SEASONAL AND MONTHLY RAINFALL VARIABILITY OVER KERALA IN A WARMING CLIMATE

by

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THESIS

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DECLARATION

I, hereby declare that this thesis entitled "**SEASONAL AND MONTHLY RAINFALL VARIABILITY OVER KERALA IN A WARMING CLIMATE**" is a bonafide record of research work done by me during the course of research and the thesis has not previously formed the basis for the award to me of any degree, diploma, associate ship, fellowship or other similar title, of any other university or society.

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Suja Mary Sam

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ABBREVIATION

%	Percentage
mm	Millimeter
m	Meter
cm	Centimeter
mm/day	Millimeter per day
m/s or ms^{-1}	Meter per second
km	Kilometer
Sq.km	Square Kilometer
ha	Hecta
mb	Millibar
hPa	Hecta Pascal
kt	Knot
m/s	Meter per second
^{0} C	Degree Celsius
i.e.,	That is
etc	Et cetera
/	or
~	Approximately
+	Positive
-	Negative
Lat.	Latitude

Long.	Longitude
M	Mean
CV	Coefficient of Variance
σ / SD	Standard Deviation
N	North
S	South
E	East
NW	North West
NE	North East
SW	South West
WG	Western Ghats
ET	Evapotranspiration
GHG	Greenhouse Gas
ENSO	El Nino Southern Oscillation
AMO	Atlantic Multidecadal Oscillation
AZM	Atlantic Zonal Mode
PDO	Pacific Decadal Oscillation
AZO	Atlantic Zonal Oscillation
NIO	Northern Indian Ocean
DMI	Dipole Mode Index
NOAA	National Oceanic and Atmospheric Administration
IMD	Indian Meteorological Department
ISM	Indian Summer Monsoon

AMS	Asian Monsoon System
MOK	Monsoon Onset over Kerala
IPCC	Intergovernmental Panel for Climate Change
SST	Sea Surface Temperature
TEJ	Tropical Easterly Jet
ASM	Asian Summer Monsoon
ISMR	Indian Summer Monsoon Rainfall
SWM	South West Monsoon
KSWM	Kerala South West Monsoon
PMR	Post-Monsoon Rainfall
ITCZ	Inter Tropical Convergence Zone
SO	Southern Oscillation
IO	Indian Ocean
pIOD	Positive Indian Ocean Dipole
nIOD	Negative Indian Ocean Dipole
EIO	Equatorial Indian Ocean
BOB	Bay of Bengal
AS	Arabian Sea
KSMR	Kerala Summer Monsoon Rainfall
ASMR	Asian Summer Monsoon Rainfall
QBO	Quasi Biennial Oscillation
LLJ	Low Level Jet Stream
IOD	Indian Ocean Dipole

TT	Tropospheric Temperature
EEIO	Eastern Equatorial Indian Ocean
OLR	Outgoing Long-wave Radiation
SST	Sea Surface Temperature
SLP	Sea Level Pressure
JJAS	June, July, August and September
LULC	Land Use and Land Cover
GPCP	Global Precipitation Climatology Project
ERSST	Extended Reconstructed Sea Surface Temperature
ICOADS	International Comprehensive Ocean-Atmosphere Data Set
VIMF	Vertical Integrated Moisture Flux
q	Specific Humidity
u and v	x and y component of wing
p	Pressure
psurf	Surface pressure
ptop	Pressure at the top of the atmospheric layer
g	Acceleration due to gravity
pr_wtr	Precipitable water
MOK	Monsoon Onset on Kerala
JJAS	June, July, August and September
SPSS	Statistical Package for the Social Service

INTRODUCTION

Chapter 1

INTRODUCTION

Indian economy faces serious challenges due to the existence of inter-annual variability of summer monsoon and it is an important factor that oversee the performance of different sectors such as agriculture, hydroelectric power and associated industries throughout India as it provides annual rain-potential of 70-80% (Mooley *et al.,* 1983) of the total rainfall. Hence it is necessary to understand and predict the fate of this variable for better planning. Monsoon rainfall unpredictability has a significant impact on agricultural productivity, which in turn has an impact on the Indian economy. Although the inter-annual variation of ISMR is just 10% of the mean, it has a significant impact on food grain production (Gadgil, 2003).

The variability in monsoon makes the heavily populated India more vulnerable to natural disasters like flood and drought. In recent decades, monsoon variability has increased (Singh *et al.,* 2014; Dash *et al.,* 2011; Vinnarasi and Dhanya, 2016) with gradual decrease in monsoon circulation and rainfall but the incidence of extreme events are increasing (Kulkarni,2012; Roxy *et al.,* 2015). For example several regions of India experience calamitous flood and drought events due to multi- day deluges which includes 2002 ,2004,2009,2012 ,2014 (drought years), 2005 Mumbai flood, 2015 Chennai flood, 2018 Kerala flood, 2019 eastern India flood and 2020 Telangana & Assam flood.

The El Nino Southern Oscillation (ENSO), Indian Ocean Dipole (IOD), Atlantic Multidecadal Oscillation (AMO), Atlantic Zonal Mode (AZM), and Pacific Decadal Oscillation (PDO) were linked to ISMR variability (Nair *et al.,* 2018; Sabeerali *et al.,* 2019).The decrease in ISMR are linked to weakening of monsoon circulation with combined influence of Indian Ocean (IO) warming (Roxy *et al.*, 2015; Mishra *et al.,*2012), increased frequency and intensity the El Nino/Southern Oscillation (ENSO) anomalies exist in equatorial Indian Ocean (Kumar,2006; Roxy *et al.,* 2014), elevated air pollution (Krishnan *et al.,* 2016), change in land use patterns (Paul *et al.,* 2016) and climatic anomalies in Atlantic Ocean (Yadav *et al*., 2018). Various researches suggest that increase in extreme events are associated with increase low level westerlies variability over Arabian Sea (AS) (Roxy, 2017), rapid warming of equatorial IO (Ajayamohan, 2008; Guhathakurta *et al.,* 2017), increased number of low pressure system (Sorland and Sorteberg, 2016), local surface warming along with humidity rise over Indian sub-continent (Roxy, 2017) etc.

Future forecasts show an increase in precipitation over South Asia due to warmer climate and weakening monsoon circulation, according to the IPCC (2007) study. Monsoon intensification is caused by an increase in atmospheric moisture transport (Christensen *et al*., 2007) as well as an increase in the land-sea temperature differential (Kulkarni *et al.,* 2010). The existence of a high temperature difference between the equatorial Indian Ocean and Northwest India is favorable to substantial monsoon activity over India (Kulkarni *et al.,* 2010). According to Roxy *et al.* (2014), the western tropical Indian Ocean is already warming at a higher rate than any other region of the tropical Indian Ocean for more than a century, altering the zonal SST gradient across the tropical warm pool region and affecting Asian monsoon circulation and rainfall. Due to the periodic occurrence of El-Nino, it act as a vent to exchange heat stored from the ocean to the atmosphere through a modified Walker circulation (Rhein *et al.,* 2014). The feedback from warmer IO could explain the recent cold conditions across the eastern Pacific Ocean. The warming trend over the Indian and Atlantic oceans leads to La-Nina like conditions over the Pacific (Kucharski *et al.,* 2011). The pole-ward shift in LLJ in response to the widening of the tropical belt under global warming causes a reduction in rainfall over the extreme southern peninsular Indian region (Sandeep *et al.,* 2014), whereas the weakening of vertical velocities caused by upper tropospheric warming causes a reduction in precipitation along India's southwest coast (the Western Ghats), which is particularly visible during July (Rajendran *et. al,* 2012). The Tropical Easterly Jet (TEJ), a notable upper-level circulation feature during the monsoon season, is also diminishing due to the warming of the middle to high troposphere over the equatorial Indian Ocean region (Sathyamoorthy, 2005; Abish *et al*., 2013).

