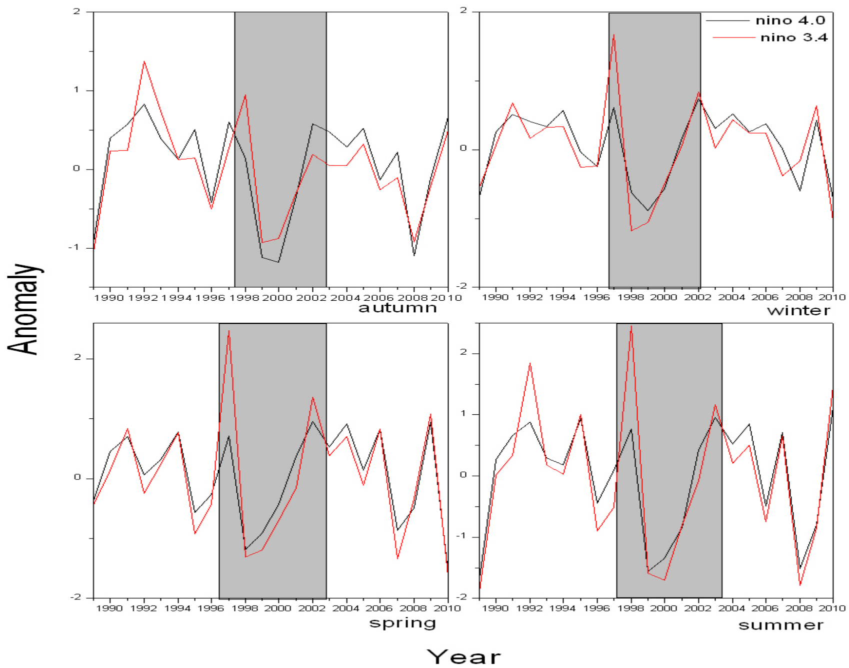

Figure 4.5. Nino 4 and Nino 3.4 indices during four seasons. The shaded box represents the period of abrupt change in anomaly

Figure 4.6 illustrates seasonal difference in temperarture during ENSO and non –ENSO years for the four seasons at the three stratosphere layers in the southern hemisphere. During the warm episodes of ENSO in autumn, the mid latitude region near to southern tip of South America, shows distinct cooling of stratospheric layers. This cooling is intense over stratopause (1 hPa). At the same time, areas of significant warming are noted in the eastern hemisphere near the polar latitudinal belt at 25 hPa level in the mid-stratosphere. In response to the warm phase of surface perturbation during winter, the stratosphere temperature change between 10^0 S and 60^0 S is significant. Considerable (statistically significant) cooling occurred in between 10^0 S and 40^0 S in the upper stratosphere. Above the southeast of south American region, significant warming is also observed. During El Nino years mid latitudinal warming is seen in the upper (5 hPa) and middle stratosphere (25 hPa) levels.

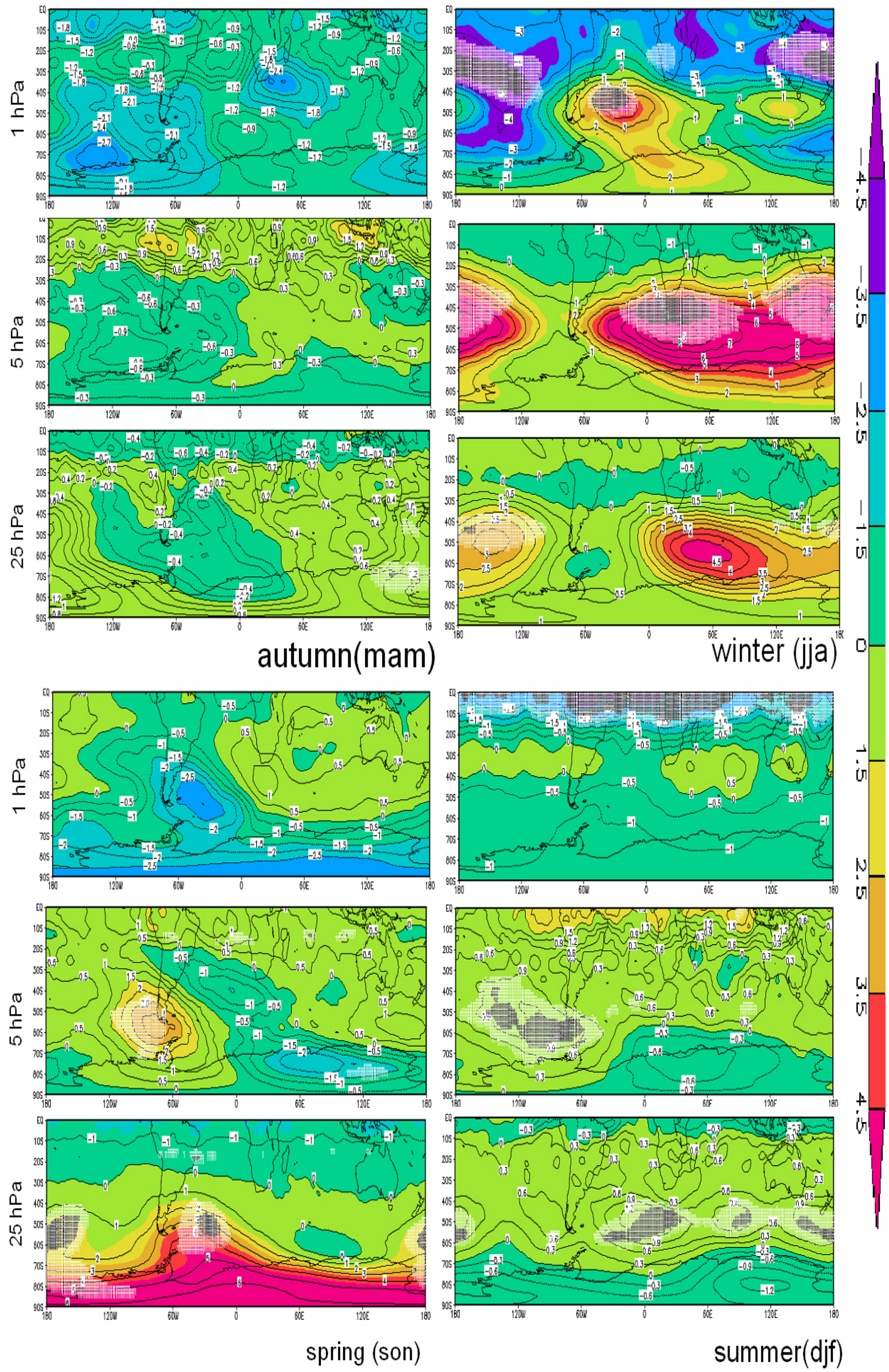

Figure 4.6. Stratospheric temperature difference during El Nino minus La Nina years using Nino 4 Index. The light dotted is significant at 95 % level and thick dotted are significant at 99.9%.

The mid-latitude upper stratosphere shows significant warming in the entire longitudinal belt. On the other hand, warming at 25 hPa is significant in the western hemisphere. Distinct variation in temperature is found during the spring in austral hemisphere. Around 50^0 S and 60^0 W weak cooling is observed in the stratopause level while significant warming is observed in the upper stratosphere. The extratropics exhibits intense warming in the mid-stratosphere while the warming is significant especially in the western hemisphere in between 45^0 S and 60^0 S. While comparing with other seasons, the summer in austral hemisphere has different response to the warm phase of El Nino. The significant cooling observed in and near equatorial region in the upper stratosphere is not observed in other seasons. In between 30^{0}S and 80^{0}S, significant warming is observed at 5 hPa and 25 hPa level**.**

For comparison, we observed the stratospheric response using the Nino 3.4 index (fig 4.7). In the stratopause level (1 hPa), the eastern hemisphere near 40^{0}S is cooling significantly during autumn. In the mean time, mid-stratosphere exhibits significant warming in the latitudinal belt near 40^0 S. In winter, near 140^0 W, upper stratosphere is cooling significantly while considerable warming is observed in the middle-stratosphere. During spring in both the 5 and 25 hPa levels the area near to 60^{0}S and 60^{0}W is warming significantly. In summer, areas of significant cooling observed near to equator at 1 hPa level. In the upper stratosphere, significant warming occurred near 60^0 W and 50^0 S, while polar region exhibits areas of significant cooling. Predominant warming also marks at 25 hPa level nearer to 55^0 S.

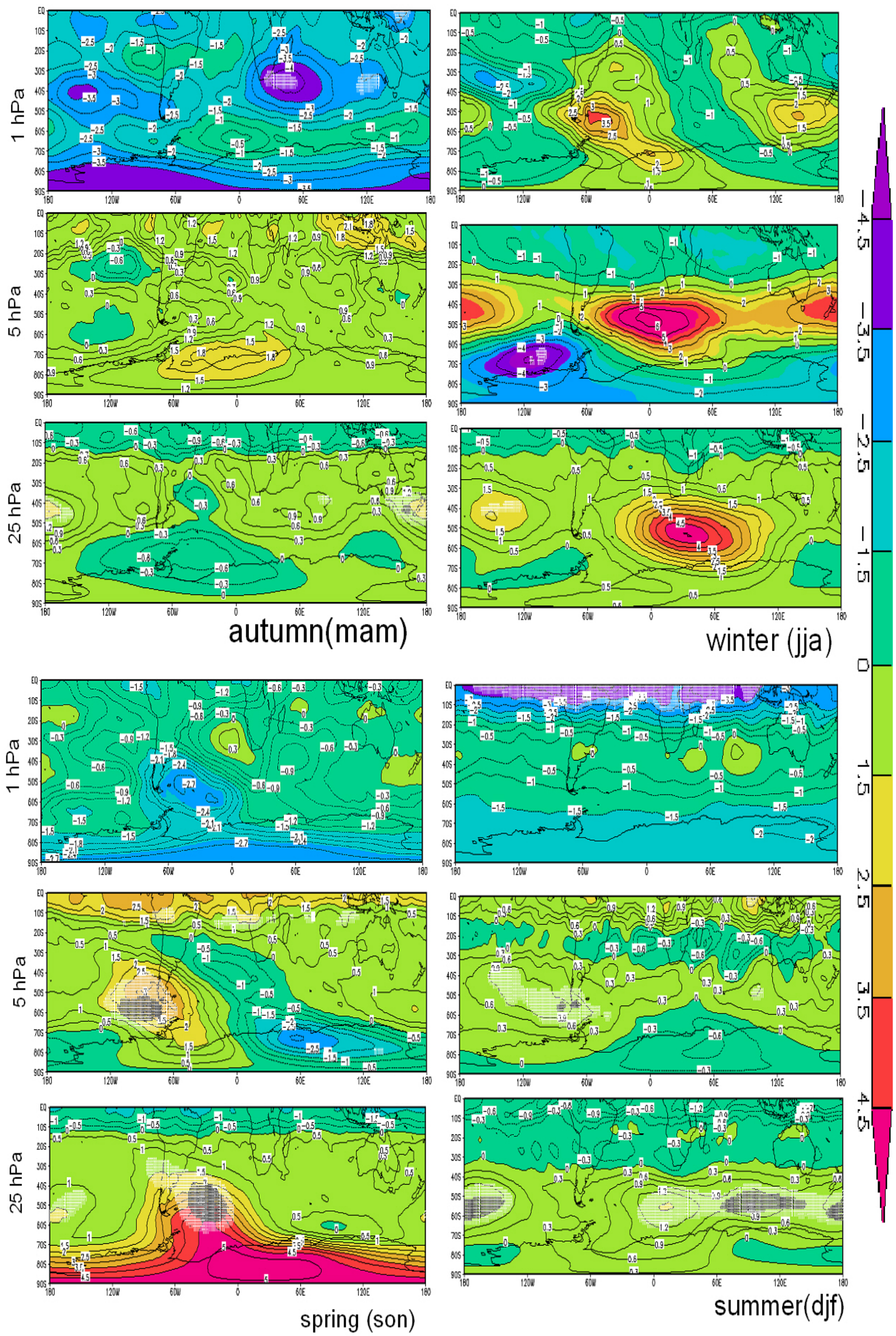

Figure 4.7. Stratospheric temperature difference during during El Nino minus La Nina years using Nino 3. 4 Index. The light dotted is significant at 95 % level, and thick dotted are significant at 99.9%.

In autumn, Nino 3.4 has core regions of significant cooling while it is not observed in the Nino 4 region. During winter, the Nino 4 region has significant relation from south of 20^0S, but the significant variations are not observed with Nino 3.4 index. Significant variation with Nino 4 index is also observed in spring with the Nino 3.4 index. The significant summer variations are same for both the indices. While comparing the two indices, most of the variations during warm and cold episodes are similar in all seasons for Nino 3.4 and Nino 4 indices. Compared to Nino 3.4 region, the spatial and seasonal variations are more significant with Nino 4 region.

The significant variation of stratosphere temperature in response to oceanic indices is interesting, so the area averaged of 'S' (55^0 S – 65^0 S, 250^0 W – 290^0 W) region (see fig 4.4) during spring (SON) has used to understand the correlation and interannual variability between them. The area is selected because this particular region at 5 hPa has shown significant warming with both the spring (SON) indices (Nino 3.4 and Nino 4). From the figure 4.8, it can be observed that the stratospheric temperature anomaly (dashed line) and Nino 4 index (thick line) show in phase relation, which is stronger after 2000 (marked by dotted line). The correlation between them is significant at 0.01 level.

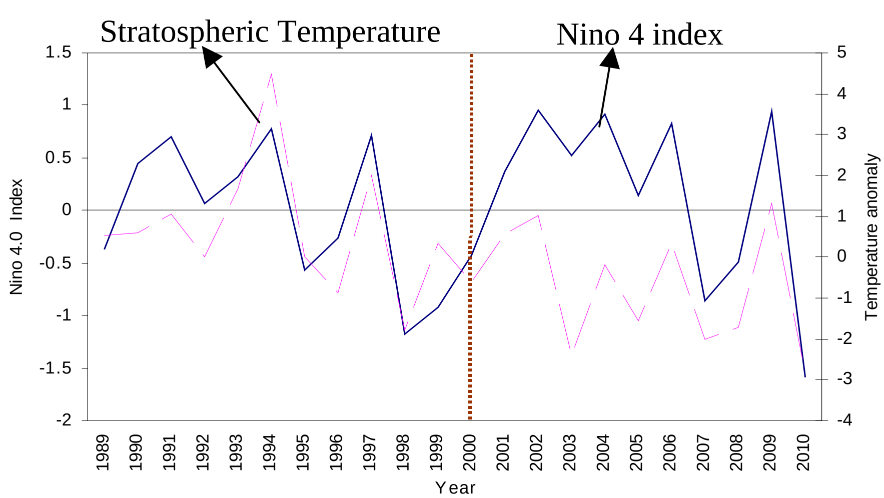

Figure 4.8. Time series of Nino 4 index and stratospheric temperature anomaly of 'S' region during spring

4.4 Discussion

Southern hemisphere stratosphere is undergoing rapid changes during the beginning of 21st century onwards which could be a result of anthropogenic influence; or as a part of decadal variation; or may be the partial influence of external forcing or the combination of all. But an unusual change over the SH stratosphere has its own importance because it has been noticed that the stratosphere could affect the climate variables of surface to a large extend (IPCC, 2007). In the present study, the annual trend shows statistically significant variation in the upper stratosphere (cooling) and warming near middle stratosphere and stratopause level. Using the ERA- 40 data, we observed that the dramatic phase reversal in the stratopause and upper stratosphere level over the extra tropical region has occurred only after the documented shift of 1976. In which the 1976 climate shift is considered as a major contribution from anthropogenic origin than the natural variations.

The upper stratosphere cooling in annual temperature can be related with difference in the properties of gas, water vapour and ozone (Shine et al., 2003; WMO, 2007), while the influence of green house gas reduces down the stratosphere (Langematz et al., 2002). During autumn and winter, we noticed areas of significant cooling near 5 hPa extending from pole to mid latitude region. At the same time the stratopause (1 hPa) in the tropical region is warming significantly. The spring variations are quite interesting in the polar latitudes with warming at all levels, that could be a result of increased extra tropical wave disturbance resulting in October warmings (Yamazaki, 1987). The increase in temperature during the spring months is also noted by assessment

(WMO, 2007). During summer, cooling in the 5 hPa level and warming at 1 hPa is observed over tropics.

These annual and seasonal trends strongly suggest that temperature variation is intense near 5 hPa and 1 hPa levels, where the former is dominant over polar region and the latter is pre dominant in tropical latitudes. In the middle stratosphere level at 25 hPa, significant warming is noticed during winter and summer. Autumn and winter seasons are showing large trends**.** The large winter trends in upper stratosphere compared to spring are well described (Hood et al., 1993; Hollandsworth et al., 1995). The standard deviation in temperature over SH is intense during winter to spring time**.** During July, high temperature variations are observed in mid latitudes of stratopause and as well as in the upper stratospheric region, at the same time the middle stratosphere over polar region are highly varying during October.

The change in polar properties can be contributed by the exchange of CO_2 and heat between atmosphere and ocean (Oinas et al., 2001; Mignone et al., 2006; Russell et al., 2006). But the upper stratospheric changes are mainly through green house gases (Jonsson et al., 2009; Stolarski et al., 2010). In the case of extremity, sudden changes in the extremity of cooling/warming days are observed around late 1990's, which continued upto 2010. These changes in the extremity are increasing during all seasons mainly at 1 and 5 hPa heights. In which, there is evidence that the mid latitude wave activity can disturb the Antarctic temperature thereby change the interannual variability (Huck et al., 2005). From 2002 onwards interannual fluctuation in the frequency of days is noticed at 25 hPa level, that point towards high biennial oscillation in recent years.

According to WMO (2007), poleward of 63^0 S, the total column ozone decreased in the late 1990's but afterwards high degree of variability are noticed and higher ozone vales in 2000, 2002 and 2004 resulting from dynamical activity. Early signs of warm perturbed atmosphere during 2002 (major stratwarm year) in autumn and winter seasons are noteworthy features in mid latitudes. The abrupt change in the extremity of cooling/warming days during 1998 to 2001 is distinct, these variations are intense at 1 hPa level. While analysing the surface indices (Nino 3.4 and Nino 4), it is observed that similar abrupt change in anomaly is found during 1998 to 2001 is distinct, and after that weak El Nino to omderate La Nina are seen. It can be interpreted that both the surface and the stratosphere show disturbed nature during 1998 to 2001. The significant effect of ENSO over the stratosphere is interesting, indicating the forcing from the surface to upper stratosphere in and near 45 ~ 50 km.

In SH stratosphere, the response over surface variability is stronger over mid latitudes, which are prominent during winter, spring and summer. Present study indicated that the standard deviation (SD) of mean temperature is larger during winter and spring**.** In SH, dynamical activity are higher during winter and spring as a result polar vortex and ozone variability are higher, in addition to that the present study shows that the surface perturbation influences significantly in temperature change over stratosphere during these seasons. Significant summer warming in the stratosphere of tropical region during the warm episodes can detail how the surface features are altering the properties of stratosphere where the extra tropical disturbance are not affected. During winter and summer seasons the extremity of warming/cooling days in the stratopause and upper stratospheric region from tropics to mid-latitudes seems to be influenced by the surface disturbance. Even though high variations in stratospheric temperature are noted between the difference of warm and cold episodes of indices, we focus only in the statistically significant area, the regions in which the influence of oceanic parameter are stronger.

The variation in stratospheric temperature during recent decades shows signals of contribution from the external forces like ENSO. Further analysing the two indices (Nino 3.4 and Nino 4) with stratospheric temperature, both are showing similar spatial and temporal relation with southern hemisphere, but the significant change is observed more with the Nino 4 Index. So it can infer that the stratospheric relations are more with the western Pacific region. Recent studies reported the variation of ENSO and its possible effect in the vertical from troposphere to upper atmosphere, due to the wave actions (Sassi et al., 2004)**.** Analysing the time series of both area averaged stratospheric temperature at 5 hPa level and Nino 4 index shows significant correlation for the entire period, the inphase relationship are quite stronger after 2000. The coherent pattern between the stratosphere temperature and ENSO is a noticeable factor.

It is noted that the radiatively induced perturbations can reach up to surface from the stratosphere during winter in high latitude and change ozone and greenhouse concentration that accounts for the change in stratosphere (Gillet and Thomson, 2003). Model study has substantiated that the stratosphere can influence the troposphere (Polvani and Kushner, 2002; Norton 2003). Solomon et al. (2010) pointed out that stratospheric water vapour change could be a factor of climate change. Even the interannual variability of ozone depends on the polar vortex temperature changes (Newmann et al., 2004; Huck et al., 2005). The important inference from the study is that the signature from troposphere can affect the stratospheric characteristics while the reverse effect is also possible. So it is important to analyse the stratospheric properties, as other features (QBO, ENSO etc) can disturb it and also the changes in the stratosphere can disturb the surface variables.

4.5 Conclusion

In the stratopause and in upper stratospheric level of southern hemisphere a rapid change in the temperature properties has occurred during the four seasons. The cooling in polar latitudes at 5 hPa level is reversed at 1 hPa, where warming occurred in tropics, during autumn, winter and summer. During spring, polar latitudes experience a warming trend, which shows increasing disturbance in the polar latitudes. From mid latitudes to polar region the latitudinal distribution of temperature is highly variable during winter to spring. Middle stratosphere polar region is disturbed mainly in October where the upper stratosphere and stratopause variations has stronger perturbations in mid–latitudes during July and October. The extremity of cooling/warming days increased from late 1998 onwards. The highly perturbed state of stratosphere temperature during 1998 to 2001 is also noticed in both the oceanic indices. The surface pertubrant has shown significant relation with southern hemisphere stratosphere temperature, mainly from tropics to mid latitudes. The relations are statistically significant during winter, spring and summer. Compared to Nino 3.4 index, Nino 4 index is showing more statistically significant spatial and temporal relationship. Significant correlation is noted for the area averaged stratospheric temperature and Nino 4 index, and the coherence pattern is found stronger after 2000.

Chapter 5

Changes in Tropical Circulation associated with Southern Extratropical Variability

5.1 Introduction

Atmospheric circulations play vital role in the distribution of mass and thereby controls the climate and weather phenomenon to a great extend. Hadley cell is one of the circulations dominating in the tropical belt and it is integral in determining the ascending and descending motions in the equator and in sub-tropics (Oort and Yienger, 1996; Trenberth et al., 2000). Through the low level rising (convergence) and sinking (divergence), it can modulate the precipitation pattern (Kumar et al., 2004) so this tropical cell is important in understanding the scarcity and abundance of rainfall in a region. Another well-known zonal circulation in the tropics is the Walker circulation, which has connection with the mean state of Pacific. This tropical east-west transport can also induce change in rainfall in many areas (Ropelewski and Halpert, 1989; Allan et al., 1996). These mass transports also have seasonal variations (Lindezen and Hou, 1988) and thereby affects the precipitation over the tropical belt (Meehl and Arblaster, 2002; Zhou et al., 2011).

But the tropical circulations like Hadley and Walker shown changes in their characteristics in recent decades and resulted in wide scale variation in climate anomalies (Kumar et al., 1999; Fu et al., 2006; Tretkoff, 2011). Intensification of winter Hadley cell and weakening of Walker circulation in the late 20th century are investigated exhaustively using the velocity potential at 200 hPa (Tanaka et al., 2004). Studies show that the changes in precipitation and circulation pattern has attributed due to anthropogenic activities (Houghton et al., 1990; Bhaskaran et al., 1995**;** Hartmann et al., 2000; Hougton et al., 2001; Marshall et al., 2003; Stott et al., 2006; IPCC, 2007; Smith et al., 2007; Solomon et al., 2007). Similar impact of climate shift is even noted in the Pacific during mid-1970s that lead to transition of cold to warmer sea surface temperature (Meehl et al., 2009). Even extratropical oscillation has varied during the late 1970 (Marshall, 2003) which is known to modulate the tropical climate anomalies (Xie and Tanimoto, 1998).

The tropical ENSO phenomenon can be modulated by the extratropics through the change in Hadley cell (Wang, 2002a; Liu and Yang, 2003). The extratropical link to Hadley circulation is described through the northern hemisphere mass fluctuation (Wang, 2002b) known as North Atlantic Oscillation (Hurrel, 1996). These studies indicate that the North Atlantic Oscillation (NAO) index modulates the Hadley and Ferrell cells by increasing its strength during its positive phase.

Another dominant mode of extratropical variability in southern hemisphere is the Southern Annular Mode (Gong and Wang, 1999). The flip-flop of mass between middle and high latitudes leads to oscillation of this mode to its positive and negative phases. These modes are known to influence the temperature, precipitation and circulation pattern (Watterson, 2000; Boer et al., 2001; Rao et al., 2003; Hall and Visbeck, 2002; Oke and England, 2004; Screen et al., 2009) and transport heat and water vapor globally (Rintoul et al., 2001). The impact of this feature is even noted in the latitudinal belt of northern hemisphere (Reason and Rouault, 2005; Gillet et al., 2006; Zhiwei et al., 2009). The coupled effect of North Annular Mode (NAM), and the Southern Annular Mode (SAM), the leading mode of variability in northern and southern hemispheres, respectively over the tropical circulation pattern has been studied extensively (Previdi and Liepert, 2007), and it is deduced that the changes in tropical circulation can be influenced by the variability of these indices. The result shows that the southern hemisphere sub-tropical dry zone has expanded towards poleward during one standard deviation increase in SAM index**.** Thomas and Wallace (2000) have also noted the poleward movement of Hadley cell due to the variability in SAM.

The main objective of the study carried out in this chapter is to investigate the effect of Southern Annular Mode during winter and summer in changing the tropical circulation pattern and the velocity potential. For this study, velocity potential at 200 hPa is used to identify the tropical circulations like Hadley and Walker. Time series analysis has been carried out to analyze the interannual variability of the tropical circulation and its association with SAM. In the context of the understanding of the tele-connection of the extra tropical influence on tropical circulation, the present study is important in providing the information of tropical circulation changes associated with southern hemispheric variability.

5.2 Data and Methodology

The data used in the present study is the monthly mean zonal and meridional wind at 200 hPa level from NCEP/NCAR (Kalnay et al., 1996), for a period from 1950 – 2010. This data has a resolution of $2.5^{0} \times 2.5^{0}$. In addition to that, the influence of southern extratropics in tropical circulation is studied using the updated Southern Annular Mode Index data (Nan and Li, 2003) from 1950 – 2010. The index above/below one standard deviation is used to represent high and low years of Southern Annular Mode.

The seasonal variation of velocity potential is analysed during the two seasons, viz., northern hemispheric (boreal) winter (DJF) and summer (JJA). To demonstrate the Hadley and Walker circulation, Tanaka et al. (2004) used 200 hPa velocity potential during different months. Present study also adopted the methodology used by Tanaka et al. (2004) to understand the Hadley and Walker circulation climatology for the post-satellite period, during summer and winter. In addition to that we observed the change in circulation due to the impact of high and low SAM of summer and winter periods. Hadley circulation is defined as the zonal mean of velocity potential, where Walker circulation is described as the annual mean of the deviation from zonal mean. Time series analysis of Hadley index is also analysed.

Horizontal wind can be divided into a non divergent and divergent part

$$\vec{v} = \vec{v}_{\psi} + \vec{v}_{\chi} = \vec{k} \times \nabla \psi + \nabla \chi$$

Where ψ is the stream function and χ is the velocity potential. The first part is non divergent and the second is the divergent part. Velocity potential is expressed in $10^{-5} m^2 s^{-1}$ which is referred as 1 unit. The Walker and Hadley circulations are thermally driven, associated with convergence and divergence. So these circulations can be better represented by the divergent/convergent component of flow. The 200 hPa velocity potential is used in this chapter so as to represent the divergent and convergent centre. The upper troposphere divergent is associated with convergence in the lower troposphere (Wang 2002 a, b), as a result the study can present the information of circulation pattern and its changes.

5.3 Results

The change in tropical circulation associated with Southern Annular Mode (SAM) is analyzed during different seasons. The changes are observed using the velocity potential at 200 hPa. Area of positive velocity potential represents rising motion, whereas negative velocity potential shows the sinking motion. The tropical circulation like Hadley and Walker circulations are also isolated.

5.3.1 Seasonal velocity potential

The seasonal velocity potential shows the area of convergence and divergence, which help in understanding the rising and sinking motions of a particular area. During solstice (winter and summer), the mean position of maximum insolation varies and thereby results in the change in the area of upward and downward motion.

5.3.1.1 Winter

Climatology of velocity potential and divergent wind during boreal summer and winter are examined for a period from 1980 – 2010. In addition to that, the change in velocity potential and divergent wind during high and low SAM phases are separately analysed (see fig 5.1 a, b and c).

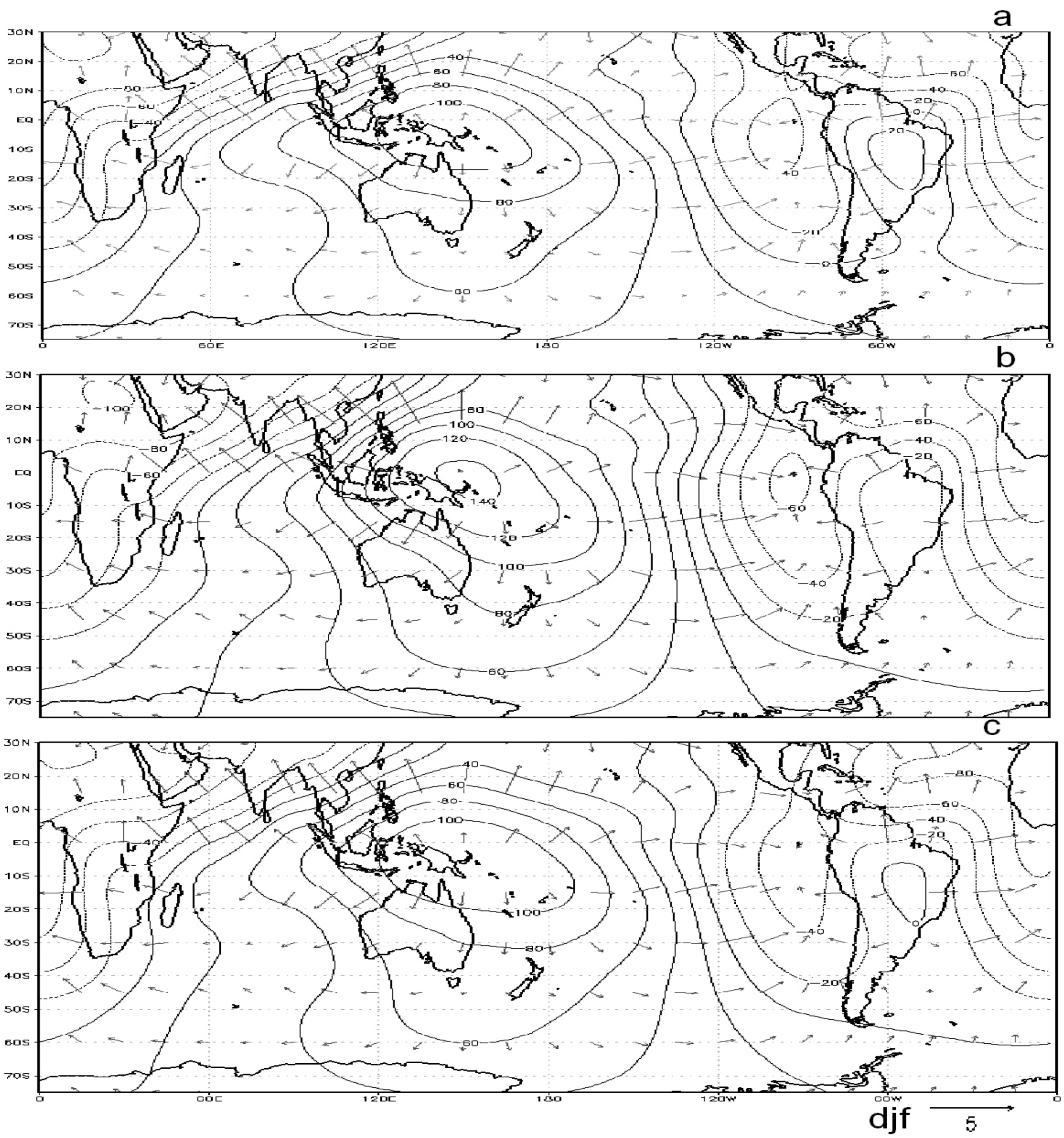

Figure 5.1. Winter mean velocity potential ($10^{-5}\, m^2 s^{-1}$) and divergent wind (m/s) at 200 hPa: a) climatology (1980-2010) b) low SAM c) high SAM.

Figure 5.1a shows the climatology of velocity potential and divergent wind during boreal winter (DJF). In winter, positive velocity potential with a maximum value of 100 units (1unit =$10^{-5}\,m^2 s^{-1}$) is located in the southwest Pacific (north of Australia). At the same time, the negative velocity extends from east Pacific to African region, with a peak value of −80 units. The divergent wind flows from the positive peak towards the north of Asia and to the east Pacific, representing Walker circulation pattern. Meridional outflow from positive area also occurs, which indicates the Hadley cell circulation.

The velocity potential and divergent wind during the negative SAM years in winter (DJF) is shown in figure 5.1b. During low SAM years, the positive area in the west Pacific region seems to be shifted to the east towards equatorial region from the climatological mean position, and is intensified to about 40 units. The negative value in the east Pacific (northwest of South America) also intensified to 20 units. The minimum is seen in the African region with a peak velocity potential value of −100 units. Strong out-flow occurs from the positive peak to east Pacific, towards north of Asian region and to the sub-tropics.

The winter velocity potential and divergent wind during high SAM years is shown in figure 5.1c. During high SAM years, the positive peak with 100 units is located in the west Pacific and minimum of −100 units is observed in the African region. Negative velocity potential value of −40 units is located in the east Pacific. The divergent flow occurs towards the negative peak of velocity potential from the positive region, which is evident from the flow towards the east Pacific and to the north Asian region and also towards the north and south latitudes.

It is apparent that, during low SAM years, both the positive and negative velocity potential value has intensified by 20 units from the climatological values. High SAM years show broadening of divergent location in the west Pacific compared to the climatological mean, at the same time the peak values remains the same.

5.3.1.2 Summer

Figure 5.2 a show the boreal summer (JJA) climatology of velocity potential at 200 hPa level. The corresponding divergent wind is also plotted. During boreal summer, the positive peak with velocity potential of 180 units is observed over western Pacific, which is 80 units greater than the boreal winter values and the peak value is shifted towards the north. The negative peak is located over South Atlantic Ocean with peak value of -120 units. Southeast Pacific shows negative velocity potential of –80 units. The divergent wind flows from maximum velocity potential value to minimum. There is upward motion in Pacific region associated with out flow towards east Pacific and to the Asian regions. Strong southward flow is observed towards the southern hemisphere representing Hadley circulation.

Velocity potential and divergent wind at 200 hPa level during negative Southern Annular Mode years is shown in figure 5.2 b. During low SAM years, the positive peak is about 200 units in the central Pacific region and the minimum of –160 units is observed in the south Atlantic region. Large positive values of velocity potential is located in the Pacific region which is shifted to the east to the climatological mean position with divergent wind spread strongly to the southern hemisphere, Asia and to the east Pacific. During low SAM period, the positive peak value is enhanced by 20 units than the climatological mean value and the negative value in the South Atlantic is enhanced by 40 units.

The 200 hPa level velocity potential and divergent wind during high SAM is shown in fig 5.2 c. In summer, the positive value of 180 units is located in

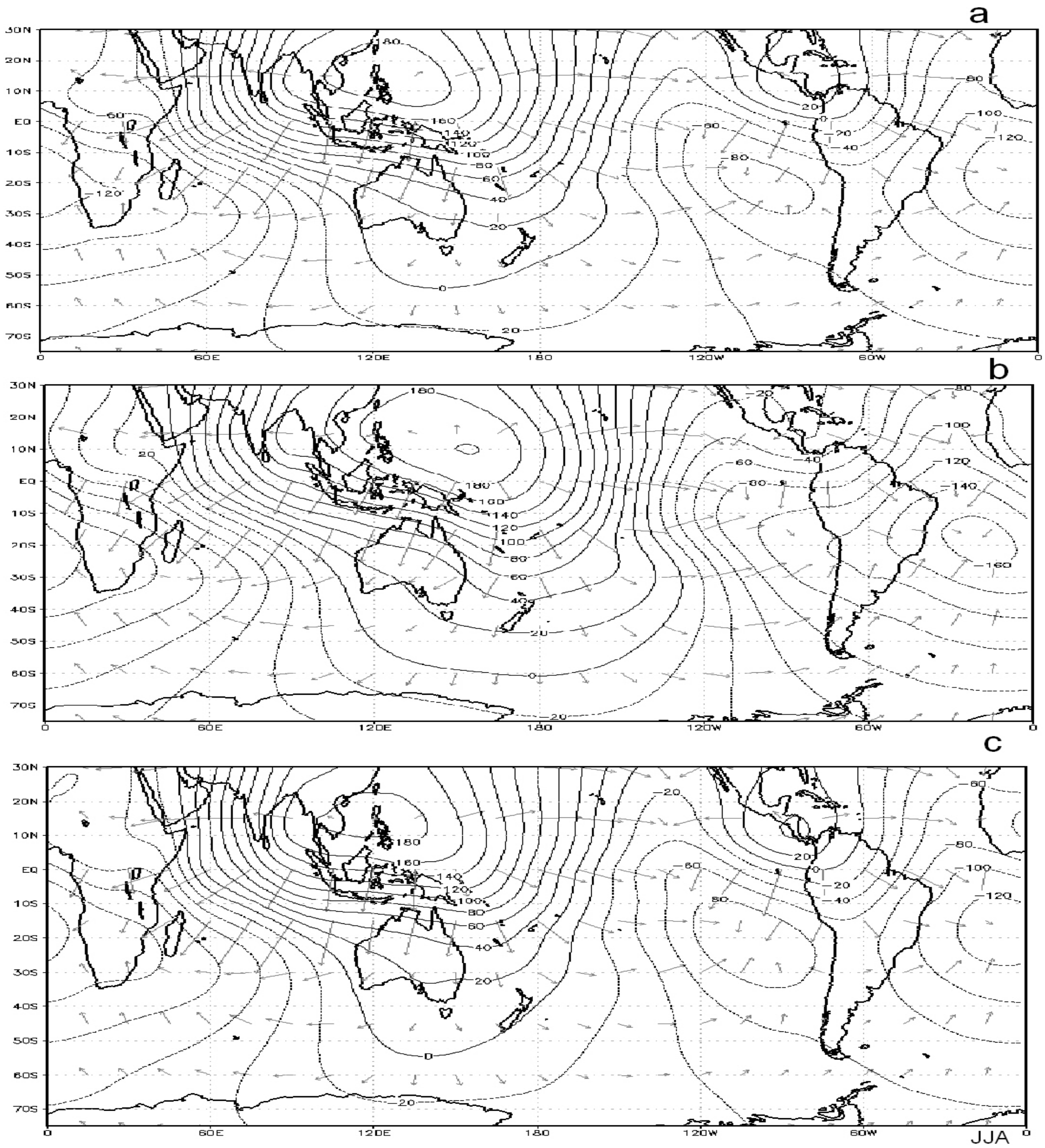

Figure 5.2. Summer mean velocity potential $(10^5\, m^2\, s^{-1})$ and divergent wind (5 m/s) at 200 hPa level: a) climatology (1980 -2010) b) low SAM C) high SAM.

west Pacific, which is slightly less than the climatology, while negative value of -120 units is observed in South Atlantic. The negative value of –80 units is observed in east Pacific. Strong divergent wind is seen from west Pacific to the southern hemisphere (Hadley circulation), towards the west and also to the east representing Walker circulation. During positive SAM years the divergent area is reduced and the convergent area in the east Pacific is enhanced. During summer, the transition of SAM phase shows anomalous variation. This feature can be observed in the western and eastern parts of Pacific.

5.3.2 Zonal mean Velocity potential (Hadley circulation)

The seasonal Hadley cell circulation is observed through the zonal mean of velocity potential during boreal winter and summer periods. During this period the maximum heating due to incoming solar radiation is shifted to northern hemisphere.