The most essential factor to consider is ENSO-related modulation of ISMR variability. Because the ENSO index is negatively connected with regional scale summer monsoon rainfall and exhibits significant variability due to regionally diverse feedbacks, it is vital to investigate its impact on the summer monsoon in greater depth under a global warming scenario. ISMR has major contribution towards the global atmospheric circulation as the Indian summer monsoon system dominates the northern summer Hadley circulation. In India, numerous research have been undertaken to determine the variability and patterns of monsoon rainfall. Improving our understanding towards monsoon inter-annual variability and associated extremes may help us to reduce the infrastructure damages and loss of life in the future. The variability of rainfall in Kerala is taken into account in this study. It is a long, narrow coastal state in peninsular India, located between latitudes 8^{o} 15'N and 12^{o} 50'N and longitudes 74^{o} 50'E and 77^{o} 30 E. It is known as the “Gateway of the Summer Monsoon”. It is a narrow strip of land that runs between the Arabian Sea and the Western Ghats on the west and the Western Ghats on the east. The region's topography ranges from steep terrain to more or less flat terrain. Therefore, the state is divided in three natural zones viz., the eastern high lands, the hilly midlands and western lowlands. Recent trends of uncertainty in monsoon variability and distribution occurs as a result of man-made interventions in the geographical and topography features which has drastic influence over atmospheric circulation. Kerala receives 300cm of annual rainfall, which is three times that of Tamil Nadu and two times that of Karnataka. The southwest monsoon (June-September), northwest monsoon (October–November), pre monsoon (March–May) with high thunder storm activities, and winter months (December–February) with minimal clouding and rainfall are the rainfall seasons in Kerala. A detailed study about the monsoon onset and rainfall are carried out by (Ananthakrishnan and Soman, 1988).Recent studies shows that the south west monsoon (SWM) rainfall over Kerala is dwindling while the post-monsoon rainfall (PMR) shows increasing trend (Abhilash *et al.,* 2018). From 1871 – 2011, the SWM rainfall shows a decrease of 10.9mm whereas there is an increase of 7.5mm in PMR (Mini *et al.,* 2016). The northern and high range zone regions of Kerala normally receives adequate amount rainfall during SWM seasons whereas southern zone regions experience relatively less rainfall (Ajithkumar and Sreekala, 2015).Therefore it is necessary to understand the physical process and potential predictors that controls the extreme seasonal rainfall in Kerala. The objectives of my study includes:

1. To understand the monthly and seasonal variability exist in Kerala monsoon rainfall
2. To evaluate the role of Indian and Pacific Ocean Sea surface temperature (SST) on modulating the Monsoon.

REVIEW OF LITERATURE

CHAPTER 2

REVIEW OF LITERATURE

The Asian monsoon system (AMS) is a "large-scale coupled ocean–atmosphere phenomenon" that affects the Indian subcontinent and other adjacent locations with distinctive seasonal precipitation and circulation patterns (Webster *et al.,* 1998). Being the key element of global climate, Asian monsoon system have influence over the tropics and subtropics of the Eastern Hemispheres and also affects the atmospheric circulation over extra tropics (Webster *et al*., 1998). The Asian summer monsoon displays a enunciate variability on different time ranges ranging from intra-seasonal to inter-annual and inter-decadal (Chang and Krishnamurti, 1987; Fein and Stephens, 1987; Webster *et al*., 1998). Although the total precipitation over the Asian monsoon region increases significantly, the low-level tropical circulation weakens due to increased static stability (Wang *et al*., 2014), reduction in meridional thermal gradient (Ueda *et al*., 2006) and intensity of the Walker circulation decreases due to climate warms (Vecchi and Soden, 2007; Haarsma and Selten, 2012).

The Indian summer monsoon rainfall affects a large portion of the Indian subcontinent from June to September (ISMR). It is one of the most significant worldwide general circulation phenomena that has an impact on global weather and climate (Ghosh *et al.,* 2009). Because it insulates warm-moist air over South Asia from cold-dry air from extra-tropics, the summer monsoon in India is far more dominant than other monsoon systems due to its terrain (Boos and Kuang, 2010).

ISMR is accountable for 70-90 percent of the total annual precipitation in the country (Shukla and Haung, 2016). In the Indian subcontinent, seasonal (JJAS) rainfall plays a critical role in agricultural productivity, industrial growth, water management, and other areas, and any inter-annual and inter-seasonal variability in ISMR has a significant impact on the economy. ISMR's inter-annual variability is just 10% of the mean, yet it has a significant impact on agricultural production

(Gadgil, 2003). The inter-annual variability of ISMR is used to predict the mean seasonal rainfall (Goswami *et al.,* 2006; Pillai and Chowdary, 2016).

The ISMR possess temporal and spatial variation (Goswami, 2005) which leads to large scale floods and drought over different parts of the country. Yadav (2016,2017a, 2017b) and Preethi *et al.* (2017) revealed a westward shift of the rain band in recent decades, resulting in copious rainfall in west and central India and below-average rainfall in east and northeast India. Forecasts suggest an increase in precipitation and a weakening of the monsoon circulation in the future (Christensen *et al.,* 2007; Meehl *et al.,* 2007b). ISM rainfall shows great spatial variability and the highest rainfall was reported over the western coast of India (owing to orographic effects) and over the head of the Bay of Bengal (BOB) along the monsoon zone (Sikka and Gadgil, 1980).

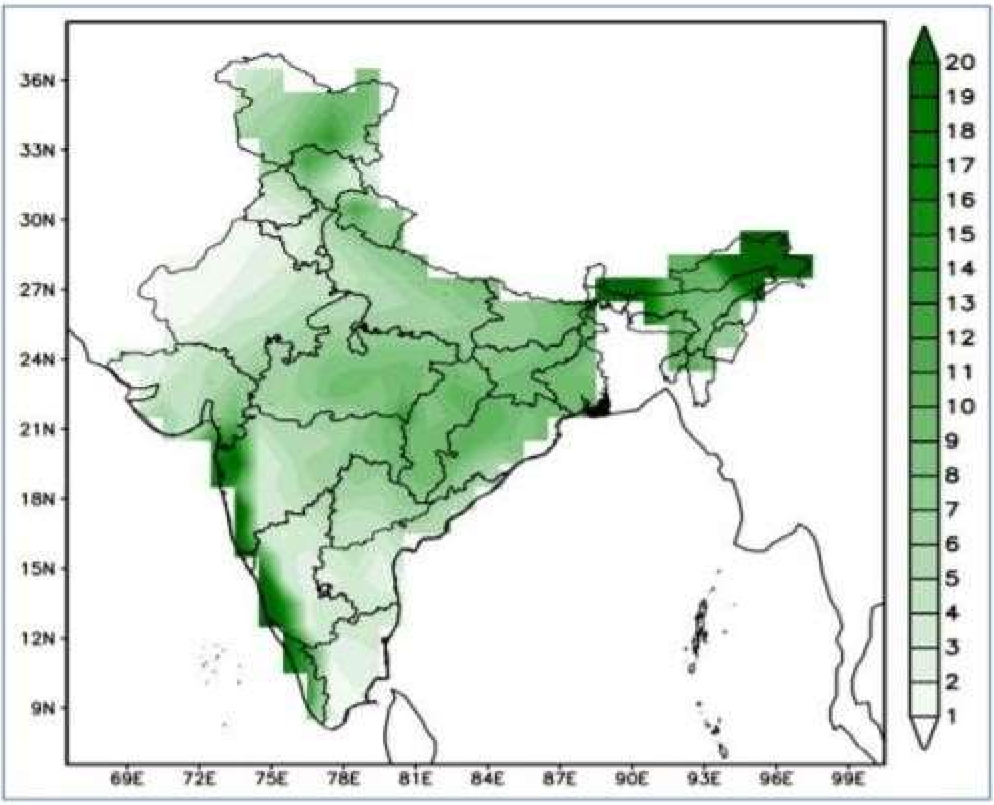

Figure 2.1: Climatological distribution of the Mean summer monsoon (JJAS) rainfall (mm/day) during the 1901-2009 period (Source : Ashok *et*

ISMR's inter-annual variability is not stable over time (Saha *et al.,* 1979), but it is linked to a number of oscillations, including the El Nino Southern Oscillation (ENSO), Indian Ocean Dipole (IOD), Atlantic Multidecadal Oscillation (AMO), Atlantic Zonal Oscillation (AZO), Atlantic Zonal Mode (AZM), and Pacific Decadal Oscillation (PDO) (Nair *et al.,* 2018; Sabeerali *et al.,* 2019; Hrudaya *et al.,* 2020). ENSO and IOD are two of them that have a substantial impact on ISMR (Krishnaswamy *et al.,* 2015).

2.1 Monsoon

Monsoon is distinguished as the seasonal reverse of wind which is accompanied by precipitation, formed as result of differential heating between the land and sea (Maharana, 2019) which is dominant feature of tropical and subtropical climate. The word monsoon is derived from an Arabic word "mausam" meaning season. The Asian-Australian region (Figure 2.2), which encompasses the Indian subcontinents, Southeast Asia, and China, has the most noticeable monsoon climate. Other areas with monsoonal features include the maritime continent, Western Africa, Central America, South-western North America, and South America, which are frequently plagued by severe droughts and floods due to substantial year-to-year fluctuation in monsoon rainfall. The seasonal shift of wind, the humid summer, and the dry winter are all characteristics of the monsoon climate.

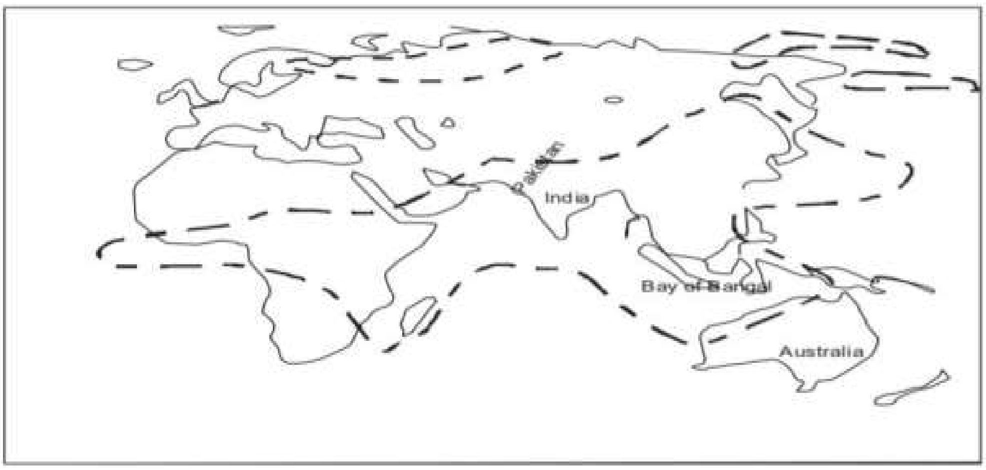

Figure 2.2 Monsoon region countries (Source: Zubair et al., 2013)

During ISM, in the lower troposphere, there is cyclonic and convergence flow, whereas in the higher troposphere, there is anti-cyclonic and divergence flow. Temporal and spatial variations in ISMR are exhibits diurnal to multi-decadal time scales and regional variation. Seasonal reverse of wind, emergence of upper level low-level jet (LLJ), and tropical easterly jet (TEJ) are all linked to ISM strength and variability (Raman, 2009).

2.2 Monsoon Mechanism

Monsoon are large scale seasonal wind system blowing over vast areas of the globe, persistently in the same direction only to be reversed with the change of seasons (Rama Sastry,1983)".The characteristic features for identifying monsoon system includes (Ramage, 1971):

- The prevalent wind direction must change by at least 120^{0} between January and July.
- The average frequency of the wind should exceed 40 per cent
- The mean resultant wind velocity must exceed 3m/s at least one of the months
- Every two years, over 5^{0} degrees latitude/latitude, a cyclone-anticyclone alternation is required. According to this, the monsoon region encompasses sections of Africa, South Asia, and North Australia.

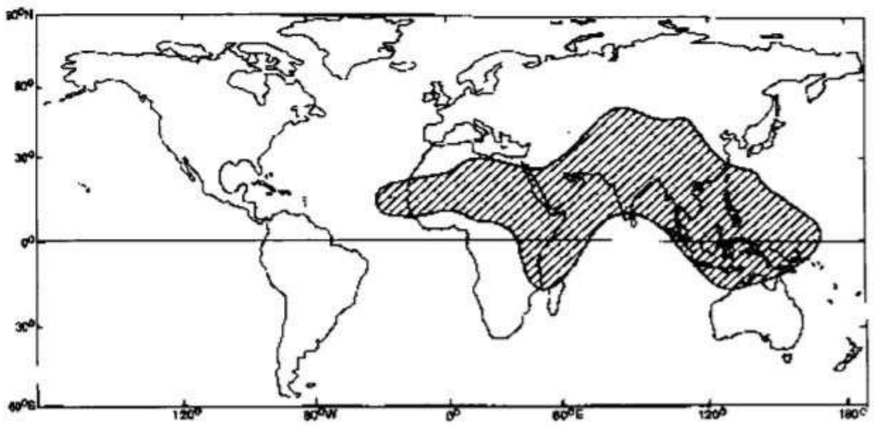

Figure 2.3 Monsoon Circulation (Source: Ramage, 1971)

The monsoon regime is the result of interactions between planetary and regional factors which are present on the surface as well as in the upper troposphere. There are varies hypothesis that explains the origin and mechanism of monsoon.

1. **Classical Theory or Thermal Concepts** (Saroha, 2017; Kumar,2014**)**

The thermal origin of Indian monsoon was explained by Admand Hally in 1686. According to him, monsoon is due to the differential heating and cooling of land and sea. Over the northern hemisphere, the sun is vertical over the Tropic of Cancer during summer season. As the result the Indian land mass is more heated than the neighboring oceans which creates a low pressure system over the Indian subcontinent. A thermal induce pressure formed in the ocean, moves towards the Indian subcontinents results in the onset of southwesterly winds that carries large amount of moisture and cause heavy rainfall over the landmass. On the other hand, the land cools faster than ocean in winter which causes the reversal of pressure gradient towards the sea which trigger the onset of wind from northeast to southwest which carry little amount of moisture.