5.3.2.1 Winter

Figure 5.3 shows the zonal mean of velocity potential in winter, representing the Hadley circulation. In winter, the velocity potential in the southern hemisphere at 200 hPa level is positive with maximum value at 12.5^0 S and negative in the northern hemisphere with high at 27.5^0 N. The positive (negative) peak represents upward (downward) motion, showing the upper branch of Hadley cell with rising motion in southern hemisphere and sinking motion in the northern hemisphere. During low SAM period, the positive velocity potential value in the southern hemisphere and negative value in northern hemisphere is less than climatological mean, which shows decrease in ascending motion over southern tropics. Moreover, the peak value is observed at 10^0 S while the negative value peak at 30^0 N, showing the poleward expansion of Hadley cell in northern hemisphere while it is reduced in the southern hemisphere. The high SAM period also shows decrease of upward motion in southern tropics, and distinct downward

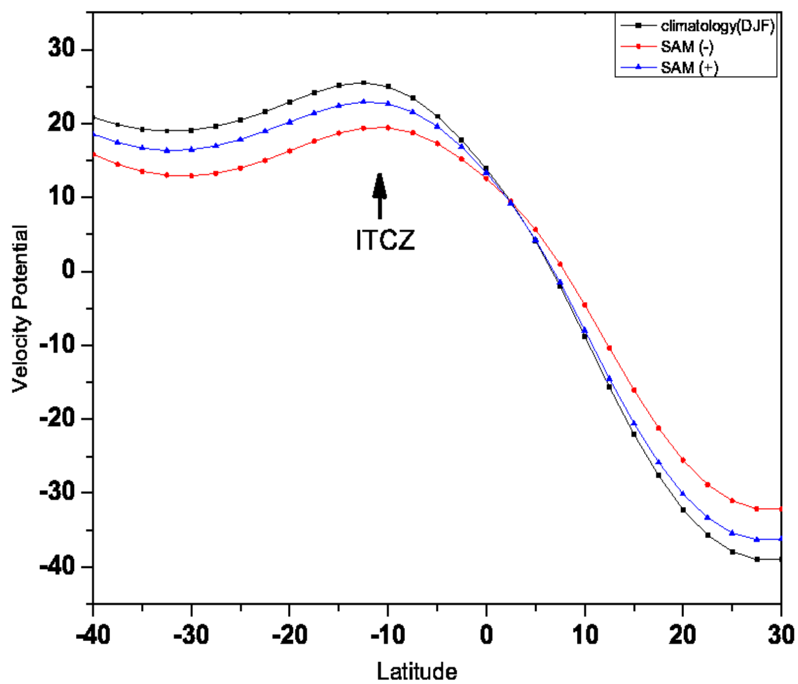

Figure 5.3. Zonal mean of the velocity potential ($10^{-5}\, m^2 s^{-1}$) in winter (DJF): a) climatology (1980-2010) (line with square) b) low SAM (line with dot) and c) high SAM (line with arrow).

movement in the northern hemisphere. But the negative vertical velocity is greater than observed during low SAM period. The peak value in northern and southern hemispheres is same. The mean ITCZ position in the southern hemisphere is at 12.5^0 S during high SAM period, while it is at 10^0 S during negative SAM years.

5.3.2.2 Summer

In summer (fig 5.4), the zonal mean velocity potential at 200 hPa is positive in northern hemisphere with peak at 15^0 N and negative in northern hemisphere with maximum at 27.5^0 S. These peaks represent the location of rising and sinking motion with the meridional gradient in equator region representing the Hadley cell circulation from the northern tropics to southern tropics. The ITCZ associated with upward branch of Hadley cell is located at north of equator at 15^0 N. During low SAM years, rising motion is slightly enhanced in the northern tropics than climatology. The negative value at southern hemisphere tropics is slightly decreased during high SAM years but the positive value in northern hemisphere is similar to the mean climatology.

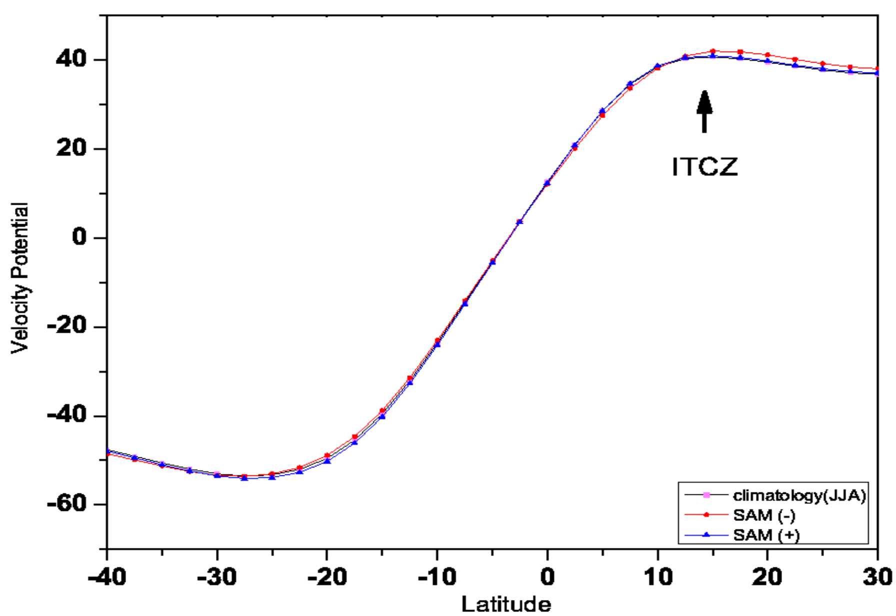

Figure 5.4. Zonal mean of the velocity potential $(10^{-5}\, m^2 s^{-1})$ in summer (JJA): a) climatology (line with square) (1980 - 2010), b) low SAM (line with dot) and c) high SAM (line with arrow).

5.3.2.2.1 Regional zonal mean velocity potential (Hadley circulation)

It can be noticed that the variability in SAM has not showing much variation from the mean climatology during summer, so the study concentrate on the Pacific region to understand whether the SAM variation can modulate the regional Hadley circulation during summer. For this, tropical domain over Pacific region (170^0 E – 80^0 W) is chosen and the zonal mean of velocity potential is examined. The zonal mean velocity over the Pacific is shown in fig 5.5. The mean climatology shows sinking motion over the southern hemisphere and rising motion over the northern hemisphere with gradient over the equatorial region representing local Hadley circulation.

But, during the low phase of SAM, the negative velocity is decreased considerably over southern hemisphere showing a peak value of –8.974 that is 25 units less than climatology mean value. At the same time, the rising motion is enhanced in the northern hemisphere. The SAM high phase shows enhancement in sinking motion over the southern hemisphere than the

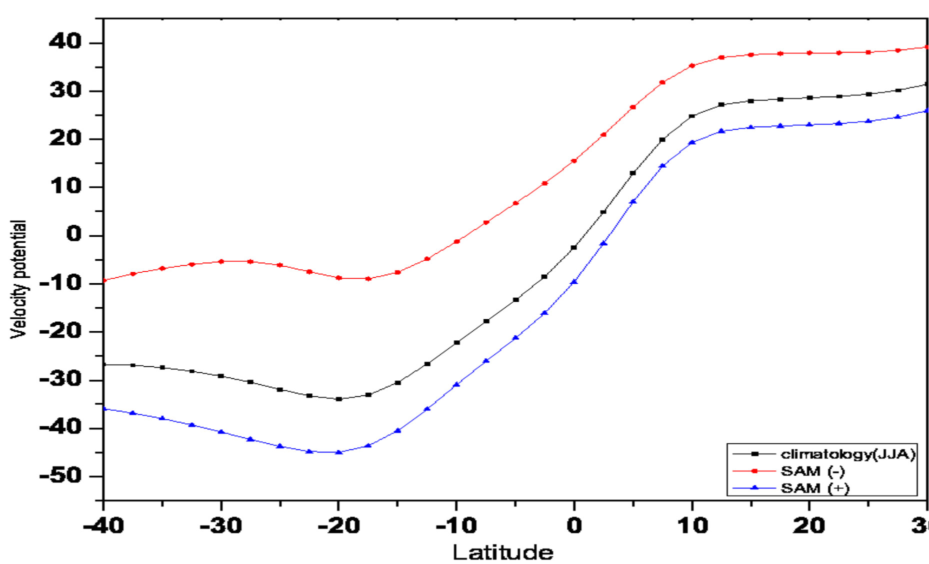

Figure 5.5. Zonal mean of the velocity potential ($10^{-5}\, m^2 s^{-1}$) in summer (JJA) in Pacific: a) climatology (1980 - 2010) (line with square), b) low SAM (line with dot) c) high SAM (line with arrow).

mean, while the upward motion in the northern hemisphere is slightly reduced. So it can infer that, even though the global zonal mean velocity potential is not showing much variation, the regional Hadley circulation is responding to the changing phase of SAM.

5.3.3 Walker circulation

Figure 5.6 a shows the annual mean of the zonal deviation field along with the divergent wind, which clearly represents the Walker circulation by removing the seasonal effect of monsoon and Hadley circulation. Peak velocity value of 100 units is seen in equatorial western Pacific and with a minimum value of −40 units in equatorial region of southeast Pacific. The divergent wind flows from the maximum velocity potential to minimum in the east Pacific. This clearly shows the pattern of Walker circulation. During low SAM years (fig 3.5 b), the positive and negative values in equatorial west and east Pacific regions are increased by 20 units than the climatological mean. This indicates that the rising and sinking motion in the west and east Pacific region is increased during negative SAM years. While high phases of SAM (fig 5.6c) shows increase in the peak velocity to about −100 units in the southeast Pacific. The positive value shows peak in the west Pacific but the area is enhanced than the climatological position and the negative area is also spread towards south.

It seems that, during high SAM period the upward and downward motions are taking place in large areas of western and eastern Pacific regions. The low SAM phase shows increase in upward and downward motion in the west and east Pacific regions, respectively. The divergent wind is also observed to the southern hemispheric region representing local meridional circulation.

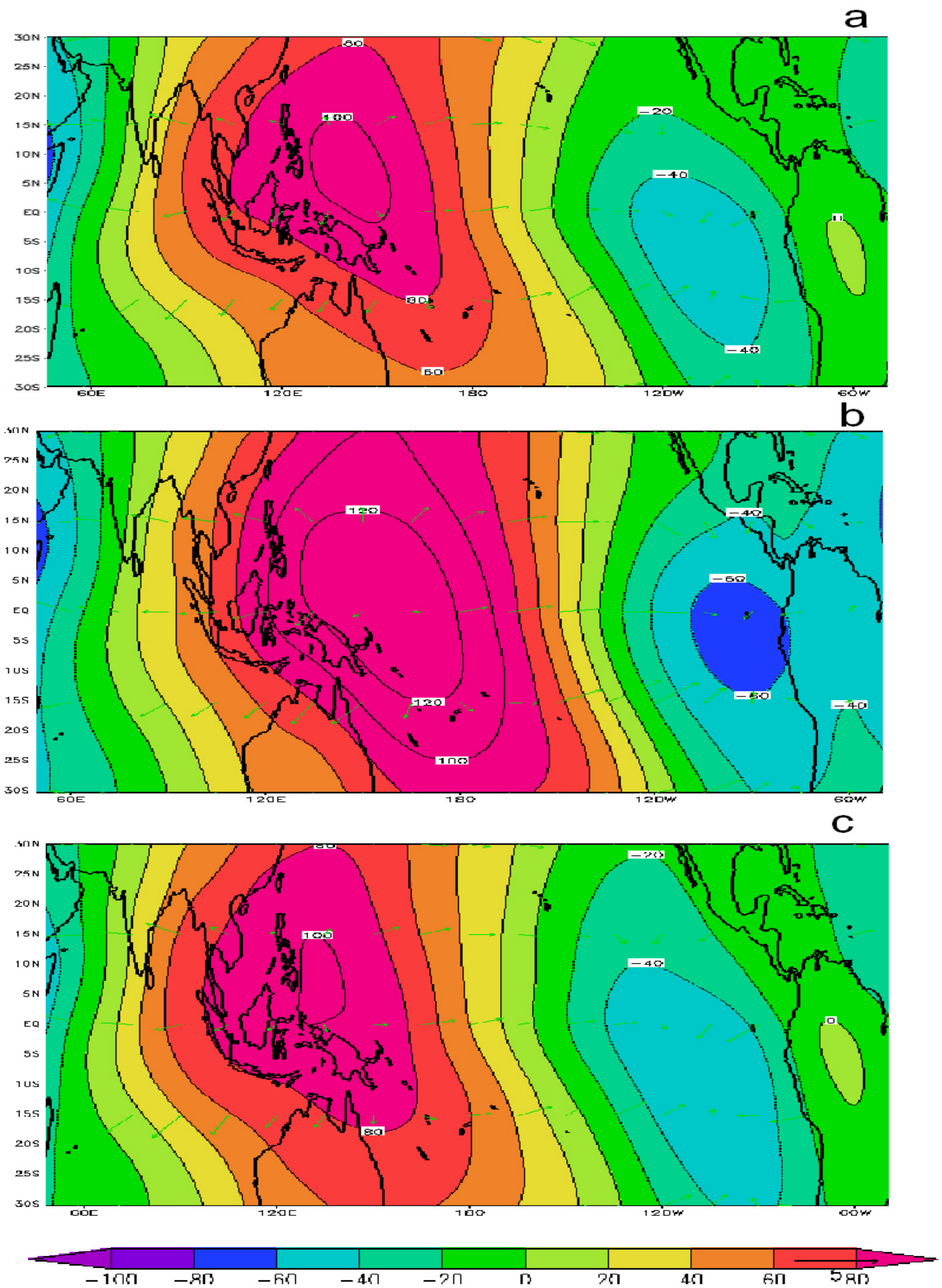

Figure 5.6. Annual mean of the deviation from zonal mean of velocity potential (Walker circulation) ($10^{-5}\,m^2 s^{-1}$) and divergent field (m/s) : a) climatology, b) low SAM, c) high SAM.

5.3.3.1 Interannual variability of Hadley circulation index during winter and summer

The time series of the intensity of velocity potential during the period from 1980 to 2010 is shown in fig 5.7 (a, b). The intensity is represented in terms of Hadley circulation index. The interannual variability of index is taken over northern hemisphere during winter and summer in the latitude of maximum velocity potential. During winter (fig 5.7a), the negative velocity potential has an average of about −35 units with a standard deviation of 3.817. High negative values are observed during 1983, 1989, 1992, 1995, for a period

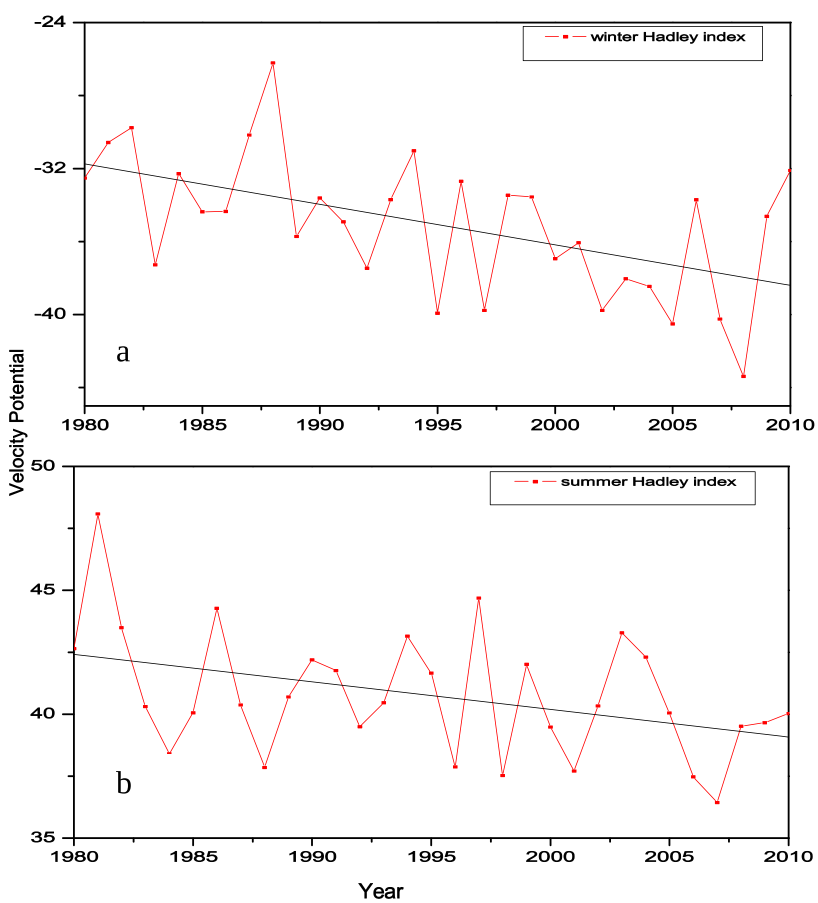

Figure 5.7. Time series of Hadley circulation index during the period from 1980 - 2010: a) winter (DJF) b) summer (JJA).

from 2002 to 2005, 2007 and 2008, while lower value is observed only in 1988. The winter shows intensification of Hadley circulation index. The time series of summer Hadley circulation index is shown in figure 5.7 b for a period from 1980 to 2010. The average summer value is 42.69 with a standard deviation of 2.53. The index shows larger decrease in the positive values during 1984, 1985, 1988, 1992, 1998, 2000 and during 2005 to 2010. The summer time series exhibits decreasing strength of Hadley circulation index.

5.4 Discussion

The present study has studied the SAM associated variability on the velocity potential and circulation pattern during boreal winter and summer. The climatological summer velocity potential value is high over the western pacific while low over the Atlantic to African region. During low SAM, the enhanced positive peak value is shifted to the east of climatological mean position and the negative peak is enhanced by 40 units. While the positive SAM period exhibits same as in climatological, with slight variation in the mean position of peak values. During negative SAM years the divergent flow from the Pacific region is enhanced showing increase in upward motion. At the same time the convergence in the Atlantic and southeast Pacific is also enhanced as a result of sinking motion. During positive SAM, the observed maximum velocity potential is 20 units less than the negative SAM period, where in the case of maximum negative potential, the value is 40 units less during high SAM than that observed during negative SAM.

The winter climatological positive value is 80 units less than summer values and the negative value is less than 40 units. During low SAM, the divergence is enhanced in the Pacific with a positive value greater than 40 units compared to the winter climatology and is shifted to the east. The negative value also seems to be enhanced by 20 units in the southeast Pacific. As a result, the ascending and descending motion is enhanced during low SAM period. During high SAM the area of divergence and convergence is slightly enhanced. In winter, the maximum positive potential during low SAM is 140 units and during high SAM is 100, but there is no difference in the maximum value of negative velocity potential during the phase difference of SAM. The negative peak at southeast Pacific is slightly varied during high SAM showing 20 units less than the observed value during low SAM period.

Understanding the Hadley cell using the zonal mean velocity potential, it is observed that the low SAM during summer shows slight increase in upward motion in the northern hemisphere than climatology. But it is well evident that the pattern has not varied due to the change in the phase of SAM. During summer, the ascending limb of Hadley cell can be varied due to monsoon and the seasonal march of summer induce sharp land sea temperature contrast. So the localized cell has their upper hand in controlling the prevailing condition. The intensity difference of local Hadley cells are well observed in Pacific, Atlantic and monsoon regions (Hu et al., 2007). The study further analysed the Pacific region in order to understand whether the SAM can modulate the local circulation. From the result it is obvious that the SAM has high impact in local circulation. While comparing the climatology, the descending motion is slightly reduced in the southern hemisphere while the ascending motion is enhanced in the northern hemisphere during the low phase. The high phase shows increased sinking in southern hemisphere.

During winter, the zonal mean velocity potential (Hadley circulation) shows variation due to the extratropical influence. The ascending and descending limb is reduced than climatology during the period of low SAM. While the positive SAM exhibits slight variation from climatology.

In the present study, Walker circulation is represented as the annual deviation of zonal mean. As a result, the Hadley and the seasonal and monsoon effect is removed. From the analysis it is clear that during low SAM years the ascending and descending movement has increased in west and east Pacific area where the difference between the western Pacific region and Peru coast is 200 units. While for positive SAM years it is decreased to 120 units and is comparable with the climatolgy. So it can infer that the anomalous variations occur in sea surface temperature over the western Pacific during negative years of southern extratropical phenomenon and moreover the Walker circulations also get stronger. It is noted that the change in Walker circulation is important in the genesis of El Nino events (Philander, 1990). Time series analysis of Hadley circulations shows intensification during winter and slight decreasing index during the summer periods.