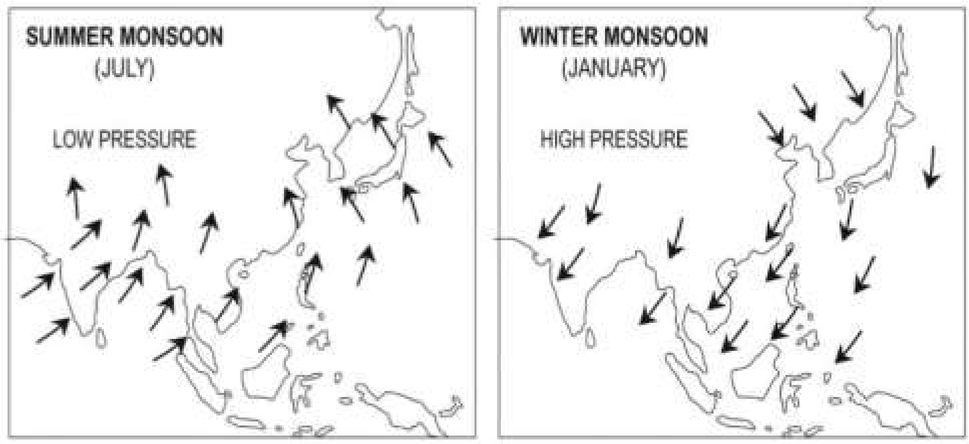

Figure 2.4 Thermal or Classical Theory of Monsoon (Source: Saroha, 2017)

2. **Aerological concept** (Kumar, 2014) **:**

R. Scherhag proposed the aerological concepts of monsoon origin in 1948. He observed that any changes in air temperature above the friction layer is associated with the wind direction changes. According to his view upper air monsoon phenomena is mostly controlled by the annual oscillation of temperature and pressure. The upper tropospheric level lacks complete reversal of zonal monsoon circulation (free troposphere). The thermal monsoon (formed as the result of anticyclones and depression over the Indian subcontinents) on the surface was opposed by the free troposphere monsoon system creates a strong and different type circulation.

3. **Dynamic Concept Or Shifting Of Inter Tropical Convergence Zone (ITCZ)** (Saroha, 2017; Kumar,2014)

This concept is put forward by H. Flohn of German Weather Bureau in 1951. According to him, monsoon is formed as result of shifting of ITCZ. During summer season, the sun is vertical over the Tropic of Cancer which make all the wind and pressure belt to moves towards the north. Along with this, the equatorial westerly of doldrums shift northwards and collectively form into south west monsoon winds. Heavy rainfall is received during this period as the south west wind are present on-shore. During winter season the ITCZ moves towards south and form north-east trade wind. These are off-shore winds that are generally dry and devoid of rains.

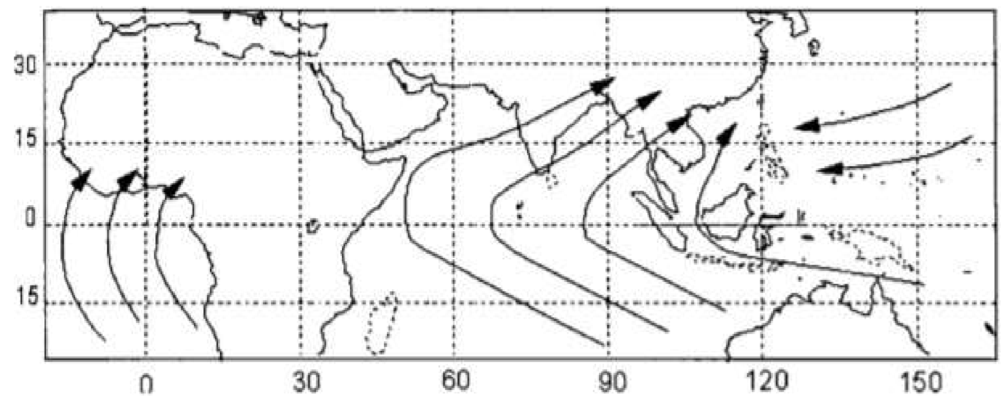

Figure 2.5 (A) Surface winds during northern hemispheric winter monsoon (Source: Webster, 1987)

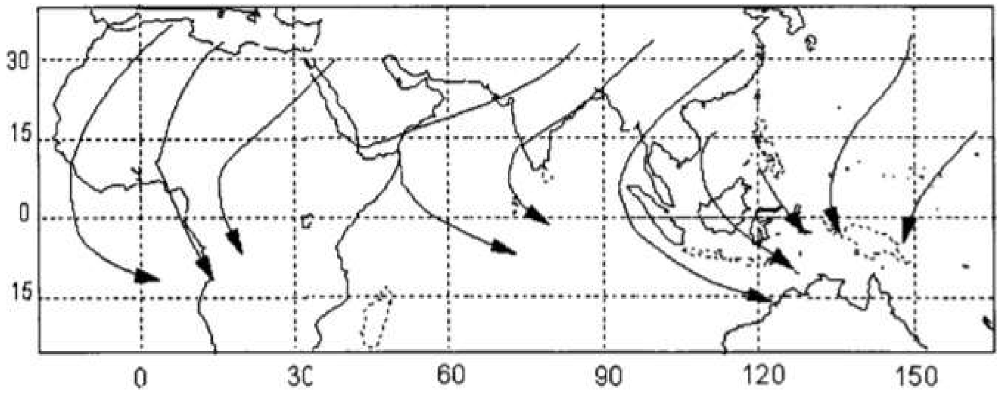

Figure 2.5 (B) Surface winds during northern hemispheric summer monsoon (Source: Webster, 1987)

4. **Upper Air Circulations :**

Influence of upper air circulation in monsoon formation was first expressed by M. T. Yin (1949) and P. Koteswaram (1952). According to them, the onset, withdrawal and intensity of Indian monsoon were influenced by upper atmospheric conditions of Tibetan plateau, position and intensity sub-tropical westerly jet stream and tropical easterly jet streams (TEJ).

During winter season, the position of sub-tropical westerly jet is over the north India and the jet bifurcates into two branches due to the presence of Himalaya and Tibetan Plateau. The northern and southern branches of jet occupies a position over north to Tibetan Plateau and over south of Himalaya. A high pressure (anticyclone) system is formed in upper troposphere which develop towards the south of southern branch of jet stream over Afghanistan and NW Pakistan. Consequently, the wind are forced to descend over the NW part of India, resulting in atmospheric stability and dry conditions. This sub-tropical westerly jet stream help the western disturbance to pass over the Indian continent which leads to snow fall over western Himalaya and rainfall over Great plains.Only the northern branch of the subtropical westerly jet stream exists after the first week of June, and it travels to the north of the Tibetan plateau. This results in the formation of dynamic depression over north western part of Indo-Pakistan. Burst of monsoon takes place when this dynamic depression entrenched over the thermal depression.