The climatology obtained in this study is compared with that of Tanaka and their research group (Tanaka et al., 2004). In January and July it seems the positive and negative values are less than 20 units from their observations. The climatlogy of Walker circulation is varied slightly in the area of convergence and divergence, as seen in the present study. The updated Hadley circulation index in this study is showing further intensification during winter till 2010, while the summer cell shows slight decreasing trend.

The difference noted may be possibly due to the period of consideration and also due to seasons. Tanaka et al. (2004) has considered individual months, whereas the present study focused in the winter (DJF) and summer (JJA) seasons. The present study has shown the climatology of post satellite era (1980 -2010), the period, after the shift is observed in many parameters. It has been already noted that the Hadley cell has anomalous changes after late 1970 (Joanson et al., 2009; Feng et al., 2011).

The present investigation has observed noticeable changes in both the Hadley and Walker circulations due to the variability in southern extratropical latitudes. It is noted that the SAM showed its positive trend in late 20th century (Marshal, 2003). The models predict further intensification of Southern Annular Mode (Thompson et al., 2000; Thompson and Solomon, 2002) due to green house changes (Kushner et al., 2001; Cai et al., 2003). This shows that the variations are occurring form the late 20th century in certain features. The study has presented the changes of circulation change associated with SAM that will help in better understanding of tropical variability.

5.5 Conclusion

The present study analysed the variation of velocity potential and circulation during the phase difference of Southern Annular Mode and found that anomalous changes occur in the positive and negative values of velocity potential. These variations are observed during both winter and summer seasons. The winter Hadley cell shows large variation during the high and low phase of southern extratropics. The summer variation in Hadley circulation is confined only to regional circulation. Anomalous Walker circulation is observed during the positive and negative phases of SAM. Moreover, the climatological value of circulation is carried out during post-satellite period that helps to understanding the climatology after the climate shift period. The updated winter Hadley cell shows further intensification, while slight negative trend is observed during the summer period.

Chapter 6

Signature of extratropical influence on the summer monsoon over India

6.1 Introduction

The spatial and temporal disparities of summer monsoon rainfall over India influence its economy and agriculture to a great extend. The variability in precipitation in turn depends on the circulation pattern and SST anomalies associated with it (Shukla, 1975; Rao and Goswami, 1988; Yamazaki, 1988; Clarke et al., 2000). Occurrence of monsoon variability due to the global variation of phenomenon like El Nino - Southern Oscillation (ENSO), Quasi Biennial Oscillation (QBO) and Indian Ocean Dipole (IOD) are also well documented (Webster et al., 1999; Ashok et al., 2001; Claud et al., 2007).

El Nino - Southern Oscillation is one of the dominant modes that influences the Indian monsoon rainfall (Shukla and Paolino, 1983; Rasmusson and Carpenter, 1982; Webster and Yang, 1992). But recently, weakening relationship of ENSO with Indian summer monsoon is reported (Kumar et al., 1999; Sudipta et al., 2004). In addition, the relationship between IOD and Darwin pressure index seems to be reduced due to the decrease in the IOD occurrence (Swadhin and Yamagata, 2003), which is one of the index influencing the Indian monsoon. It has been reported that the southwest monsoon flow over Indian region decreased due to the altered relation of ocean and atmosphere (Ramesh et al., 2009). Even the monsoon and

Tropospheric Biennial Oscillation (TBO) relation has changed after 1976 (Pillai and Mohankumar, 2009). These recently altered relationships of Indian monsoon with various parameters may sometimes excite other dominant mode of variability that can influence the precipitation pattern.

Studies have shown plausible relationship of monsoon with high latitude variability. This relationship is established through the linkage of Indian monsoon with North Atlantic Oscillation (NAO). It is reported that, there exist significant relation of monsoon with the boreal winter and spring-time NAO (Dugam et al., 1997). Indian Ocean warming has related with the changes of polar vortex in northern and southern hemisphere (Shuanglin, 2009), which act as an element for the monsoon (Shukla and Misra, 1977; Joseph and Pillai, 1984).

In southern hemisphere also there exist a flip-flop mass distribution between middle and high latitude, known as Southern Annular Mode (Gong and Wang, 1999). The influence of dominant mode of extratropical variability, called as Southern Annular Mode (SAM), on the moisture transport, sea surface temperature (SST) and regional rainfall are well documented (Boer et al., 2001; Rao et al., 2003; Screen et al., 2009). The seasonal impact of Antarctica variability is noticed upto the tropical northern belt (Reason and Rouault, 2005; Gillet et al., 2006). Further study shows that precipitation anomalies in southern and northern Australia are associated with SAM (Belinda et al., 2007). Decrease in precipitation due to SAM anomalies over the northern hemisphere is detailed through the monsoon over China (Nan and Li, 2003; Zhiwei et al., 2009), revealing that the leading mode influences in a wider sense. This indicates that SAM can influence both northern and southern hemispheric monsoon patterns. But the relation of these modes (southern and northern) to Indian monsoon is not yet well studied. Since the Asian summer monsoon circulation originates in the oceanic region of southern hemisphere, it is interesting to search for a possible association between SAM and summer monsoon over Indian subcontinent.

The main objective of this study is to investigate whether there exist any possible link between Southern Annular Mode (SAM) and Indian Summer Monsoon Rainfall (ISMR). Precipitation changes were considered to understand the possible link between SAM and ISMR. The circulation changes, SST anomalies and moisture properties associated with variability of SAM were analysed during boreal summer.

6.2 Data and Methodology

For the present study, gridded rainfall data ($1^0 \times 1^0$) procured from Indian Meteorological Department (Rajeevan et al., 2006), updated for a period of 58 years extending from 1951 – 2008 is considered. The updated Southern Annular Mode Index (SAMI) is used for the same period from 1951 – 2008 (Nan and Li, 2003). The SAMI of Nan and Li (2003) is used because it shows better negative correlation in the zonal-mean sea level pressure anomalies between 40^0 S and 70^0 S than that between 40^0 S and 65^0 S, developed by Gong and Wang (1999).

Southern Annular Mode Index is the difference of normalized monthly zonal mean sea level pressure between middle and high latitudes. High and low SAMI years are taken above and below one standard deviation in order to understand the variability of Indian summer monsoon during June (correspond to the onset phase) and July-August (represents the wide spread rainfall period). Correlation analysis of SAMI and ISMR has been carried out for a period from 1951 – 2008. Composite difference of high – low SAMI years is also used to establish the linkage between SAMI and ISMR. Difference of weak minus strong rainfall is computed for below/above one standard deviation.

Sea surface temperature (SST) data obtained from National Oceanic and Atmospheric Administration/National Climate Data Center (NOAA/NCDC) Extended Reconstructed Sea surface temperature version 3b (ERSST. v3b) for a period from 1951 – 2008 (Smith and Reynolds, 2004), has been utilised in this study to understand the difference during high and low SAMI period.

For the analysis of circulation anomalies associated with the high and low SAMI index, we used monthly re-analysis data of NCEP/NCAR (Kalnay et al. 1996). The SAMI high/low related variation on circulation pattern of wind, specific humidity and moisture transport is studied. The profile of vertical velocity (w) is used to understand the variation in ascending and descending area during monsoon season.

The horizontal flow of moisture (qu) in kg m^{-1} s^{-1} from surface to 300 hPa level is calculated, using the equation

$$qu = \left(\frac{1}{g}\right) \int_{300}^{p_S} q\, u\, dp$$

Where q is the specific humidity, u is the zonal component of wind field, p_s is the pressure at the surface (1000 hPa) and dp is the pressure difference between the layers.

6. 3 Results

Southwest summer monsoon over the Indian sub-continent is generally expected to begin around early June and cease by the end of September. Thar desert and the adjoining areas of the northern and central Indian subcontinent heats up considerably during the hot summer season causing a low pressure area over the north and central parts of India. To fill up the low-pressure zone, the moisture laden winds from the Indian Ocean dash into the sub-continent. The monsoon current for the Indian subcontinent is basically a product of southeast trade winds originating from a high pressure mass centered over the South Indian Oceanic region, therefore the SAM may have an influence on the Indian summer monsoon.

6.3.1 Association between June SAM index and the onset phase of Monsoon

In the present study, the possibilities of in-phase relationship between SAM index and monsoon onset period are studied. For this, onset phase of the monsoon during the month of June over the Indian region is correlated with June SAM index. The study further extends to interpret SAM index associated variability on the monsoon parameters. A correlation study has been made using the June SAM index with rainfall data at each grid point over the Indian region during the month of June to study the relationship between SAM and early phase of monsoon over the Indian subcontinent.

Linear correlation between June SAM index and ISMR during the onset phase of monsoon (June) for the period 1951-2008 is shown in figure 6.1. Significant positive correlation is observed in the Indo-Gangetic plain. The area of significant positive correlation between the rainfall during the onset phase and the difference of SAM index locates in the region of monsoon trough to the north of 20^0 N. Large areas of significant positive correlation are noticed between the belt of 24^0 N and 34^0 N. Though, negative relations are observed over the extreme southwest and southeast coast, north and north east of India, the relations seems to be insignificant.

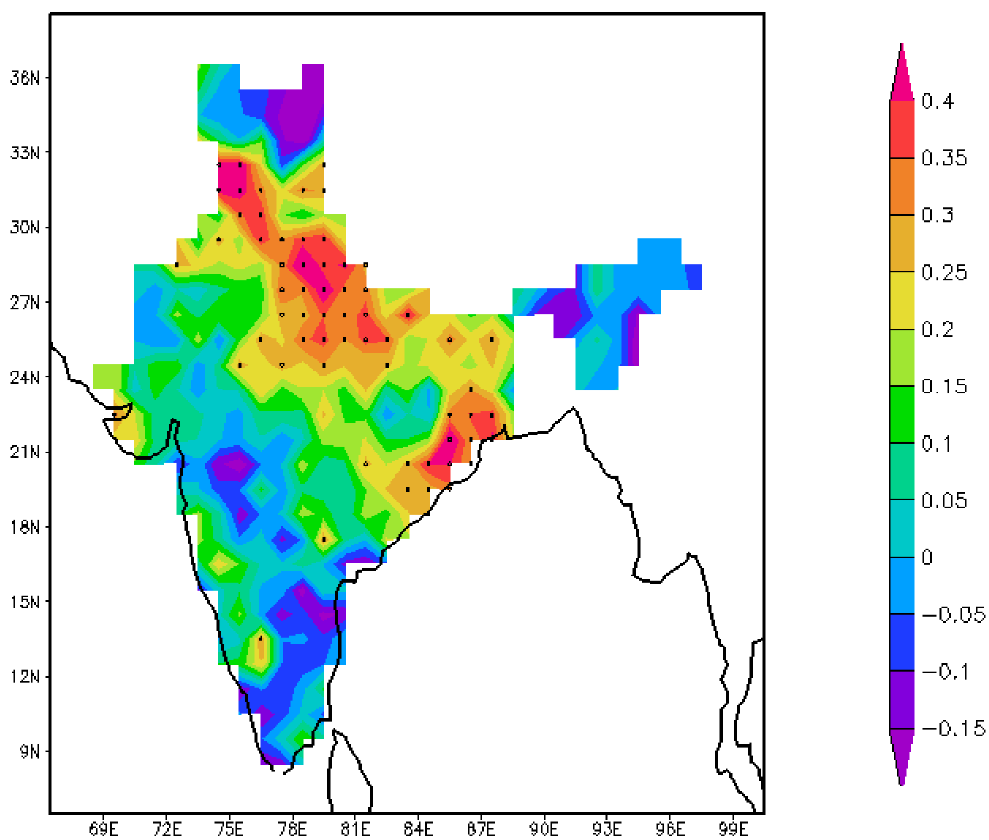

Figure 6.1. Correlation between the June SAMI and June precipitation in India for the period 1951-2008. The positive (negative) correlation coefficients that are significant at the 95% confidence level are dotted.

Since, significant relation is observed between June rainfall and June SAMI, the effect of SAMI on ISMR is analysed using other parameters that affect the monsoon. For this, zonal wind pattern at 850 hPa level during the June is utilised.

One of the important parameters that affect the widespread rainfall during Indian summer monsoon season is the low level jet stream, which transports moisture from Arabian Sea to the mainland. The zonal wind begins as southeast trades far in southern hemisphere, migrate to northern hemisphere during boreal summer. After crossing the equator, it blows as southwesterly due to Coriolis force and provides moisture to Indian subcontinent; thereby intensifies the monsoon activity. The zonal wind difference during June corresponding to high and low SAMI is shown in figure 6.2. During high SAMI, the zonal wind strength increases from south of

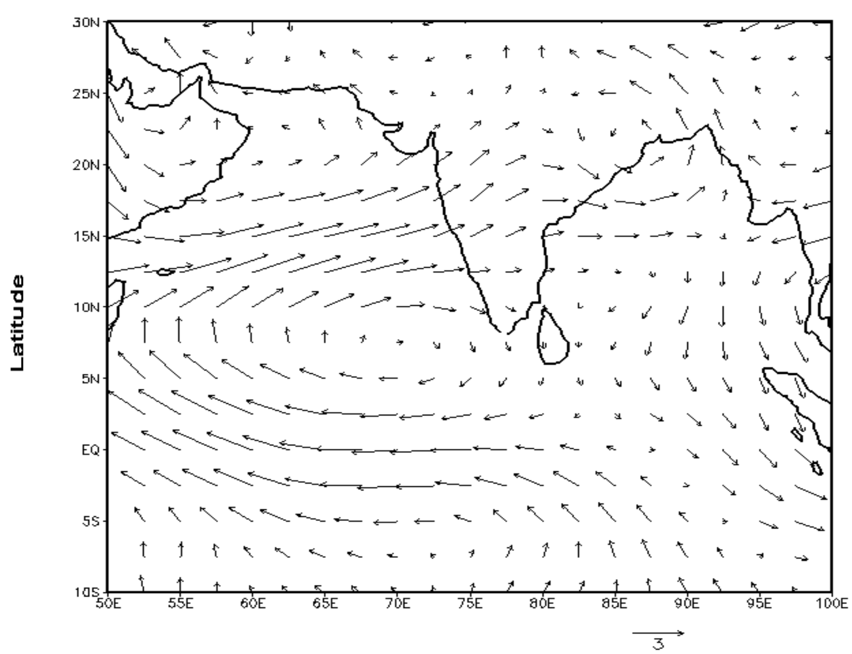

Figure 6.2. Composite difference of the wind at 850 (ΔV_{850}) hPa between high and low SAMI years during the month of June.

the equator, and further enhancement is observed along north of 12^0 N over the Indian mainland and nearby regions. Strong anomalous cyclonic flow occurs around 20^0 N and 85^0 E over the monsoon trough region, while weak anticyclonic flow occurs over the southernmost region, centered over west of Srilanka.

Zonal wind pattern shows anomalous variation over India and nearby regions. Moreover enhancement of westerly flow and anomalous cyclonic flow occurs near the area of observed positive (significant) correlation (see figure 6.1) that may influence the precipitation pattern existing there. So the study has further focused over this area (25^0 N - 30^0 N; 75^0 E - 85^0 E), where significant positive relations are observed. Vertical velocity and specific humidity over the specified area is analysed to find whether ascending motion and increase in moisture are observed during high SAMI period. These changes have effect on the intensification of rainfall over this region**.** Figure 6.3 shows area averaged vertical velocity (25^0 N - 30^0 N; 75^0 E - 85^0 E**)** of June during the composite of high and low June SAMI. Both high and low

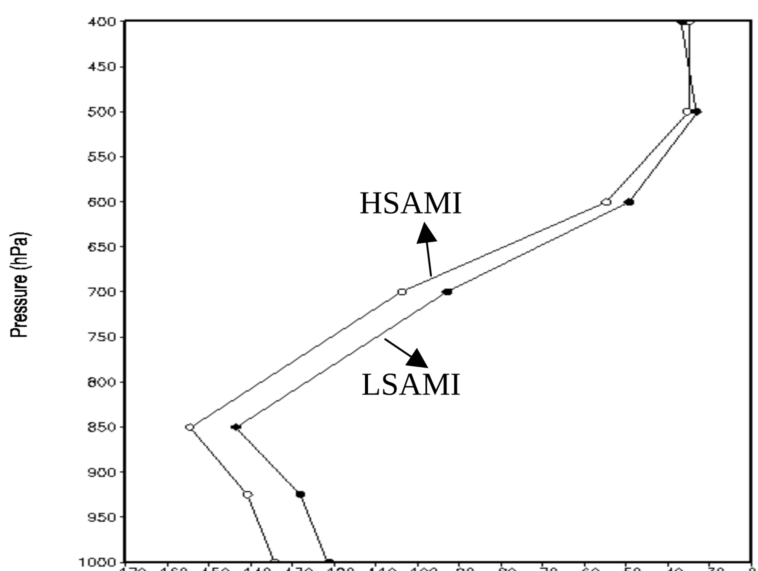

Figure 6.3. Composites of the June vertical velocity (in 10^{-3} Pa s^{-1}) high (dotted line) and low SAMI (dark dotted line) years, where high/low SAMI is represented as HSAMI/ LSAMI.

SAMI show increase in ascending motion in the troposphere extending from 1000 hPa to 850 hPa over the domain. Above this level, the magnitude of ascending motion decreases to the upper tropospheric level. During high SAMI period, ascending motion enhances in the troposphere compared to that during the period of low SAMI years. The increase of ascending motion during high SAMI years is stronger in the lower troposphere up to 850 hPa than low SAMI period. Though the vertical motion is enhanced during high SAMI, the difference between the high and low SAMI decreases as it extends towards upper troposphere.

Variability of moisture over the particular region is analysed using specific humidity. Figure 6.4 shows area averaged specific humidity (25^0 N - 30^0 N; 75^0 E - 85^0 E) of June during the composite difference of high and low June

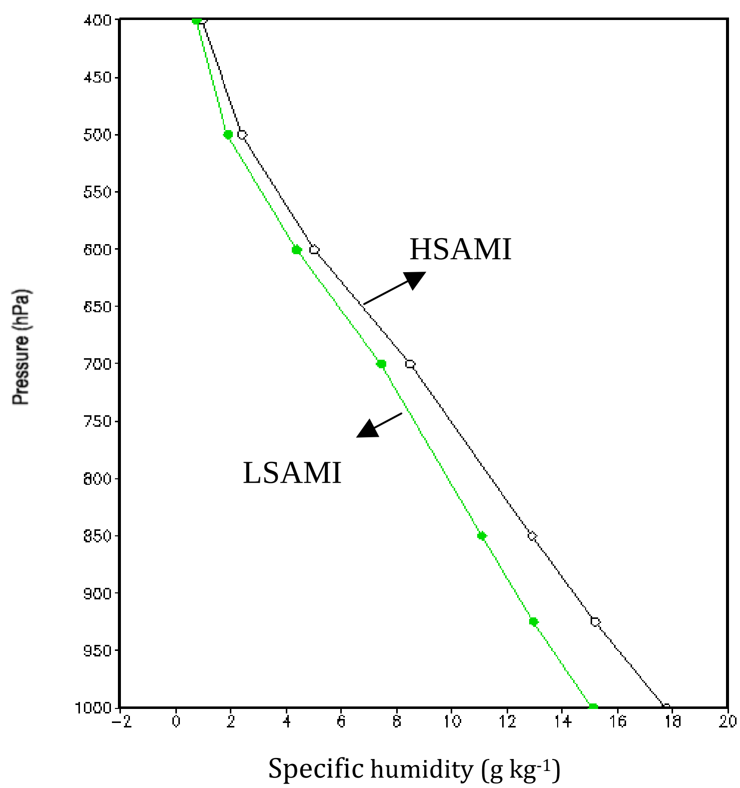

Figure 6.4. Composites of the June specific humidity (g kg^{-1}) for June high (dotted line) and low SAMI (dark dotted line) years, where high/low SAMI is represented as HSAMI/LSAMI.

SAMI. Compared to low SAM index, it is evident that high SAM index leads to an increase in specific humidity in the troposphere extending from 1000 hPa to 500 hPa. The difference of specific humidity between high and low SAMI is larger in the lower troposphere, and it gradually reduced towards the upper tropospheric region. In the lower troposphere level, an enhancement of 5-8 % of specific humidity is noted during high SAMI period. It can infer that increase in humidity as well as the enhancement of upward vertical velocity induces precipitation over the specific region during the period of high June SAMI years.

6.3.2 SAMI relation to widespread rainfall period (July-August)

During July-August, precipitation occurs over most parts of India. Prevailing synoptic conditions have strong undulations during this time. So we consider June and July-August SAMI in order to understand the variability of monsoon features during July-August period. The in-phase and lead-lag relationship is analysed to find whether there exist any predictive skill over the precipitation and circulation pattern over Indian region due to southern high latitude circulation (SAMI).

6.3.2.1 Correlation between SAMI and July-August rainfall

Figure 6.5a shows correlation between June Southern Annular Mode index and the July-August Indian summer monsoon rainfall for the period 1951 - 2008. June SAM index is negatively correlated over most of the areas of Indian region. Areas of significant negative relations are observed in between the belt of 18^0 N and 27^0 N, an out of phase relation occurs over the northernmost region, and also over the southwest sector of India. Significant positive correlations are noticed over the northeast of India.

Figure 6.5b shows the correlation between July-August SAM index and the July-August Indian summer monsoon rainfall for the period 1951 - 2008. Significant negative relations are observed in north Indian region and positive relations are observed to the northeast India. Negative relations are observed over the eastern part of central India along 84^0 E and along southwest coast of India, but the relations are insignificant. From figures 6.5 a & b, it is well evident that the rainfall of July-August is significantly more correlated with June SAMI than July-August SAMI.

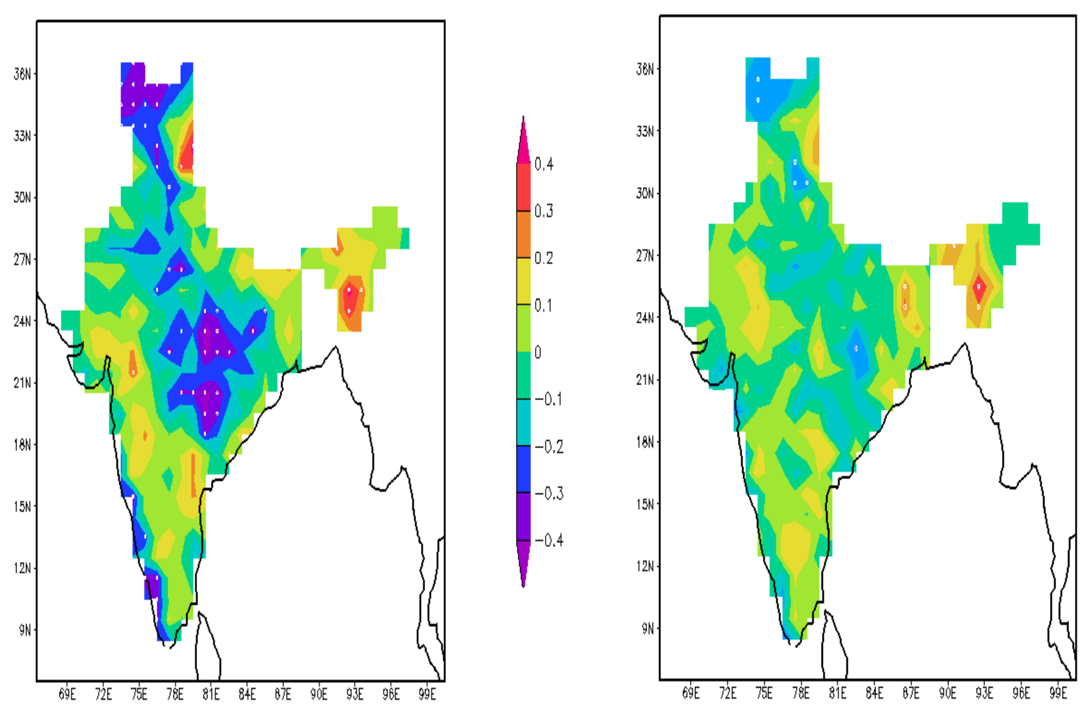

Figure 6.5. Correlation between the July-August precipitation for the period 1951-2008 and a) June SAMI and b) July–August SAMI. The positive (negative) correlation coefficients that are significant at 95% confidence level are dotted. The contour interval is 0.1.

6.3.2.2 Composite difference of June SAMI and July-August rainfall

Even though the correlations are not so high, the spatial structure shows a negative relation over most of the area, particularly in the southwest coast of India, where rainfall is maximum during these months. So, this study explore the difference of July-August ISMR corresponding to high and low June SAMI above/below 1.5 sigma level. In addition to that, weak minus strong ISMR for July-August is also plotted to find the magnitude and spatial similarity between them.

Figure 6.6a shows the difference of weak minus strong monsoon rainfall during July-August. During weak monsoon period, negative anomalies are strong in the southwest coastal regions, the regions along 81^{0} E and the extreme north of India, where positive anomalies are noticed in the northeast India. The difference of July-August rainfall corresponding to high and low June SAMI is shown in figure 6.6b. During high June SAMI, rainfall anomalies are negative over the southwest coastal regions, south central India along 81^{0} E and 22^{0} N and in the northern most regions of India. The magnitude of negative anomalies is stronger in between 8^{0} N and 20^{0} N along the southwest coast, and along the area near 22^{0} N and 81^{0} E. Strong positive anomalies occur over northeast India and at the same time weak positive anomalies take place over some areas of southeast India.

During weak monsoon conditions, precipitation reduces along southwest coastal regions and at the same time heavy rainfall occurs in northeast regions. High SAMI period also shows similar characteristics of a weak monsoon conditions with negative anomalies over the west coast and positive anomalies over the northeast regions. The spatial distribution of anomalies is similar in figures 6.6a and 6.6b, which is shown in the patched area along the southwest sector, north, northeast region and area around 22^0 N. At the same time, there is difference in the magnitude of anomalies over the specified region. It seems that high June SAMI results in decrease of rainfall along the areas of southwest coast, central and northern India, while precipitation enhances in northeast India.

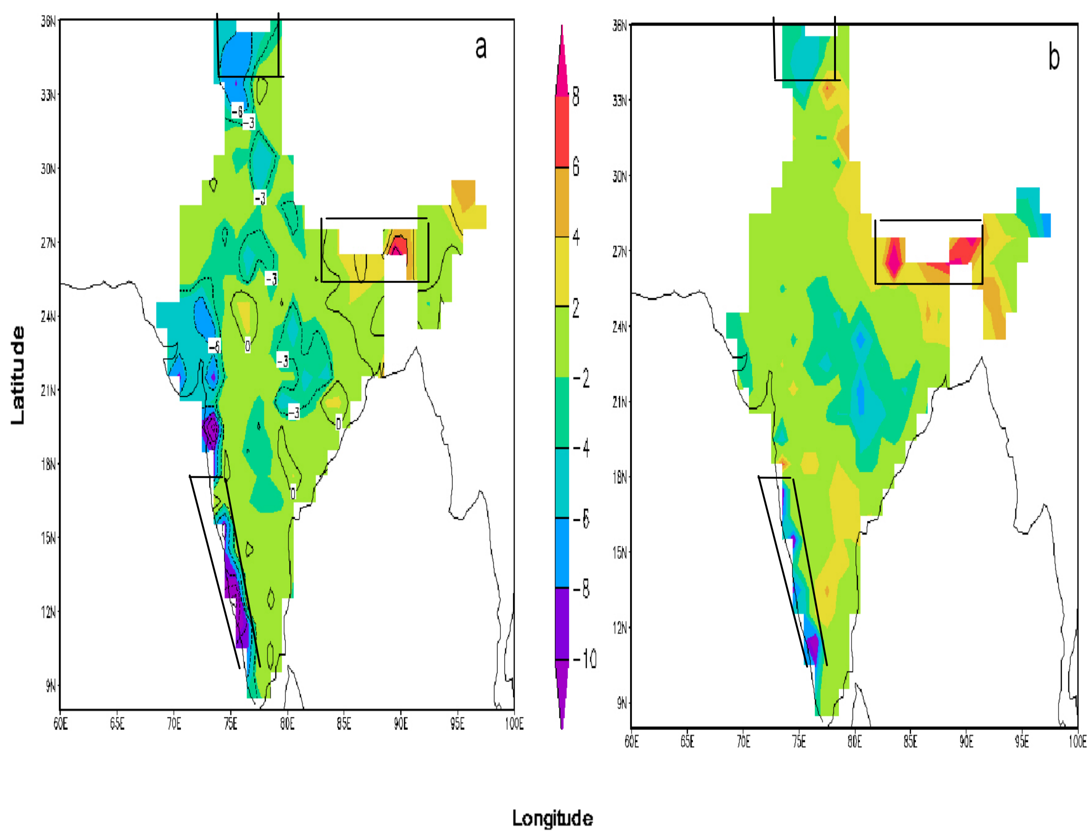

Figure 6.6. Composite difference of the July-August rainfall (mm/day) a) between weak minus strong monsoon and b) corresponding to the difference of high and low SAMI years during June

6.3.2.3 June SAMI linked variability on monsoon parameters

From the above results, it is evident that June SAMI has strong relation with Indian rainfall during July-August. As a result, Indian summer monsoon variability due to high *minus* low Southern Annular Mode has been studied in detail for the parameters like SST, wind, moisture transport and vertical velocity corresponding to the difference of June SAMI is during July -August. Indian Ocean SST is one of the major elements that affect the Indian monsoon (Webster et al., 1998; Yamazakhi, 1988)**.** So the variation of SST associated with SAM is an important parameter considered for the analysis. Moreover, the gradient between the SST in south Indian Ocean and Indian landmass is an important feature for the development of monsoon system. The composite difference of averaged July-August SST corresponding to strong minus weak June SAM index is shown in figure 6.7. Increase in SST is observed during high SAMI to the south of 25^0 S and also over the areas of

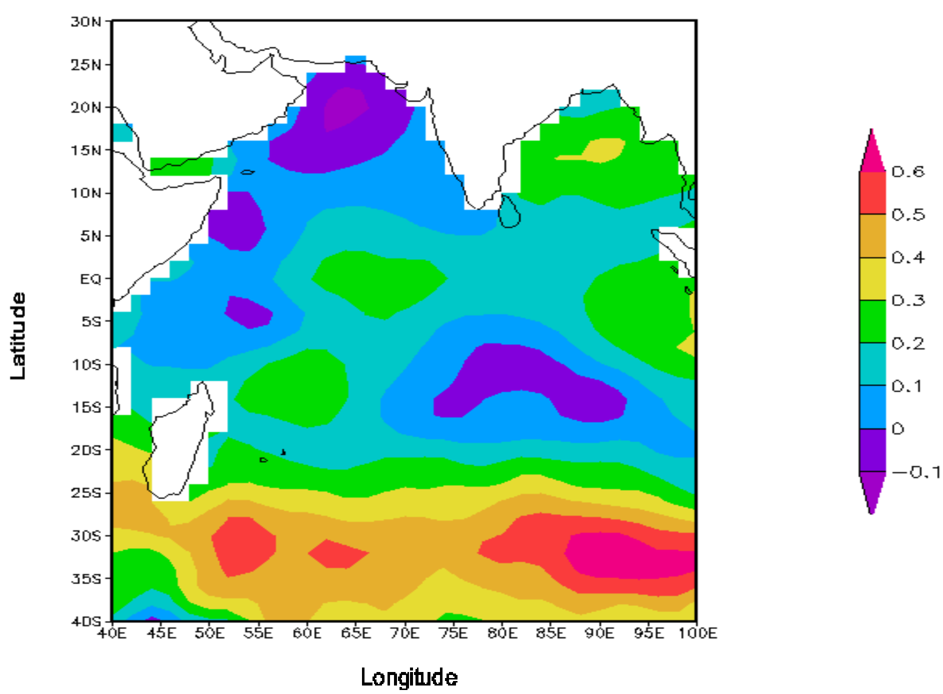

Figure 6.7. Composite difference of July- August SST between the high and low SAMI years during June.

equatorial region (east of 60^0 E) and the Bay of Bengal region. At the same time, during high SAMI, SST slightly decreases in the Arabian Sea and to the west of 60^0 E around equatorial region and area located near 10^0 S and 85^0 E.

The magnitude of warming is intense near 30^0 S, while cooling of 0.1^0 C is observed over the Arabian Sea around 20^0 N. This gradient of SST is favourable for the development of a weak monsoon condition during high SAMI. Moreover, the pattern also induces a weak monsoon circulation and thereby moisture transport. Circulation pattern also indicates the elements of strong monsoon. The major flow at 850 hPa level determines the strength of monsoon and associated rainfall.

The composite difference of averaged July-August vector wind at 850 hPa corresponding to high *minus* low of June SAM index is shown in figure 6.8. During high SAMI, zonal wind at 1.5 km shows enhanced southerly flow

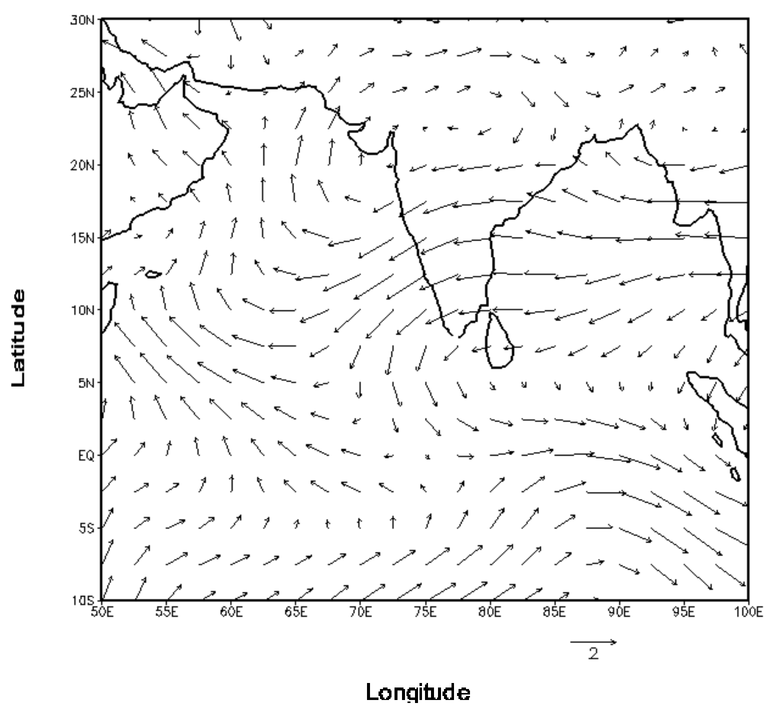

Figure 6.8. Composite difference of July- August wind at 850 hPa between high and low June SAMI years.

towards equator between 50^0 E and 65^0 E. Anomalous easterly flow occurs over south India and northeasterly along the west coast of India, accompanied by enhanced northerly flow along 20^0 N and 60^0 E. Strong anomalous anticyclonic flow is observed in central India during high SAMI. Though the magnitude of variation is less, the flow pattern is disturbed in the southwest coastal regions of India, which results in anomalous variation in precipitation pattern over south India.

The variation in ascending and descending motion is examined using vertical velocity. The height–latitude profile of composite difference of vertical velocity during July-August corresponding to high *minus* low June SAM index is shown in figure 6.9. During high SAMI, enhanced ascending motion occurs from equator to 10^0 N, where the increase in magnitude is high over the upper troposphere. While anomalous descending enhances in between 10^0 N and 25^0 N, the magnitude of variations are large in between 500 hPa and 300 hPa. At the same time, ascending motion enhances in the latitudinal belt between 25^0 N and 35^0 N during high SAMI. From the figure 6.9, it is evident

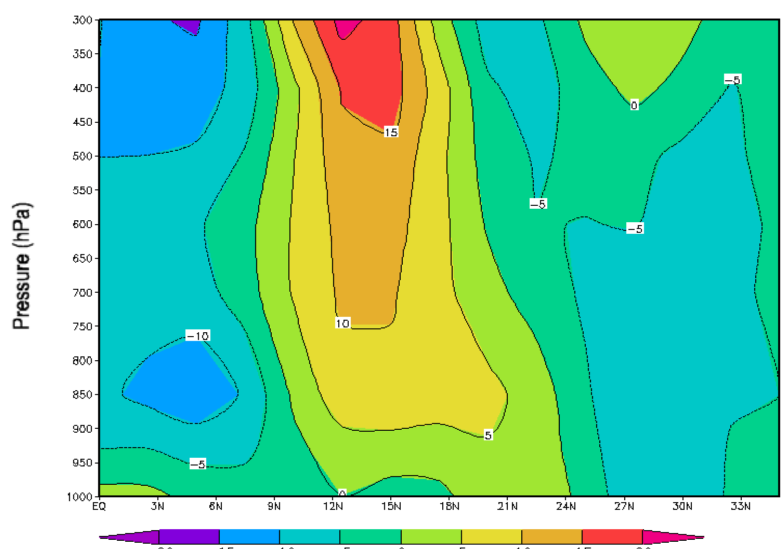

Figure 6.9. Composite difference of July- August vertical velocity (in 10^{-3} Pa s^{-1}) averaged over 65–95^0 E, between the high and low June SAMI years.

that ascending motion enhances nearby equatorial region, but descending motion enhances in south India during the expected heavy rainfall period. Change in zonal moisture transport over south India due to June SAMI is also evaluated.

During monsoon, the transport of moisture through southern area is critical for precipitation. Moisture transport during July-August corresponding to high and low June SAM index is shown in figure 6.10. During high SAMI, moisture transport decreases to the west of 65^0 E in between the belt of equator and 20^0 N. The transport of moisture increases to the south of 6^0 N along the east of 65^0 E, while to the north of 6^0 N, moisture transport decreases. Anomalous decrease is observed near Somalian coast and over the east coast of the Arabian Sea (around 13^0 N) extending to the Bay region. In the near equatorial region, moisture transport seems to be increased, the region of maximum intensity is noticed along 85^0 E. The decrease in moisture transport during high SAMI period along the Indian domain also results in reduction of precipitation.

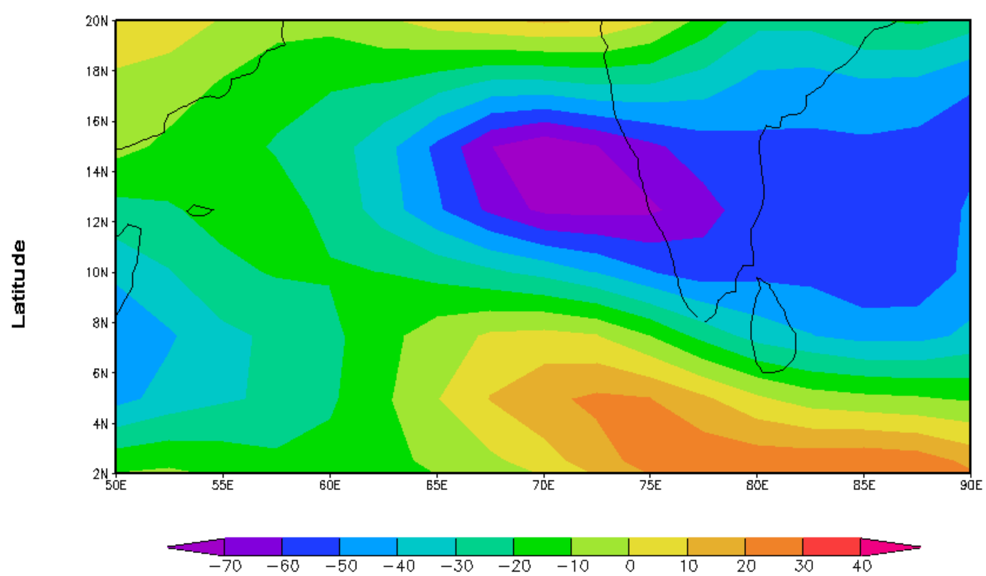

Figure 6.10. Composite difference of July-August moisture transport (kg m^{-1} s^{-1}) corresponding to high and low June SAMI years.

From the above analysis, we observed that the variability of June SAM index influences the summer rainfall over India during onset and wide spread rainfall period. Anomalous pattern similar to weak monsoon condition is observed during high June SAM index. Inorder to understand the interannual variability of SAM index during June, time series is plotted for a period from 1951 to 2008. During 1950 to 1980, positive trend of June SAM index is observed (see fig 6.11). The index showed abrupt positive departure for a couple of years during late 1970's. From early 1980 onwards SAMI showed a negative slope till mid-1990. Thereafter, a gradual increase of positive trend is observed in June SAM index.