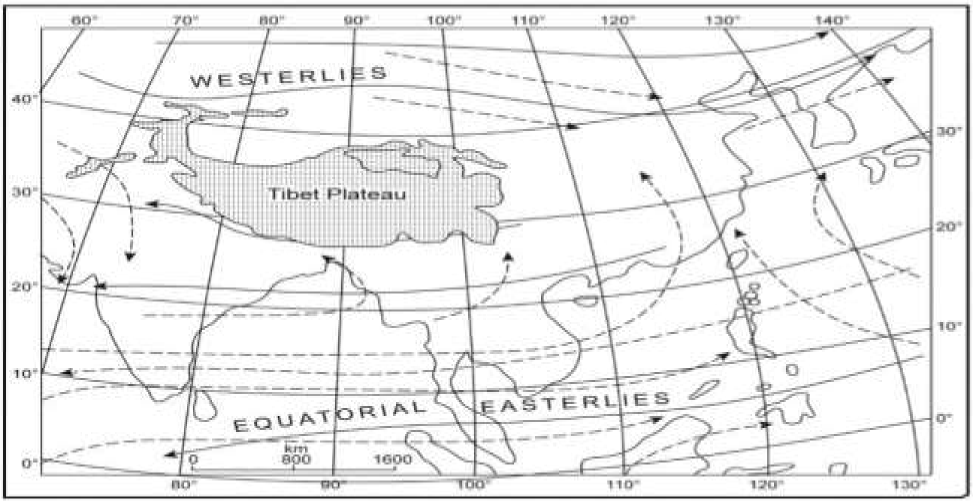

Figure 2.6 Upper air circulation Features and Monsoon

During summer season, Tibetan Plateau (2-3° C greater than the surrounding area) gets heated and acts as a high altitude heat engine which produces a thermal anticyclone in the upper atmosphere. This anti-cyclone weakens the westerly sub-tropical jet stream, and gives rise to the tropical easterly jet stream. This will intensifies the high pressure cell over the Indian Ocean and create a surface pressure gradient between Indian Ocean and India thereby activates south-west monsoon. When the temperature over Tibetan Plateau remains high for a long duration, it strength to the easterly jet stream and results in heavy rainfall over India.

2.3 Onset of Monsoon

In the Indian meteorological calendar, the advent of the Indian Summer Monsoon (ISM) over the southern tip of the Indian peninsula, also known as Monsoon onset over Kerala (MOK), is considered the start of the southwest monsoon (SWM) season. The delay in monsoon commencement has an impact on farmers as well as policymakers. It is difficult to declare the monsoon onset. The advent of ISM represents a transition in the Indo-Pacific region's large-scale atmospheric and oceanic circulation, and while there is no universally recognized description for this transition, it is revealed at the surface by rainfall variability. As a result, monsoon onset is marked by a sudden, significant, and constant rise in rainfall.

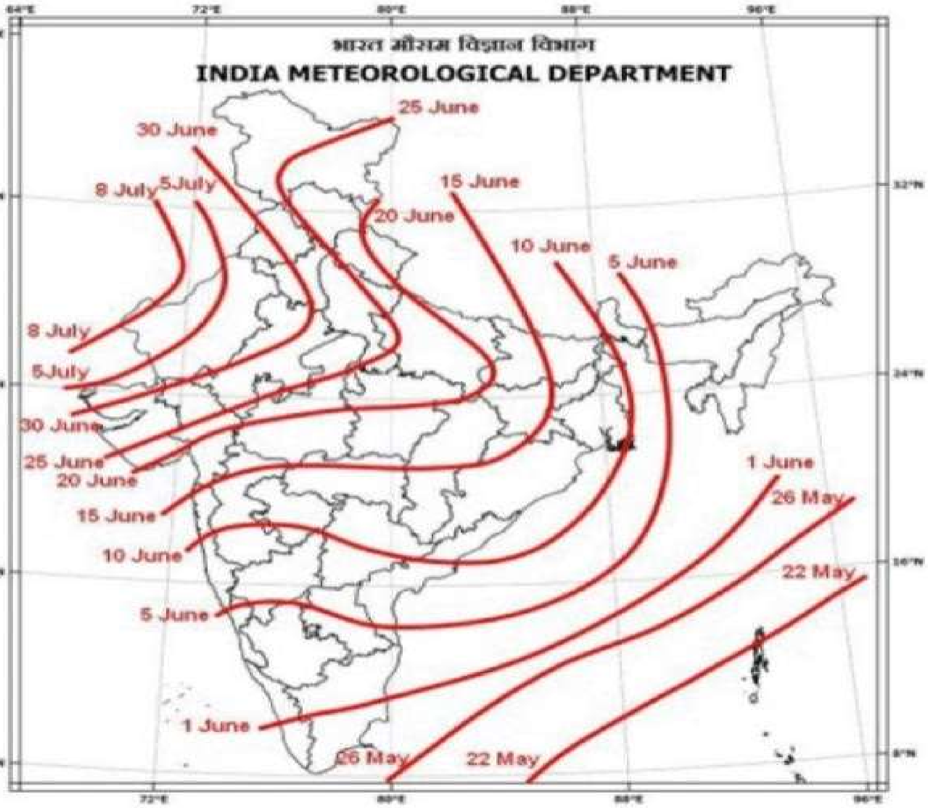

Figure 2.7 Normal dates of SW monsoon onset over Indian region
(Source : www.imdmumbai.gov.in)

For many years, understanding MOK features and projecting the date has been a source of contention.To define the MOK, different criteria are used by various researchers. Ananthakrishnan and Soman (1988) used daily rainfall data collected from different rain gauge network all over Kerala and derived dates of MOK for the period 1901-1980. They define MOK as the first day of the shift from light to heavy rainfall, which means that the average rainfall received in each provision for the next five days must be at least 10mm. According to this criteria, the onset date for south Kerala is 30 May, and the onset date for north Kerala is 1^{st} June, with a 9-day SD. To determine the start and withdrawal of the monsoon, Fasullo and Webster (2003) employed vertically integrated moisture transport, whereas Zeng and Lu (2004) used the normalized precipitable water index. The monsoon onset indexes provided by Xavier et al. (2007) and Wang et al. (2009) are based on large scale meridional temperature gradient reversal in the upper troposphere of the Tibetan Plateau (south) and zonal wind average at 850hPa over the south of AS, respectively. Puranik *et al.* (2013) predicted MOK using OLR and zonal kinetic energy. Every year, the Indian Meteorological Department (IMD) issues an official prediction on the start of the ISM based on subjective forecasting methodologies. From 2006, IMD follows Joseph *et al.,* (2006) criteria that consider rainfall and circulation variables. IMD forecaster consider certain subjective features while declaring the MOK date which includes rainfall data collected from different stations of Kerala, the lower tropospheric winds and the moisture availability up to 500hPa (Rao,1976). The long-term mean date of MOK varies from 30th May to 2nd June with SD 7.28 days.

The commencement of the monsoon is affected by the seasonal reversal of surface and higher atmospheric wind across the southern Peninsula, and accompanying convection is caused by the temperature and pressure contrast between the Indian continent and the Indian Ocean from March to May. MOK covers rainfall across a vast area in the north Indian Ocean, which is accompanied by deep convections (Joseph *et al.,* 1994). The moisture required for this large area is primarily created in South IO, and it is carried out by powerful cross-equatorial LLJ (Findlater, 1969; Joseph and Sijikumar, 2004). Through Ekman pumping of moist air from atmospheric boundary layers, the cyclic shear vorticity zone of LLJ generates deep convections and rains. The onset of ISM over Kerala is linked to a northward shift of the Sub Tropical Jet (Yin, 1949), as well as other factors such as Eurasian region warming due to diabatic heating (Murakami and Ding, 1982), reversal of the meridional temperature gradient (Flohn, 1957; Li and Yanai, 1996), and Tibetan Plateau influences (Yanai *et al.,* 1992; 2006).