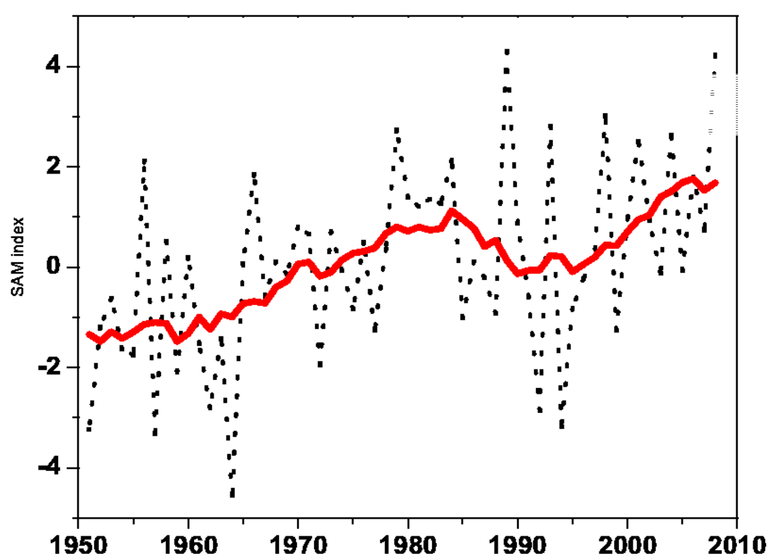

Figure 6.11. Time series of SAM Index during June (dotted lines) along with 10 year moving averages (solid line).

6.4 Discussion

The study reveals that, Southern Annular Mode (SAM) influences the Indian rainfall, and the parameters associated with it during the summer monsoon period. When June SAMI is correlated with June ISMR, positive relations are observed over northcentral India. During positive phase of SAMI, enhanced westerly flow and anomalous cyclone occurs at 850 hPa level in the near regions of positively correlated area. At the same time anomalous anticyclone is observed over the extreme part of south India. During June, average rainfall of about 98.34 mm occurs in the positively correlated area (25^0 N - 30^0 N; 75^0 E - 85^0 E). During high SAM years, the specific humidity slightly increases in troposphere from 1000 hPa to the 500 hPa. The magnitudes of humidity variation between high and low SAM years are stronger in lower troposphere. At the same time, vertical velocity also shows enhanced ascending motion over this region. This reveals that ascending motion is enhanced and the atmosphere is fed by high moisture content over this region during positive phase of SAM, so precipitation may enhance over the region.

During July–August, monsoon has active and break spells of rainfall, where break situation affects the low level flow and also the moisture transport (Joseph and Sijikumar, 2004), thereby affecting the precipitation anomalies. Increase in break spell, during July-August over Indian region, and decreasing low level flow along with reduced moisture transport is documented (Ramesh et al., 2009). This makes clear that the properties of monsoon have changed during recent decades, and pose a question whether any other relationship is also influencing Indian summer monsoon during this season. So the in phase and lead-lag relation between Indian monsoon and SAMI seems to be a possible link.

To understand the predictive potential of SAM, we correlated June and July-August SAMI with July-August rainfall. The relation shows that the June SAMI has significant relation compared to July- August rainfall. Correlation between the June SAMI and rainfall over India during July-August shows significant negative correlation over most of the areas except in the northeast region of India**.** The southwest sector of India where the maximum rainfall occurs during summer monsoon due to orographic ascent, observed significant negative relation during high SAM years. During high SAMI years, the precipitation over Indian region shows pattern similar to weak monsoon rainfall. This is well evident in the southwest sector of India, where negative anomalies observes during the composite analysis of high SAMI.

During high SAM years, SST warming is observed over tropical regions of southern hemisphere and at the same time cooling over western sector of Arabian region. This can alter the horizontal temperature gradient that prevails during good monsoon, which in turn modifies the monsoon flow and moisture transport. Warming of SST is also noticed in the Bay of Bengal region. It is reported that, the western Arabian Sea cooling may cause reduction in rainfall (Shukla, 1975). While, changing relationship of Indian ocean with monsoon has already been noticed after the climate shift (Clarke et al., 2000).

The low level flow at 850 hPa (Somali jet stream) over South India seems to weaken slightly during high June SAM index. The flow pattern over southwest coastal regions which seems to be changed will possibly affect the precipitation in west coastal regions of India Anomalous anticyclone is also observed over central India. The present analysis shows a reduced moisture flow in between equator and 20^0 N during high SAMI, indicating lack of moisture during peak monsoon period. Vertical velocity shows enhanced ascending motion over the equatorial region but increased descending motion over south India. It can conclude that the positive phase of southern extra tropical phenomenon during June is not conducive for monsoon of July-August.

From the study, it is clear that SAM induct strong association to Indian monsoon. When June SAMI has positive polarity, it results in enhanced precipitation during June over northcentral India and moreover the circulation and moisture is also favourable for the increase in precipitation over that area. At the same time, high SAMI during June is not conducive for the precipitation over most parts of India during July –August, which is also evident from the anomalous circulation, moisture transport and vertical velocity. Simultaneously precipitation increases in the northeast regions of India during the positive phase of SAM. Sea surface temperature (SST) also favours a weak thermal gradient between northern and southern hemisphere, which will adversely affect the monsoon circulation, and thereby moisture transport. During this time, anomalous variations in the zonal flow at 850 is also noticed (Rao et al., 2004; Sathiyamoorthy, 2005; Joseph and Simon, 2005), these flows are important feature of monsoon.

Strong positive phase of SAM is seen during recent decades. The increasing trend of SAM is already reported during the second half of the 20th century (Gong and Wang, 1999; Kidson, 1999; Marshall, 2003; Miller et al., 2006). Moreover, SAM variation seems to be reinforced by anthropogenic force and ozone changes (Arblaster and Meehl, 2005). Another interesting thing is that, the North Annular Mode and ENSO has combined affect over Indian Summer Monsoon (Dugam, 2006). But, studies also shown that the long-recognised negative correlation between Indian monsoon rainfall and ENSO has weakened rapidly during recent decades (Sudipta et al., 2004). So, we summarise, SAM is affecting Indian monsoon and circulation to a wider sense. The observational evidence perceived in the present study reveal that SAM has a strong linkage with the circulation and precipitation pattern of the Indian summer monsoon.

6.5 Conclusion

Southern Annular Mode shows indication of link with Indian monsoon during summer monsoon period. But the relation of June SAM is distinct with June and July-August rainfall. During June, high SAMI enhances the June rainfall over northcentral India. The circulation and moisture also provides an enhancement of rainfall over this region. The correlation between June and July-August SAMI with rainfall of July-August shows that the correlation is significantly high with June SAMI than July-August SAMI. During July-August, rainfall seems to be decreased in most of the regions due to the impact of high June SAMI except in northeast India where significant increase in precipitation is observed, which is similar to weak monsoon rainfall observed during high SAMI. The SST, circulation and moisture properties demonstrate a situation of weak monsoon condition. Anomalous variation in the low level flow and reduced moisture transport occur during high SAMI may affect the monsoon. Positive phase of June SAMI is observed in recent decades, suppresses active rainfall spells over southwest India, where the summer monsoon starts has its first spell and has strong spell during monsoon period. June SAMI is useful to quantify the variability of parameter associated with summer monsoon and to some extend for the prediction of the widespread rainfall over the country during the July-August. Concisely, the Southern Annular Mode seems to be associated with the rainfall variability and distribution of Indian monsoon.

Chapter 7

Modulation of North Atlantic Oscillation on the SAM-Monsoon link

7.1 Introduction

Extra tropical influence on atmospheric parameters are observed through the anomalous behaviour of North Atlantic Oscillation (NAO), North Annular Mode (NAM) and Southern Annular Mode (SAM) (Wallace and Gutzler, 1981; Hurrell, 1995; Beniston, 1997; Hurrell and van loon, 1997; Thompson and Wallace, 1998; Gong and Wang, 1999; Wanner et al., 2001). In this, NAM and SAM are dominant mode of variability in northern and southern hemisphere respectively. These modes are oscillation of mass between mid and high latitudes that leads to positive and negative phases. While, NAO is a north hemispheric pattern, and is the pressure difference between Azores high and Icelandic low. These three extra-tropical oscillations have impact over the surface weather features, storm tracks and are identified to produce extremes in temperature and precipitations (Lefebvre et al., 1997; Cullen and Menocal, 2000; Greatbatch, 2000; Thomas and Wallace, 2000; McCabe and Muller, 2002; Rao et al., 2003; Sulan and Jianping, 2003; Turkes and Erlat, 2003).

In addition to that, these modes have known to influence the precipitation pattern over the tropical Indian subcontinent. Some of the studies suggested that NAO is related to Indian monsoon variability (Dugam et al., 1997, Kakade and Kulkarni, 2012). It is noted that April NAO index has negative correlation with Indian summer monsoon of the concurrent year (Kakade and Dugam, 2000). Summer precipitation over India has shown teleconnection with various features all over the globe, and this rainfall has high influence in the socio – economic growth of India. El Nino, this Pacific phenomenon is one of the important parameter that is noted to influence the summer rainfall over India (Sikka, 1980; Rasmussen and Carpenter, 1982; Shukla and Paolino, 1983; Webster and Yang, 1992; Mehl and Arblaster, 2002). But the effect of El Nino to Indian summer monsoon has weakened during recent decades (Sudipta et al., 2004).

Other parameters that has altered its relation to Indian summer monsoon, one of which is Indian Ocean Dipole (IOD) and the other is Tropical Biennial oscillation (Pillai and Mohankumar, 2009). The climate shift during 1970's has also contributed to increase in the break spells of Indian monsoon thereby adversely affected the circulation and moisture properties (Ramesh et al., 2009). All this distorted relationship of monsoon with various parameters has lead to investigate new relationship of monsoon (Dugam and Kakade 2004, Munot and Krishna Kumar, 2007). It has been already noted that Indian summer monsoon has relation with extra tropical variability of northern hemisphere (Goswami et al., 2006). North Atlantic Oscillation has shown significant inverse relation with Madden Julian Oscillation (MJO) during break spells of summer rainfall over India and the researchers added that it could be used as a predictor of active/break spells (Dugam, 2008).

On the other hand, the main variability in the southern hemisphere account from SAM (Boer et al., 2001; Screen et al., 2009). The SAM has prominent role in changing the temperature and rainfall pattern (Belinda et al., 2007) and its influence has noted up to northern hemisphere through the variation of precipitation in China (Zhiwei et al., 2009). The SAM oscillation during June seems to influence the summer monsoon over India through changes in precipitation, circulation and moisture properties.

The main objective of the study carried out in this chapter is to understand whether the observed June Southern Annular Mode association to summer monsoon is altered due to the different phase of North Atlantic Oscillation. During July-August, 60% of rainfall occurs all over India, moreover dry and wet spells are also frequent in these months, which will result in drought and flood. Understanding the importance of these months, precipitation during July-August is analysed using these modes. For this June SAM and NAO during April are used to find this relation. April NAO is used because, it is noted that April index shows statistical inverse relationship with Indian monsoon (Kakade and Dugam, 2000). Changes in SST and moisture transport are also considered for the study. An attempt has also been made to understand of this extra tropical variability to Indian summer monsoon, which is not well studied. Moreover, this study focuses on the predictability of summer monsoon precipitation over India through extra tropical variability. It is interesting to observe that whether there exist any other potential predicators similar to El Nino, as a teleconnection factor.

7.2 Data and methodology

The primary data used in the study is the updated Southern Annular Mode Index (SAMI) and North Atlantic Oscillation Index (NAOI) for a period from 1951 – 2008 (Nan and Li, 2003). The SAMI developed by Nan and Li (2003) has shown better negative correlation in the zonal-mean sea level pressure anomalies between 40^0 S and 70^0 S than that between 40^0 S and 65^0 S developed by Gong and Wang (1999). In addition to that, the NAOI build up by them can better represent the spatial and temporal variability of different variables. For the rainfall analysis, gridded rainfall data (1^0 x 1^0) obtained from Indian Meteorological Department (Rajeevan et al., 2006), and updated for a period of 58 years extending from 1951 – 2008 is also considered.

June Southern Annular Mode index and its relation to Indian monsoon during July-August has already discussed in the previous chapter. At the same time, it is noted that the North Atlantic Oscillation during April has significant negative relation with summer monsoon. The study further extends to understand whether April NAO can modulate the observed relation of southern extra tropics and summer monsoon. The SAM and NAO indices are classified (SAMI and NAOI) into four categories. The values of indices above/below +/- 0.3 are included for the study, as a result this modes can better represent their characteristics. The average precipitation during July-August is observed for this relationship. The difference between positive and negative SAMI during the NAOI negative phase is observed primarily. Similarly the different between positive and negative SAMI during the NAOI positive phase is also considered. Variability in SST and moisture transport are also analysed for the above two criteria. Sea surface temperature (SST) data is procured from National Oceanic and Atmospheric Administration/National Climate Data Center (NOAA/NCDC) Extended Reconstructed Sea Surface Temperature version 3 (ERSSTv3) for a period from 1951 – 2008 (Smith and Reynolds, 2004). Zonal moisture transport is also calculated using wind and specific humidity obtained from NCEP/NCAR data set (Kalnay et al., 1996).

7.3 Results

In July-August, summer monsoon has large variability in rainfall distribution over India due to active and break spells. During active spells rainfall increases in specific regions of India. So the dispersion of average rainfall during active is analysed to understand the intensity and spatial distribution of precipitation. For this, years above one standard deviation during July-August have taken, which is shown in figure 7.1.

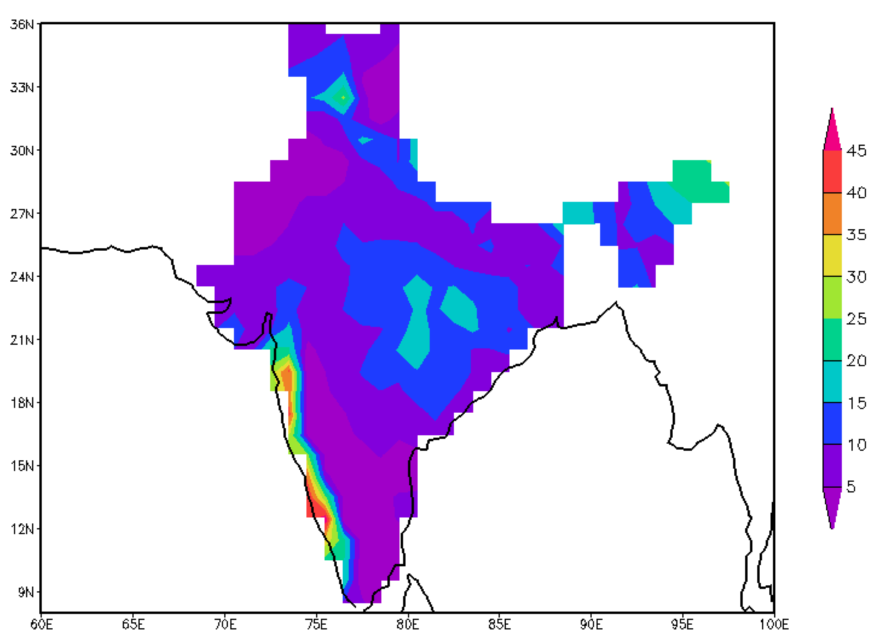

Figure 7.1. Average active rainfall over India during the July-August period

During southwest monsoon (July-August) period, heavy rainfall occurs along southwest coastal regions. Over the, west coastal region near 13^0 N and 18^0 N, precipitation exceeds 35 mm/day. At the same time, precipitation of 20 mm/day occurs along extreme northeast India and also in areas over eastern part of central India. While in central India precipitation exceeds 5 mm/day, where heavy precipitation of about 10 mm/day occurs in the east central region. Along northwest and southeast India precipitation is less than 5 mm/day.

Intense rainfall is observed over southwest region and along northeast India (see fig 7.1). Depending on strong and weak monsoon periods, the precipitation amount has anomalous change in the observed regions. This study further concentrated on the variability of rainfall that can be account due to the simultaneous effect of SAMI and NAOI. Four criteria has been defined based on positive/negative phases of SAM and NAO Indices, and is shown in table 7.1. The years in the table have index values above or less than 0.3 has been used as a result the disparity between the two modes on monsoon can be better understood.

SAM ($^+$) NAO ($^+$)	**SAM ($^-$) NAO ($^+$)**	**SAM ($^+$) NAO ($^-$)**	**SAM ($^-$) NAO ($^-$)**
1958	1952	1956	1951
1970	1954	1966	1953
1982	1957	1971	1961
1989	1959	1973	1963
1990	1962	1978	1975
1993	1964	1979	1988
2001	1972	1983	1995
2002	1977	1984	
2004	1985	2000	
2006	1991	2007	
	1992	2008	
	1994		

Table 7.1. Years representing the four combinations of Southern Annular mode and Northern Atlantic Mode.

Where SAM $^{(+/-)}$ denotes SAMI $^{(\text{positive/negative})}$ and NAO $^{(+/-)}$ denotes NAOI $^{(\text{positive/negative})}$. During 1951 to 2000, SAM $(^{-})$ NAO $(^{+})$ and SAM $(^{-})$ NAO $(^{-})$ phases seem to be active, after that, SAM $(^{+})$ NAO $(^{-})$ and SAM $(^{+})$ NAO $(^{+})$ become prominent. The positive phase of SAMI and alternating phase of NAOI are occurring recently. This shows that the alternating effects of these modes are tested whether this can modulate the July-August precipitation over India. The four modes are schematically represented in figure 7.2. The difference of (a) and (b) and at the same time (c) and (d) are analysed to study the simultaneous influence over for rainfall, SST and moisture transport during July-August.

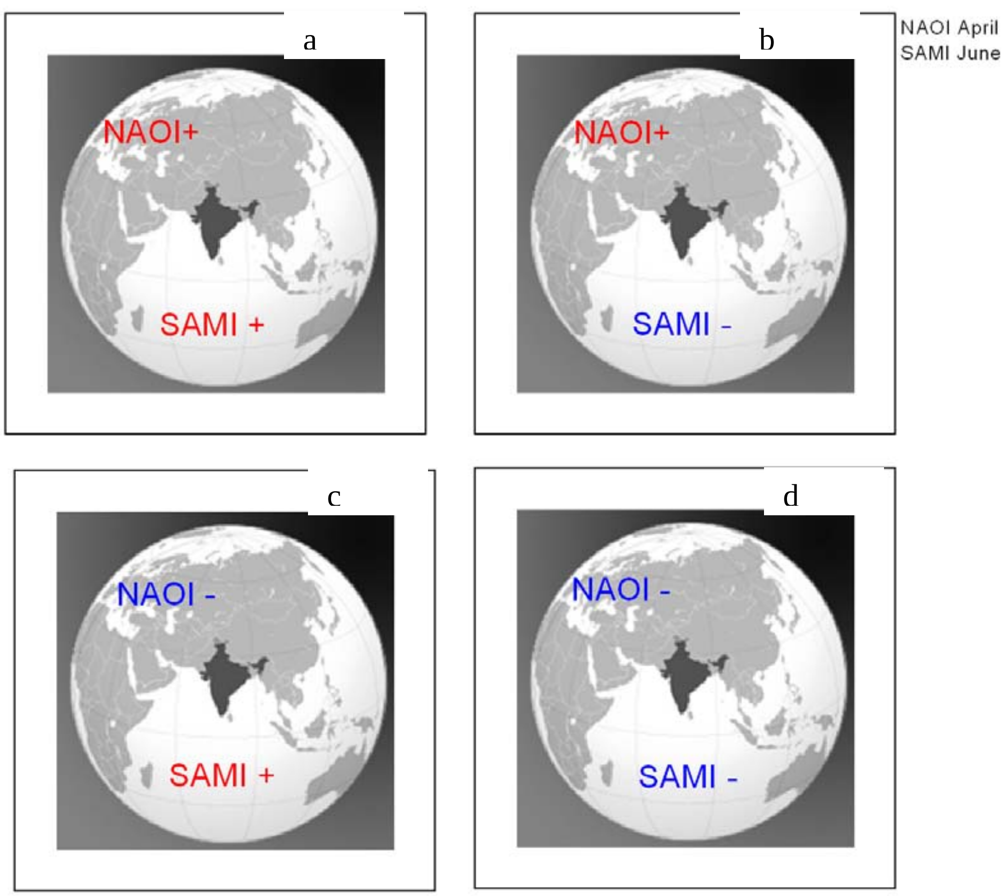

Figure 7.2. Schematic representation of the four indices (a) SAMI^{+}NAOI^{+} (b) SAMI^{-} NAOI^{+}, (c) SAMI^{+} NAOI^{-}, (d) SAMI^{-} NAOI^{-}

7.3.1 Association of July-August variability to SAMI during the positive phase of NAO

Average rainfall during July-August corresponding to June SAMI and April NAOI are shown in fig 7.3 (a, b) and their difference is shown in fig 4.3 c.