Various hypothesis has been proposed to explain the variability in ISM onset. Central Pacific, eastern Pacific and Indian Ocean SST (warm) have positive correlation with monsoon onset. Warming over the central Pacific Ocean causes the Walker circulation to ascend (cyclonic circulation) over the central Pacific and fall (anticyclonic circulation) over the Indian subcontinent. This produces a decrease in convection and latent heat over the Indian continent, lowering tropospheric temperature and delaying onset. Similarly warming of IO modifies the Hadley circulation which increases the TT. Murakami *et al.,* (1986) suggested that the intra-seasonal oscillation (ISO) with period 24-91 days, plays a key role in determining the onset date over south Asia. Between 1870 and 1989, 16 of the 22 MOK delays were linked to a moderate or strong El Nino event. The abnormal persistence of westerlies (easterlies) that exist several days before MOK and enhanced (suppressed) deep convection over the south eastern Arabian Sea and the southern Bay of Bengal are two essential aspects that exist during years with early (delayed) MOK (Sankar *et al.,* 2011).

2.4. Component of Indian summer Monsoon

The surface heat low, the monsoon trough, the cross-equatorial flow, and the low-level Somali Jet (also known as the Findlater Jet), the Mascarene High, the Tibetan anticyclone, and the Tropical Easterly Jet-stream are the regional components of the Indian summer monsoon (Krishnamurti and Bhalme, 1976).

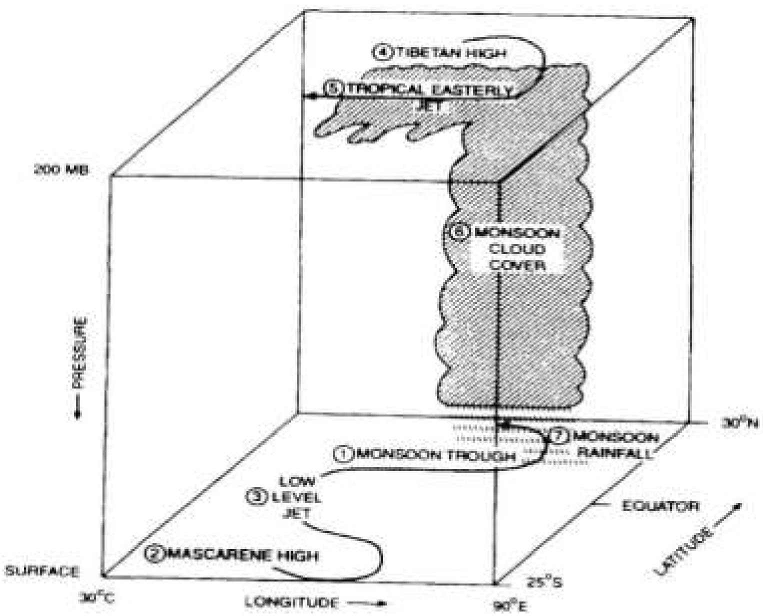

Figure 2.8 Structural elements of the Indian summer monsoon
(Krishnamurti and Balme, 1976)

1. **Surface Heat Low**

During the pre-monsoon months, a heat low gradually develops throughout the subcontinent, eventually settling over Pakistan's central region in July. This low pressure area has a shallow vertical structure, is confined below 1.5 km, and is covered by a subtropical high pressure belt. During the pre-monsoon months (April and May), the surface temperature associated with heat lows steadily rises (Rao, 1976). The severity of the heat low is linked to monsoon activity, with lower and higher pressures favoring monsoon activity in the heat low region and peninsula, respectively.

2. **Monsoon trough**

The monsoon trough was identified by Blanford (1886) and is considered as the important feature of ISM which is seen in the upper levels up to 500mb (6km). It's an elongated low pressure zone that stretches through north India's Indo-Gangetic plains, with an axis that runs roughly west-northwest to east-southeast. Heavy rain is normally limited to 1-2^{o} to south of the monsoon trough (Raghavan, 1973).The axis of the trough runs from Ganganagar to Calcutta through Allahabad and the vertically up to mid-troposphere. The temperature difference of the trough from north to south is 2^{o}C i.e., the tilting is southward with height. In the Indian latitudes, monsoon trough is considered as an equatorial trough of northern summer. The western half of a westerly trough moves north to the foothills of the Himalayas (moving eastward across North West India), resulting in decreased rainfall over the plains of central-north India (monsoon break), and the monsoon is in active phase over the Indian plains when it returns to its normal position. The position and strength of the monsoon trough influence the rainfall distribution over India. (Rao, 1976; Sikka and Narasimha, 1995)

3. **Mascarene High**

The Mascarene High, often known as the Indian Ocean High, is a high pressure system situated at 30°S, 50°E latitude. Summer monsoon features include the mascarene high and the monsoon trough across NE India. The pressure gradient created by monsoon north-south differential heating is measured by the pressure difference between these two locations. The strengthening of the Mascarene High will boost the cross-equatorial flow, forming the east African low-level jet (Findlater, 1977) and the Arabian Sea monsoon current (Sikka and Gray, 1981). The severity of the Mascarene high is linked to the onset of monsoon over India, according to Okoola and Asnani (1981).

4. Cross Equatorial flow

From May to September, a strong southerly flow with a speed of 15 ms^{-1} and a height of 1.5 km above the surface can be seen over eastern Africa. During monsoon season, core of the jet stream can strengthens the speed up to 25ms^{-1}. During boreal summer, the low-level cross equatorial flow becomes much stronger over the western Indian (Hoskins and Wang, 2006) and completes the reverse Hadley circulation by transferring moisture from the South Indian Ocean to South Asia and connecting the Mascarene and monsoon trough. It causes substantial upwelling around Somalia's coast, which lowers the Arabian Sea's SST by 3-4^{o} degrees Celsius and generates the seasonal Somali current (Krishnamurti and Bhalme, 1976; Love, 1985).

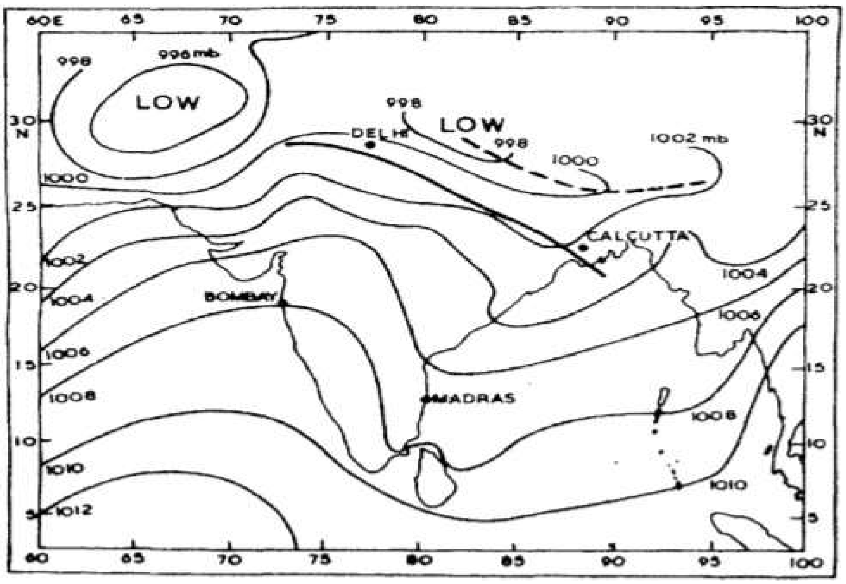

Figure 2.9.The surface heat lows and monsoon trough. Thick line represents the normal position of monsoon trough during active phase of monsoon and dotted line represents its position during a 'break' (Source: Das, 1987)

5. **Tibetan High**

During the summer monsoon season, the Tibetan High is a warm upper tropospheric anticyclone with the maximum amplification across the upper troposphere (200 - 150 hPa). The Tibetan High's development is linked to the elevated Tibetan plateau temperature, which acts as a heat engine during the summer (Flohn, 1957; Koteswaram, 1958; Yanai *et al.,* 2006; Liu *et al*., 2007). The easterly jet is strengthened and the Asian summer monsoon circulation is established and maintained when the temperature above the Tibetan Plateau is high for a long time (Yanai *et al*., 1992). Warm hydrostatic troposphere columns follow the confluence of the Tibetan High with the monsoon trough at sea level over northern India and the Himalayan foothills (Krishnamurti and Bhalme, 1976). Because of its higher height, the Tibetan plateau's thermal convection is stronger than heat lows. The heat disparity between the Tibetan plateau and the equatorial oceans, according to Fu and Fletcher (1985), is connected with the inter-annual variability of the Indian monsoon rainfall.

6. **Tropical Easterly Jet (TEJ)**

The Tropical Easterly Jet (TEJ) is a characteristic of the Asian summer monsoon system that occurs in the upper troposphere. (Sathiyamoorthy *et al.,* 2004; Rao *et al.,* 2004). TEJ is a narrow horizontal air current compose of strong winds with speed roughly around 40-50m/s which are seen in the upper levels of troposphere (100-500hPa) during June to September (Krishnamurti and Bhalme, 1976). The heat differential between the subtropics and the equatorial Indian Ocean feeds TEJ. (Koteswaram , 1958) and potential energy released from Hadley and Walker circulation is the energy source for TEJ. The strength and position TEJ determines the SW monsoon rainfall activity over India. TEJ weakens from the Indian region during monsoon withdrawal and shift southwards.

2.5 Possible causes for ISMR variability

Weather systems that occur worldwide and regionally play a key role in monsoon rainfall variability. Several studies have found that the ocean-atmosphere coupled interaction is a decisive element in the existence of inter-annual variability in monsoon rainfall (Alapaty *et al.,* 1995; Ju and Slingo, 1995; Chandrasekar and Kitoh, 1998). The land-sea temperature differential is the primary factor that influences the Asian summer monsoon. Strong monsoon activity requires a strong temperature difference between the equatorial Indian Ocean (EIO) and North West (NW) India. Other factors that influence the timing of the Indian monsoon include previous year's snowfall in the winter and spring, ENSO, and a heat low across Pakistan and northwest India. Several scientists have spent the last few years researching inter-annual and intra-seasonal variations in monsoon rainfall, as well as the associated phenomena (Gadgil *et al*., 2003; Webster *et al.,* 1998). Because of both external surface boundary forcing and interior dynamics, ISMR fluctuates from year to year (Kulkarni, 2012). ISMR is intimately linked to the heat difference between the Tibetan Plateau and the equatorial Pacific (Fu and Fletcher, 1985). Strong and weak monsoon systems are linked with increased (weakened) land-sea temperature differences in the upper troposphere during monsoon months. Singh and Chattopadhyay (1998) identified a statistical relationship between TT and ISMR anomalies for the period 1961–1990. They argue that in May, a warmer or cooler TT generates excellent or bad rain in India. Several experts (Lau and Bua, 1998; Watanabe and Nitta, 1998; Yang and Lau, 1998) discovered that SST anomalies had a bigger influence on inter-annual monsoon variance than land surface processes. Weak Walker circulations result in drier IO moisture and a lack of rainfall (Bhalme and Jadhav,1984).

The major contributors of summer monsoon rainfall over India are July and August. Since 1950, the ISMR shows significant decreasing trend and this weakening of "ISMR is linked with large-scale perturbations and circulations (Paul et al., 2016)". Recent studies shows that from 1871-2010, the monsoon rainfall shows a significant decrease of 10.9mm in 10 years while the post monsoon shows an increase of 7.5mm (Mini *et al.,* 2016). Decline in rainfall is more visible during June (decrease of 28% to 22.4%) and July (23% to 18.4% decrease) (Mini *et al*., 2016) and an increase was seen during August and September. Rainfall contribution during the summer monsoon season is declining, whereas contributions during the pre-monsoon and post-monsoon seasons are increasing (Mini *et al.*, 2016). Rainfall has been decreasing (increasing) in the southern (northern) areas of the Western Ghats since 1901 (Guhathakurta and Rajeevan, 2008). Major rainfall drops were seen over Madhya Pradesh and neighbouring areas, northeast India, and portions of Gujarat and Kerala, while significant rainfall increases were observed along the West Coast, north Andhra Pradesh, and northwest India (Kumar *et al*., 1992).

Several theories have been proposed to explain the decline in ISMR. The phases of the stratospheric quasi-biennial oscillation (QBO) have an influence on ISMR, according to Mini et al. (2016), with the strong easterly phase of the QBO being connected with weak (dry) Indian monsoon and the weak easterly/westerly phase being associated with robust (wet) monsoon (Mukherjee *et al*., 1985). According to Bollasina *et al.,* (2011) decrease in rainfall over Central North East India is due to high aerosol emissions over Northern India which creates a cooling effect over the land mass. As a result, the temperature difference between the northern and southern hemispheres narrows. Also frequency of drought events over India increases with increased aerosol emission (Ramanathan *et al*., 2005). Another key element that influence ISMR is warming of Southern IO (Rao *et al.,* 2010; 2012). Southern IO warming weakens the meridional SST gradient which debilitate the meridional H adley circulation result in enfeeble ISMR. Another element that influences the ISMR was changing land use land cover (LULC). Evapotranspiration and recycled precipitation (precipitation formed from local land surface evapotranspiration (ET) is known as recycled precipitation) decreases as a result of large-scale conversion of woody savannah to agriculture land (deforestation), resulting in a sharp drop in rainfall, particularly in Northeast India. Recent study showed that the frequency and intensity of heavy rainfall over the Indian monsoon regions are increasing (Francis and Gadgil, 2006) much prominently over the central and northwest part of India and a decreasing nature is visible over Northeast India (Mohapatra *et al*., 2018).