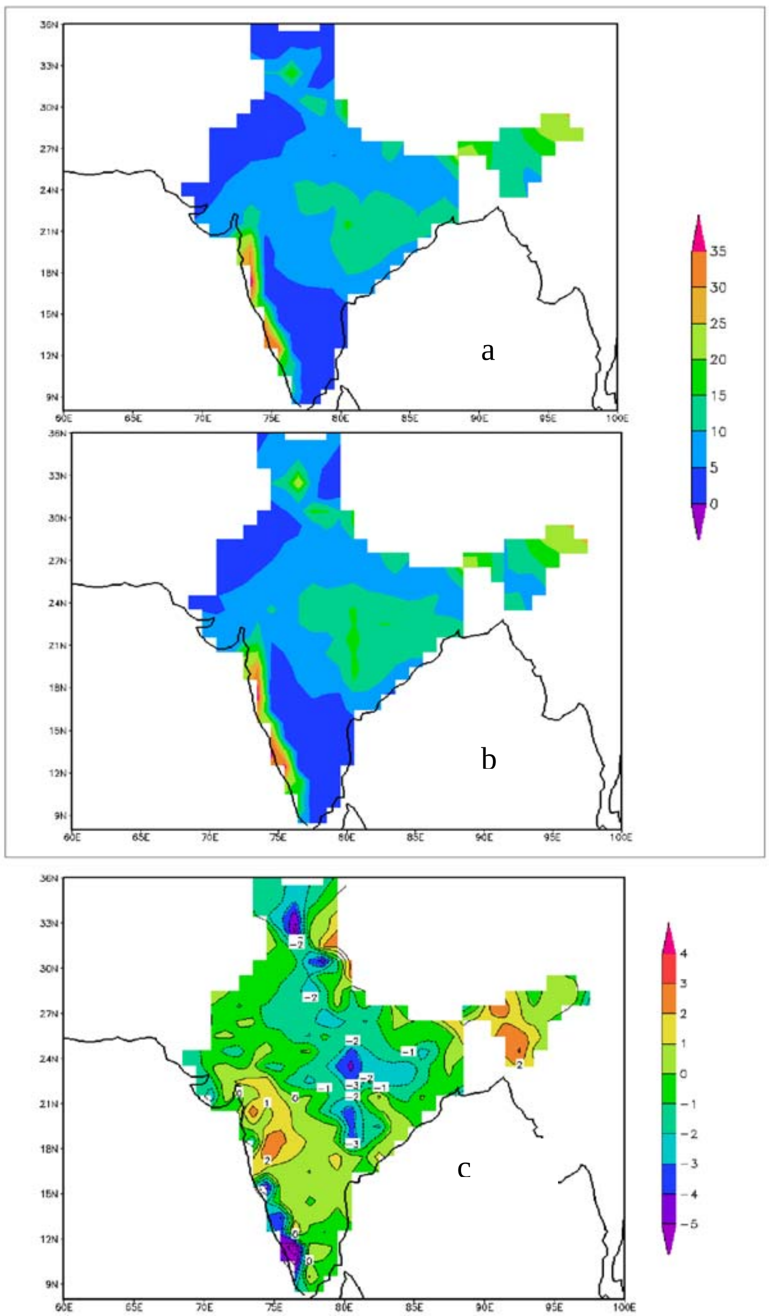

Figure 7.3. Rainfall during July-August a) positive phase SAMI during June and negative phase of April NAOI b) negative phase SAMI during June and negative phase of April NAOI c) figure a minus b

Mean rainfall during July-August corresponding to high Southern Annular Mode and North Atlantic Oscillation Index is shown in figure 4.3 a. During the positive phase of both June SAMI and April NAOI, heavy rainfall of about 30-35 mm/day occurs along Konkan coast. At the same time intense rainfall also occurs along areas of northeast India (15-25 mm/day). To the east of 77^0 E and in between the latitudes 18^0 N and 25^0 N rainfall exceeds 10 mm/day. Central India and south Gujarat region show rainfall in between 5-10 mm/day and it is less than 5mm/day in the southeast and in northwest India.

July-August rainfall corresponding to low SAMI and high NAOI is shown in figure 7.3 b. Along the southwest sector, the rainfall exceed to about 25-30 mm/day during the negative phase of June SAMI and high phase of April NAOI. To the east of 77^0 E and in between 18^0 N and 24^0 N, rainfall is in between 10 - 20 mm/day. In north India and areas of central India shows rainfall of about 5-10 mm/day. While to the northwest, extreme north and southeast India the rainfall is less than 5mm/day.

The rainfall of July-August corresponding to the difference of positive and negative phase of SAMI is analysed during the constant positive phase of NAOI, which is shown in figure 7.3 (c). During the positive phase of both SAMI and NAOI precipitation decreases along Konkan coast. Along the southwest coastal stations up to 17^0 N strong negative anomaly of rainfall is observed. Precipitation decrease of more than 4mm/day occurs to the south of 12^0 N. Precipitation enhances to the north of 17^0 N but in the coastal areas of Gujarat shows insignificant negative anomaly. Precipitation variation of about 1 mm/day is observed in southeast India. While, in the northwest India region precipitation seems to be enhanced slightly. Negative anomalies occur in north, and in areas along central Indian region, where decrease of 3 mm/day is observed in certain areas. At the same time positive anomaly of rainfall occurs in the northeast of India, where rainfall increase of 3 mm/day is noticed.

Rainfall variability is observed over many regions of Indian due to the combined effect of SAMI and NAOI. Further the study extend to understand the SST anomalies due to the simultaneous effect of SAMI and NAOI, the variability in SST can attribute towards anomalous change in the mean flow pattern and also alter the moisture properties over Arabian Sea.

July-August SST corresponding to the difference of positive and negative phase of June Southern Annular Mode Index during the positive phase of April North Atlantic Oscillation Index is shown in figure 7.4. When June SAMI

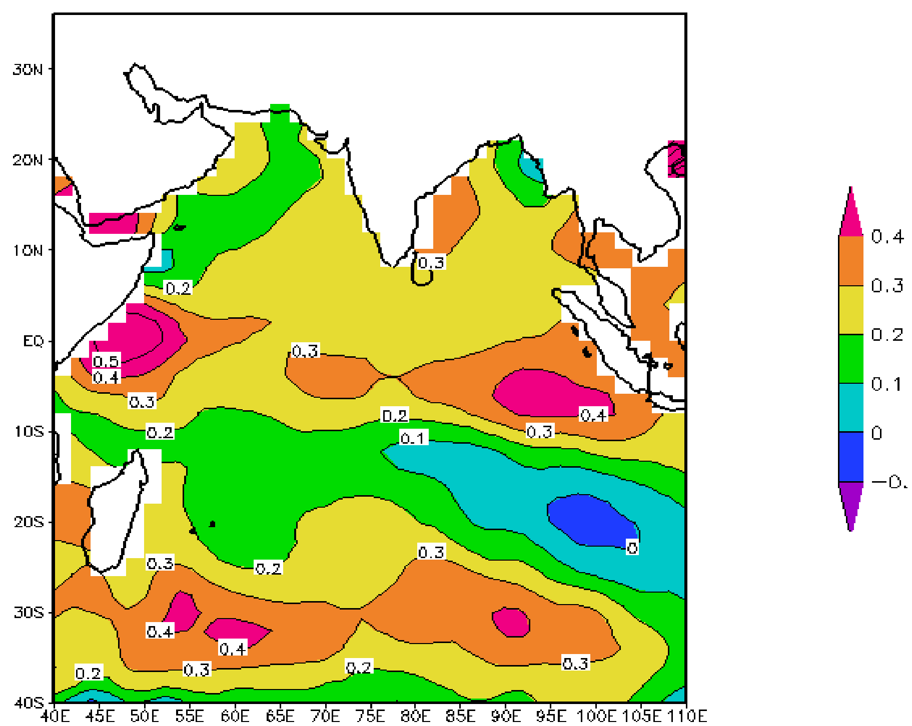

Figure 7.4. July-August SST corresponding to the difference of positive and negative phase of June SAMI during positive phase of April NAOI

and April NAOI indices are active, SST seems to increase in Arabian Sea. While in Bay of Bengal region anomalous change has noticed, increase of 0.3^0 C occurs in the west of Bay and slight decrease in the northeast Bay region. Sea surface temperature also increases in the equatorial region, which is intense in the Somalia coast and in southeastern equatorial Indian Ocean. Near 15^0 S cooling intensifies further to the south towards the eastern longitudes. The magnitude of SST greater than 0.3^0 C occurs around 30^0 S, where two core regions are observed one near 55^0 E and the other along 90^0 E.

Another parameter related to monsoon is the moisture transport during the July-August period. From the rainfall anomaly (figure 7.3.c) it is observed that higher variation in anomalies occurs in south Indian region mainly in the Konkan coast, where maximum rainfall was expected. So this study analysed the zonal moisture transport over this domain.

July-August moisture transport corresponding to the difference of positive and negative phase of June Southern Annular Mode Index during the constant positive phase of April North Atlantic Oscillation Index is shown in figure 7.5. During the positive polarity of both SAMI and NAOI, moisture

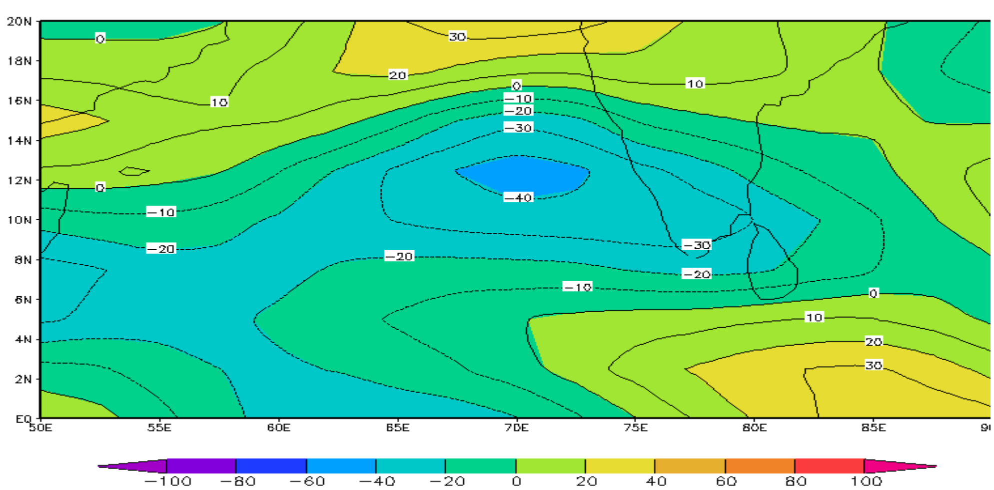

Figure 7.5. July-August moisture transport corresponding to the difference of positive and negative phase of June SAMI during positive phase of April NAOI

transport decreases in between equator and 14^0 N along the west of 75^0 E. The decrease intensifies in Arabian Sea along the belt from Somalian coast to Indian mainland and it extends to the south Bay region. Large decrease of moisture transport is noted along 70^0 E. To the north of 14^0 N slight enhancement of moisture transport occurs, which is relatively high over the west coast of India. Increase in moisture transport is also observed to the south of 4^0 N towards the east of 75^0 E.

7.3.2 Association of July-August to SAMI during the negative phase of NAOI

Average rainfall during July-August corresponding to June SAMI and April NAOI are shown in fig 7.6 (a, b) and their difference is shown in fig 7.6 c. Average rainfall during July-August corresponding to high Southern Annular Mode Index and low North Atlantic Oscillation Index is shown in figure 7.6a. During the positive phase of SAMI and negative phase of NAOI precipitation exceeds 25 mm/day in areas along Konkan coast. At the same time northeast and eastern part of central India shows precipitation of more than 10 mm/day. To the extreme northeastern region precipitation is about 20 mm/day. Precipitation greater than 5 mm/day is observed along central India, and less than 5 mm/day occurs along extreme north, northwest and southeast region and also to the extreme north of India.

July-August rainfall corresponding to the negative phase of both Southern Annular Mode Index and North Atlantic Oscillation Index is shown in figure 7 6.b. Heavy rainfall occurs along southwest India (25-35 mm/day) during low

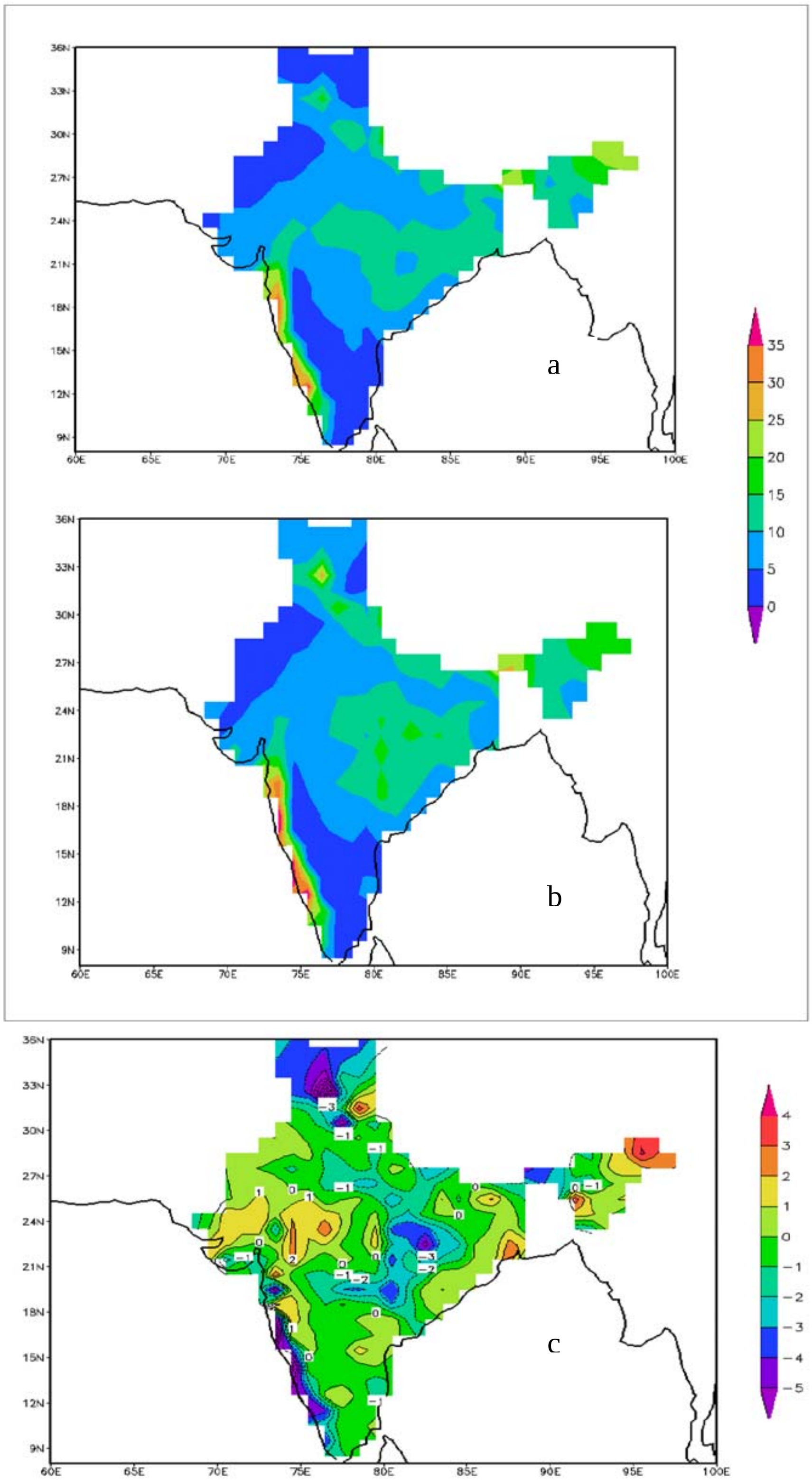

Figure 7.6. Rainfall during July-August a) positive phase SAMI during June and negative phase of April NAOI b) negative phase SAMI during June and negative phase of April NAOI c) figure a minus b

SAMI and high NAOI. This precipitation amount is maximum in between 10^{0}N and 22^{0} N. Rainfall in between 15 – 25 mm/day occurs along northeast Indian region. In a wide area of eastern part of central India precipitation exceeds 10 mm/day. Along the northwest and southeast India precipitation is less than 5mm/day.

Rainfall corresponding to the difference of positive and negative phase of SAMI during the negative phase of NAO is shown in figure 7.6.c. During the positive phase SAMI and negative phase of NAOI precipitation decreases through Konkan coast. Along the southwest coastal stations up to 22^{0} N the rainfall shows strong negative anomaly of 4 mm/day. In northwest India precipitation enhances by 1-2 mm/day. In north and areas along eastern part of central India, negative anomalies of precipitation enhances, where rainfall decrease to 2-3 mm/day in specific areas. In northeast regions the anomalies are quite different, negative along 90^{0} E and positive anomalies to the east of 95^{0} E. Southeast coast of India also shows slight decrease in rainfall anomalies.

Rainfall shows spatial variation in anomalies depending on the phase of SAMI and negative phase of NAOI. The study also analysed the variability of SST and moisture transport associated with this phenomenon. The gradient of SST between southern and northern hemisphere is one of the important parameter affect the monsoon system.

July-August SST variability corresponding to the difference of positive and negative phase of June SAMI during constant negative phase of April NAOI is shown in figure 7.7. Cooling enhances in Arabian Sea region with intense towards the north of 20^0 N during high SAMI and low NAOI, While Bay of Bengal warms slightly and it is intense over east Bay region. Slight warming occurs in the equatorial region where cooling of SST occurs south of equator upto 15^0 S and to the west of 85^0 E. Intense warming of 0.3^0 C occurs along 30^0 S, where two core region of warming are observed, one along 60^0 E and other along 95^0 E.

From the figure 7.6 (c) it is clear that anomalous change occurred in to July-

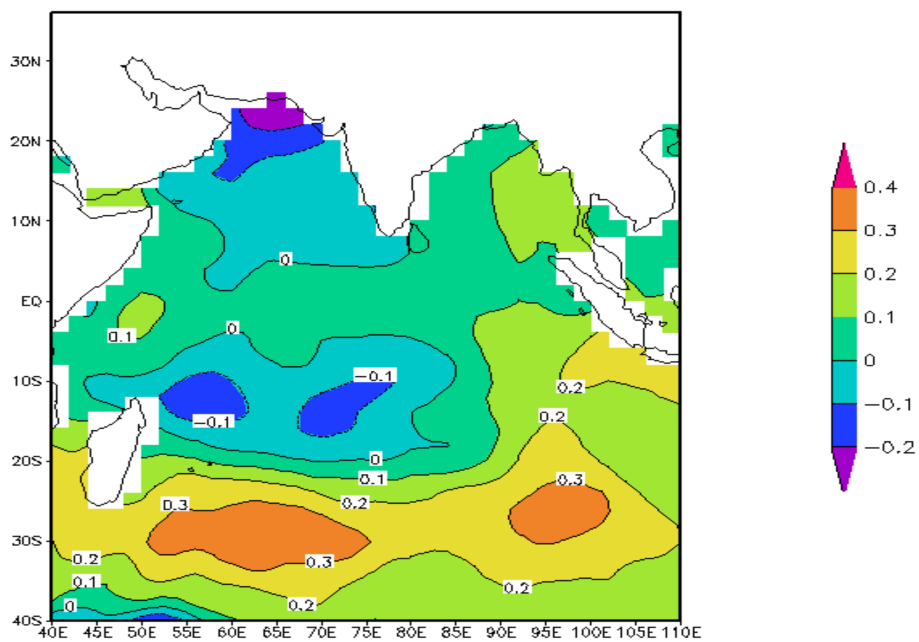

Figure 7.7. July-August SST corresponding to the difference of positive and negative phase of June SAMI during constant negative phase of April NAOI.

August rainfall due to the coupled effect of both modes, which is more pronounced over the southwest coastal Indian region. The variability in the rainfall can be account by the anomalous behaviour of zonal transport of moisture. So the zonal transport of moisture during July-August is also analysed using both indices.

July-August moisture transport corresponding to the difference of positive and negative phase of June SAMI during the constant negative phase of April NAOI is shown in figure 7.8. To the west of 60^0 E moisture transport decreases during high SAMI and low NAOI in between the belt of equator and 20^0 N. Along the east of 65^0 E, moisture transport increases in between equator and 5^0 N which is intense along 90^0 E in the near equatorial region. While to the north of 5^0 N moisture transport decrease upto the belt of 20^0 N. Anomalous decrease in moisture transport are observed in the west of coast of India and in the Somalian coast, where a reduction of 100 kg m/s is observed.

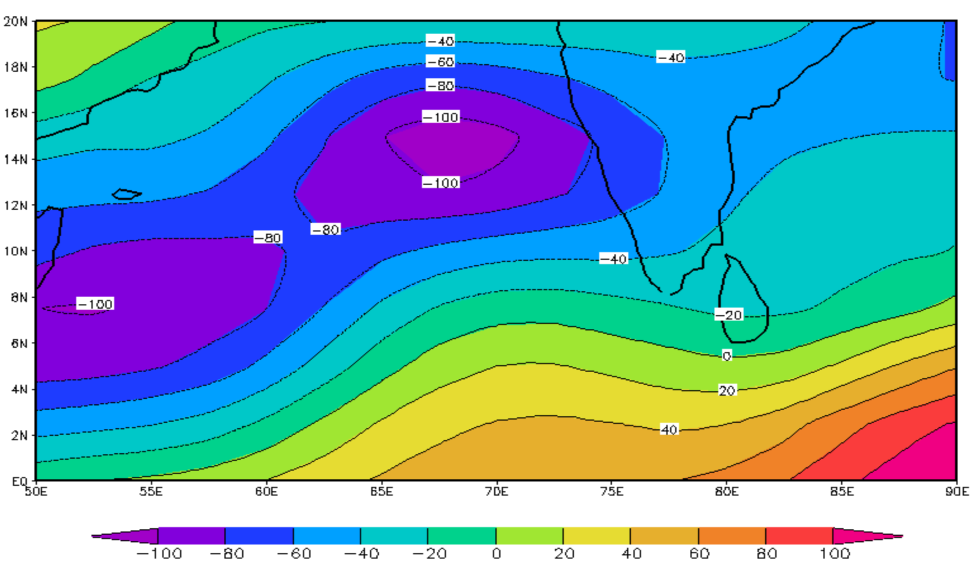

Figure 7.8. July-August zonal moisture transport ($kg\ ms^1$) corresponding to the difference of positive and negative phase of June SAMI during constant negative phase of April NAOI.

Since this study has shown spatial and temporal variability in rainfall, SST and moisture transport during July-August by connecting the two extra tropical phenomenon viz., Southern Annular Mode during June and North Atlantic Oscillation of April. It is important to note the interannual variability of these two indices. As a result present condition of these modes has analysed.

7.3.3 Interannual variation of June SAMI and April NAOI

Time series of SAMI and NAOI are shown in figure 7.9 (a), (b). During June, Southern Annular Mode shows strong deviation in indices from –4.5 to 4.5 (see fig 7.9 a), and the index showed slight positive trend till 1980. During late 1970 the index continued as positive phase for a couple of years showing intense high pressure in mid latitudes and low pressure over high latitudes.

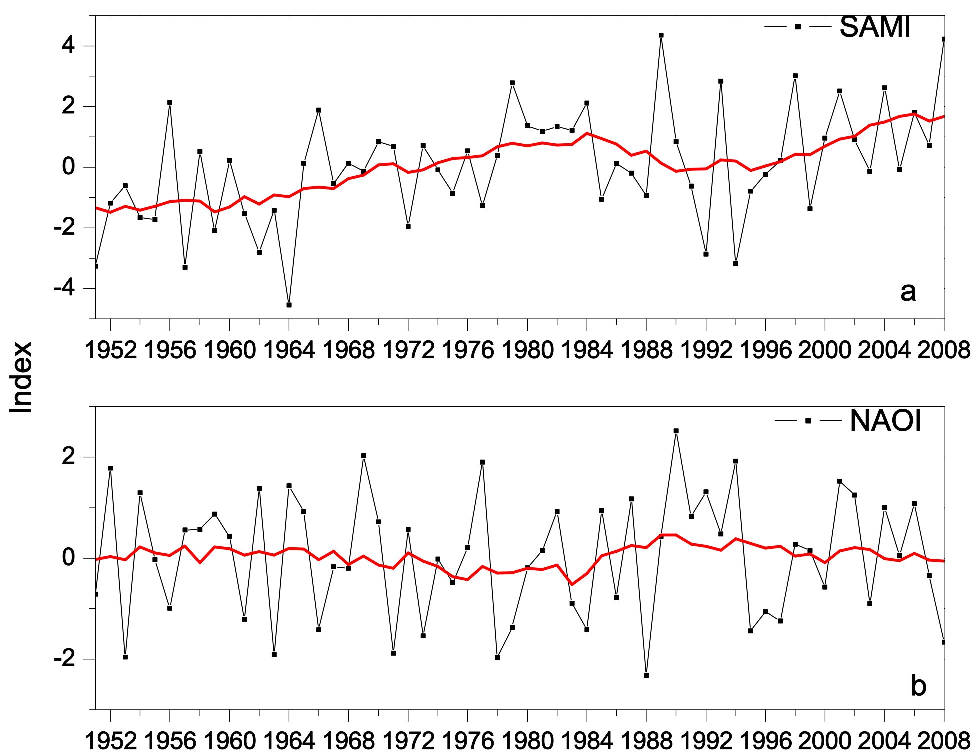

Figure 7.9. Time series of indices during June (dotted lines) along with 10 year moving averages (solid line) for: (a) SAMI during June, (b) NAOI during April

During mid –1980 to early 1990 the SAMI slope was slightly negative, and after that the SAMI is showing a positive trend. At the same time the counterpart oscillation NAOI (see fig 7.9 b) during April the value of index varies in between –2.5 to 2.5. From 1951onwards the index slope was slightly negative till late 1970, after that, sudden jump to the positive phase is observed. In early 1990s the NAO was positive for a couple of years and after that it remained low for three years successively. This shows that the index showed mark variation in 1990's. From mid 1990 onwards alternating phases of index is noted.

7.4 Discussion

Indian summer monsoon variability has been extensively studied in interannual to intraseasonal time scales (Webster et al., 1998; Annamalai et al., 1999). Teleconnection of various parameters with Indian monsoon also well studied (Parthasarathy and Pant, 1985; Verma, 1990). In the wake of breaking relationship between Indian summer monsoon and El Nino (Kumar et al., 1999) it is relevant to study the other parameters influence over monsoon. In this study, Indian monsoon during July-August is related through the simultaneous variation of oscillation in both hemispheres. So the effect of Southern Annular Mode and North Atlantic Oscillation on monsoon has observed through two different perspectives.

Primarily, the variation occurred due to the difference of high and low phase of Southern Annular Mode Index (SAMI) has analysed by keeping the North Atlantic Oscillation Index (NAOI) as positive. During the positive phase of both SAMI and NAOI, rainfall seems to decrease in the southwest coastal stations and also in the north and eastern part of central India. Negative departure of rainfall anomaly is intense over the extreme southwest region. But in the north east region rainfall anomaly increases during the positive phase of indices. The variation due to high phase of both indices are also observed in SST and moisture transport. Sea surface temperature shows increase of temperature over most of the regions. Arabian Sea exhibits slight increase where east of Bay presents an increase of 0.3^0 C. Large areas of increase is observed in SST along 30^0 S. This increase in temperature around 30^0 S is not a good sign for the monsoon**.** On the other hand, moisture transport also shows decrease in Arabian Sea and Indian mainland, which can also adversely affect the amount of precipitation observed in south India. The rainfall anomaly has shown decrease in the southwest coastal regions.

Using other criteria i.e., the difference of high and low phase of Southern Annular Mode Index by keeping North Atlantic Oscillation Index as negative is used to understand the monsoon variability. It is noticed that negative anomaly of rainfall occurs in the southwest, north and eastern part of central India during positive phase of SAMI and negative phase of NAOI. Anomalous change has also observed to the northeast region. Seas surface temperature variation also shows anomalous increase in the southern tropics while North Indian Ocean showed slight cooling with intense over Arabian Sea. This gradient developed in the oceanic region may not be conducive for good rainfall. To the south of 20^0 N anomalous decrease in moisture transport is observed where the magnitude of reduction are quite high in the Arabian Sea showing that zonal moisture transport has weakened during monsoon which will adversely affect the precipitation over south Indian region.

If the two effect are compared it can be understood that the simultaneous effect has difference in magnitude and in spatial distribution of rainfall anomalies (see fig 3c and 6c). During the positive SAMI and negative NAOI the intensity as well as spatial spread of the negative anomaly in the southwest region is more compared to the high phases of both SAMI and NAOI. Negative anomaly is also intense over all regions of India during positive SAMI and negative NAOI. In northeast, the difference between them is well evident, where positive anomaly is observed along 85^0 E in positive phase of SAMI and NAOI. But for the positive SAMI and negative phase of NAOI shows negative anomaly along 85^0 E rainfall while positive anomaly is observed in the extreme northeast. The observed rainfall anomaly over the southwest and northeast due to the coupling effect shows a condition similar to break spells. During break spells, decreased rainfall over western regions of India and increased rainfall over eastern India is observed (Singh et al., 1992; Krishnamurthy and Shukla, 2000).

Sea surface temperature also shows variability in Arabian Sea and Bay of Bengal for the two criteria considered in the study. During high phase of SAMI and low phase of NAOI, Arabian Sea shows cooling, which is not observed in other condition. At the same time southern hemisphere near 30^0 S in the southern hemisphere intense warming is noticed in both cases. Moreover, the gradient in SST shows strong and reverse temperature gradient, which is observed during weak monsoon years. Shukla (1975) has noted that cooling of western Arabian Sea will not help in developing good monsoon. Anomalous cooling of SST in the southeastern equatorial Indian

Ocean and anomalous warming of SST in the western equatorial Indian Ocean results in Indian Ocean Dipole (Saji et al., 1999). It seems that this phenomenon is notably affected by the interaction of these two modes. It seems that, positive phase of both SAMI and NAOI results may favour positive IOD while the positive SAMI and negative NAOI help in developing negative IOD.

One of the important parameter related to monsoon is the moisture transport that shows decrease in Arabian Sea to Bay region. But the interesting thing is that, the decrease in moisture transport is large during positive SAMI and negative NAOI than positive phase of these indices. The reduction of moisture transport mainly occurs during the break spell of rainfall (Ramesh et al., 2009). Another noticeable thing is that both criteria shows increase in moisture transport along equatorial eastern Indian Ocean.

It can infer that these modes are influencing the monsoon parameter and thereby precipitation. The positive trend of SAMI is not conducive for good monsoon in India during July-August. Moreover the rainfall anomalies observed during SAM is affected when the NAO phases are introduced. The modulation of SAM-Monsoon relation due to NAO is well evident in SST and moisture transport. In recent decades the SAMI is in the positive phase and the NAOI is not showing any significant trend, but the model predicts a positive trend for both modes and to be continued due to green house changes (Gong and Wang, 1999; Kidson 1999; Marshall, 2003; Miller et al., 2006). One of the interesting thing is that the climate shift occurred during late 1970 has played a vital role in establishing new relation which can be understood by the altering relationship of Indian monsoon and ENSO (Mokhov et al., 2011) and with the TBO and IOD connectivity which is described earlier. The weakening of low level jet stream during the second half of the 20th century also has affected Indian monsoon. (Joseph and Simon, 2005).

The study provides a possibility of prediction skill of July-August rainfall using June Southern Annular Mode Index and April North Atlantic Oscillation Index, the months in which monsoon rainfall has strongest impact on India and its economy. This will help in understanding the climatic anomalies associated with monsoon. Comprehensive studies on the teleconnection between the extratropics and tropics are really needed to understand the physical mechanism and thereby improving the predictable capability of Indian summer monsoon system.

7.5 Conclusion

Monsoon over India shows strong indication of linkage to June Southern Annular Mode Index. The relationship between SAMI and Indian monsoon is different when the counterpart oscillation North Atlantic Oscillation Index of April has considered. The impact of Southern modulation on Indian monsoon is analysed separately during positive and negative phase of NAOI. It is noticed that the positive phase of SAMI and negative NAOI results high variation in monsoon parameters than during the positive phase of SAMI and NAOI. The precipitation shows strong negative anomaly in southwest, north and eastern part of central India, while northeast region shows positive anomaly. The negative anomaly over southwest region is intense and spatially spread during SAMI is positive and NAOI is negative. During this phase decrease in moisture transport is also intense in the southern India and nearby regions, the SST also shows weak gradient to develop the monsoon system. Moreover SST shows slight variation during the two thresholds in Arabian Sea and Bay of Bengal region and also along the areas of IOD. The study indicates a probable predictability factor of July-August rainfall using the simultaneous effect of June SAMI and April NAOI. Further the results obtained here will help to investigate the depth of the relationship of monsoon with extratropics.

Chapter 8

Summary and Conclusion

The southern hemisphere climate has shown anomalous variation in temperature properties during the second half of the 20th century. In which the great climate shift of 1970 has varied the temperature structure over the southern hemisphere. This is well evident in the temperature in different vertical levels from troposphere to stratosphere. Through Fourier analysis it has been observed that the low periodicities oscillations seems to be dominate in the tropical region and high periodicities in mid-latitude to polar region.

The magnitude of temperature variations seems to be intense in the stratospheric levels, which is strong in the polar stratosphere. Phase change in temperature anomalies has been well evident in between 30 hPa and 1 hPa level. One of the significant variations noted after the climate shift is the cooling at 5 hPa level and warming at 1 hPa level. The western and eastern hemispheres also change its temperature properties after the climate shift. Mid – troposphere analysis through Radiosonde observation detected the 1970 climate shift in both tropical and polar region, while the mid-latitude showed the shift only in 1989. Transition of temperature anomalies from cool to warm has been observed in the *in situ* observation at 500 hPa level. It indicates that the temperature in the southern hemisphere might have attributed due to anthropogenic and natural change.

Another noticeable feature observed is the anomalous variation in upper stratosphere over southern hemisphere. This can be observed by the trend from 1989 – 2010 period. Stratosphere level at 5 hPa has been cooling intensively during winter while warming occurs at the stratopause level at 1 hPa level. The cooling trend has been intense over the polar stratosphere while the warming trend is high over the tropical region. Such strong trends are seen in the southern hemisphere during autumn and winter seasons. During spring the austral hemisphere shows distinct characteristics by polar warming which can be attributed due to the distribution of ozone.

Another interesting feature in the southern stratosphere is the extremity of cooling and warming. It can be noticed that a sudden increase in the number of extreme days has occurred after 1998 to till 2010. In specific levels an abrupt variation in the stratospheric temperature is observed during 1998 to 2001 that is also evident in the El Nino indices. The tropical and mid-latitude stratosphere regions seem to be disturbed during the low and high phases of El Nino indices. These variations are stronger during winter, spring and summer. The tropospheric perturbation and stratosphere temperature at 5 hPa shows strong in-phase relationship on interannual timescales from 1998 onwards. The effect of ENSO to the stratospheric temperature may possibly due to the effect of wave activities that generated in the lower troposphere.

The perturbation of southern extra tropical region seems to influence strongly to the tropical circulations. These variations are accounted by the dominant mode of variability of southern hemisphere, known as Southern Annular Mode. During the positive and negative phase of SAM, the tropical circulation like Hadley and Walker circulation get varied. The Walker circulation get intensified during the negative phase of SAM. The Hadley circulation of winter period strengthens during the positive phase of SAM. But, during the boreal summer, the variations in Hadley cell have observed only in regional circulation. The climatology of velocity potential obtained was slightly different form the earlier finding, this change aroused due to the period of consideration. The period observed in this case is taken only after the climate shift, as a result the present status of circulation gets revealed. Further strengthening of winter Hadley circulation has been observed till 2010.

The influence of southern extratropics on Indian summer monsoon circulation is another topic of interest. The positive phase of June SAM accelerates the low level flow at 850 hPa during June. Eventhough the precipitation is less in the central to north Indian region during June, it has been observed that the positive SAM enhance precipitation in this region. During positive phase of SAM, ascending motion increases together with moisture content over the specific region may slightly reinforce the rainfall. Though the positive June SAM accelerates the monsoon, but it adversely affects the July-August rainfall.

The monsoon rainfall during July-August is more important because the rainfall spread all over India during this period and maximum precipitation do occur in the Indian sub-continent. The negative rainfall anomalies are stronger in the southwest coast, central and north Indian region while precipitation enhances over the northeast region. During the positive phase of SAM the low level flow and vertical velocity represent a character of weak monsoon. The anomalous low level flow occurs in the south Indian region while large scale descending motion enhances in the same region during the positive phase of SAM. Sea surface temperature indicates cooling in Arabian Sea, which will adversely affect the evaporation and their by moisture properties. The moisture transport also seems to reduce in the southern India on the high phase of SAM. The positive shift of SAM anomalies has been observed in recent years, and the model also indicates further strengthening of SAM which is not a good sign for summer monsoon over India.

The southern annular mode index of June seems to influence the monsoon during July-August, this will help in understanding the July-August monsoon flow pattern before one month, indicating a predictive nature of Indian summer monsoon through the southern extra tropical variation. The character of monsoon precipitation is not seems to a sole dependent factor of SAM, but in the light of weakening relationship of El Nino- Monsoon the SAM may provide valuable information regarding the monsoon pattern.

The observed SAM-Monsoon connection seems to alter by the phases of North Atlantic Oscillation. The April NAO index can alter the relationship of SAM-Monsoon. The negative phase of NAO and positive phase of SAM produce strong negative rainfall anomalies than observed during the positive phase of both NAO and SAM. Moisture transport over south India also seems to highly reduced during the negative phase of NAO and positive phase of SAM. Sea surface temperature also favours a weak monsoon condition during these phases. It can infer that the negative NAO and positive SAM will result a weak monsoon than the positive phase of both indices.

8.1 Scope of future work

The present study illustrates the variability in southern hemisphere and its effect on tropical circulation. Detailed studies are to be carried out to understand the physical mechanism behind the sudden and prolonged variation of southern atmosphere and its influence on northern hemisphere features. The observed sudden variation of stratosphere in recent decades has to investigate further to understand whether these variations are a result of natural or by the overloading of greenhouse gases to atmosphere. The interannual relationship of ENSO–Stratosphere is an interesting topic to know how this relationship woks on intraseasonal time scale. The variations in Hadley and Walker circulations are known possibly by the trend of extra tropical modes, and this will provide information regarding the tropical circulation pattern, and through models the area of precipitation and drought can be predicted. The effect of SAM over Indian summer monsoon has to study in detail to get an inference over the predictive potential of Indian summer monsoon on intraseasonal timescale.

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List of Abbreviations

ASM	Asian Summer Monsoon
CPC	Climate Prediction Centre
ECMWF	European Centre for Medium Range Weather Forecasting
ENSO	El Nino Southern Oscillation
FFT	Fast Fourier Transform
IAV	Inter annual variability
IOD	Indian Ocean Dipole
IPCC	Intergovernmental panel on Climate Change
ISMR	Indian Summer Monsoon
ISV	Intraseasonal Variability
NAOI	North Atlantic Oscillation
NCAR	National Centre for Atmospheric Research
NCEP	National Centre for Environmental Prediction
SAMI	Southern Annular Mode Index
SST	Sea Surface Temperature
SSW	Sudden Stratospheric Warming
TBO	Tropical Biennial Oscillation
QBO	Quasi-Biennial Oscillation
WMO	World Meteorological Organisation