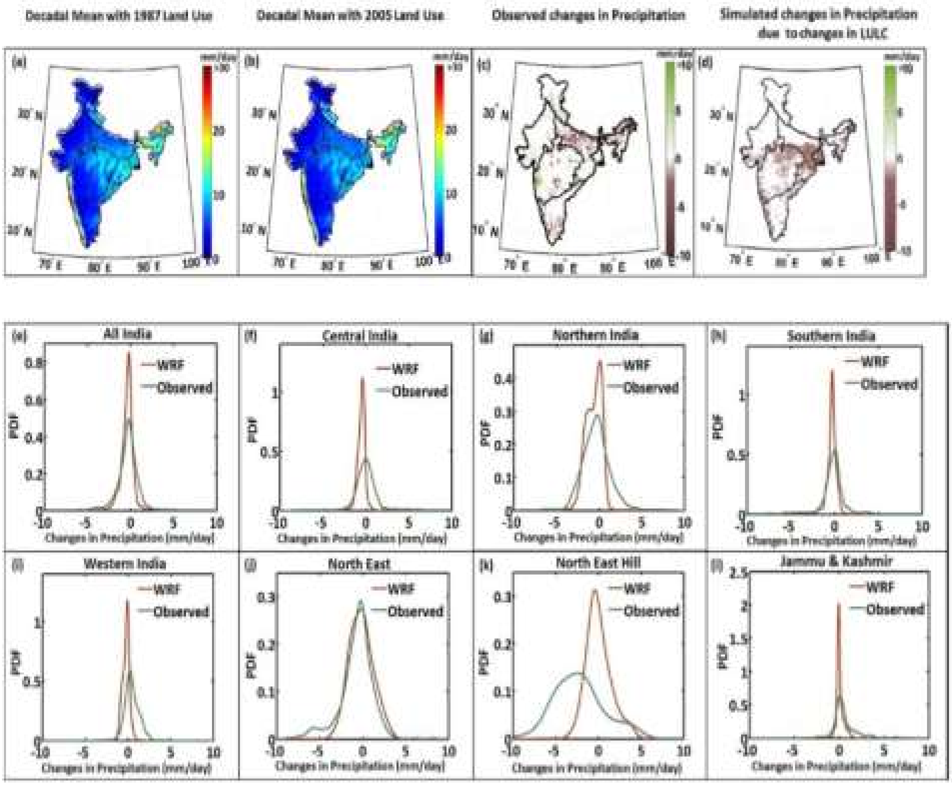

Figure 2.10 Changes in Monsoon Precipitation due to changes in LULC
(Source: Paul *et al.*,2011)

The drying trend in the ISMR has slowed in recent decades, while the frequency of extreme events has increased (Goswami *et al.,* 2006; Ajayamohan and Rao, 2008; Turner and Annamalai, 2012), which has been linked to an increase in synoptic activity over India (Ajayamohan *et al.,* 2010). Increased greenhouse gas concentrations (GHG) (May, 2002; Stowasser *et al*., 2009; Kitoh *et al.,* 2013) contribute to global warming by weakening the intensity of the boreal summer monsoon circulation (monsoon Hadley cell) and associated southwesterly monsoon flow (Krishnan *et al.,* 2013), which leads to weakening of the LLJ (Krishnan *et al.,* 2013 (Rajendran and Kitoh, 2008; Rajendran *et al.,* 2012).

Recent observational modeling studies reveal that the inter-annual variability in ISMR has an inverse relationship with Atlantic Zonal Mode, AZM (Pottapinjara *et al.,* 2016) i.e., strengthening (weakening) of ISMR are connected to cold (warm) phases of AZM. ISMR is influenced by extra-tropical Pacific and

Atlantic Ocean SST forcing and is only visible during non-ENSO years (Chattopadhyay *et al*, 2015). According to Sabeerali *et al.* (2019), the inverse link between ISMR and AZM has been strengthening in recent years due to a large increase in SST anomalies over the tropical Atlantic Ocean, which has resulted in an increase in strong AZM episodes. Induce abnormal tropospheric warming (cooling) over EIO during the warm (cold) phase of AZM. This tropospheric warming (cooling) weakens (strengthens) the ISMR by reducing (increasing) the meridional gradient of TT over the Asian monsoon domain, which leads to a reduction (enhancing) of low-level wind magnitude and moisture transport to the Indian subcontinent (Sabeerali *et al.,* 2012 ; Xavier *et al*., 2007).

2.5.1 Warming Ocean and ISMR?

Contributions from various Sea Surface Temperature (SST) anomalies, particularly the tropical pacific and IO affect the ISM resulting in a prominent inter-annual variability. Over the last hundred years, a significant increase in SST over AS, BOB and equatorial south IO (Kothawale *et al.,* 2008) were observed and it has been accelerating since 1971. The SST variation in the equatorial Pacific Ocean plays a crucial role in predicting the spatial and temporal variability of the ISMR. According to Shukla (1975) and Mooley (1988), increased SST in the Indian Ocean has a substantial impact on ISM, and there is a considerable positive link between Arabian Sea SST during the winter and subsequent monsoon rainfall.

The decrease in monsoon rainfall is attributed to the Indian Ocean warming faster than the surrounding subcontinent, lowering the land-sea temperature gradient (Roxy, 2017). According to Annamalai *et al.* (2013), the tropical west Pacific warming creates a drying trend across South Asia due to changes in atmospheric circulation. The Indo-Pacific warming is modulating the local Walker circulation, causing the low-level monsoon westerlies to weaken (Ratna *et al.,* 2016). The western Indian Ocean has cooler SST mean in summer than the rest of the Indian Ocean, which leads to strong monsoon winds and subsequent upwelling over the western IO, resulting in a zonal SST gradient. The strength and flow of moisture-laden winds toward the South Asian subcontinent are then controlled by this (Izumo *et al.,* 2008). The weakening of the TEJ is accompanied by a change (rise) in SST in the Indian Ocean, which increases convection activity (Joseph and Sabin, 2008).

Mishra *et al.* (2020) discovered that warmer AS causes a drop in monsoon throughout majority of India due to a decrease in convective precipitation. He claims that warming AS diminishes the magnitude of westerly winds in the southern BOB and AS, which causes anti-cyclonic circulation over central India, resulting in lower monsoon rainfall. Also AS warming contributes extreme rainfall across most of Indian region. According to Yadav and Roxy (2019), monsoon drying is caused by changes in the lower level circulation, such as an abnormally high upper tropospheric southerly wind across the Tibetan Plateau due to warm EIO. The rainfall received over the core monsoon zone has changed from heavy rain events to moderate events (Goswami *et al*., 2006) meanwhile (Rajeevan *et al.,* 2008) central Indian experiences extreme rainfall events and this trend is associated with the increase in SST and surface latent heat flux over tropical Indian Ocean. From 1970 to 2002, the ISMR rainfall declined at a rate of 1.5mm/year due to a constant rise in temperature (Kothawale *et al*., 2008).

2.5.2. How temperature drives precipitation?

Temperature Changes alters the atmospheric circulation and thus create large-scale precipitation patterns. Schewe and Levermann (2012) predicted the temperature increase in last 21st and early 22nd century may produce regular changes in precipitation, such as a shift in monsoon precipitation up to 70% below average. It might also push out the onset of the monsoon by up to 15 days (Ashfaq *et al.,* 2009). The IPCC's fourth and fifth assessment reports [IPCC, 2007, 2013] noted that the global mean surface temperature has risen by 0.74°C and 0.80°C, respectively, over the last century, and that this is due to increased greenhouse gas concentrations, which result in a consistent increase in surface and tropospheric temperature. Extreme precipitation intensifies when global mean surface temperature rises (Kharin *et al*., 2013; Fischer and Knutti, 2016). The observed intensity in extreme precipitation events (the rainfall per unit time) upsurges with surface temperature (Sillmann *et al.,* 2013; Fischer and Knutti, 2016). Model studies indicates that tropospheric warming results moisture enhancement in atmosphere (Trenberth *et al*., 2005; Pumo *et al*., 2019) which leads to the increase in heavy rainfall events (Kumar *et al*., 2002). From 1951 to 2000, Goswami *et al*. (2006) found a significant increase (decrease) in the frequency and amplitude of extreme (moderate) occurrences over central India during the monsoon seasons.Ocean-atmosphere heating contrasts drive regional circulation patterns, particularly the meridional gradient of the deep tropospheric heat between India IO is the important driving factor for deep Asian monsoons circulation (Webster *et al*., 1998). Strong (weak) ASM years are linked to positive (negative) TT anomalies over Eurasian, negative (positive) temperature anomalies over IO and eastern Pacific, negative (positive) SST anomalies over equatorial eastern Pacific, AS, BOB, and south China Sea (Li *et al*., 1996).

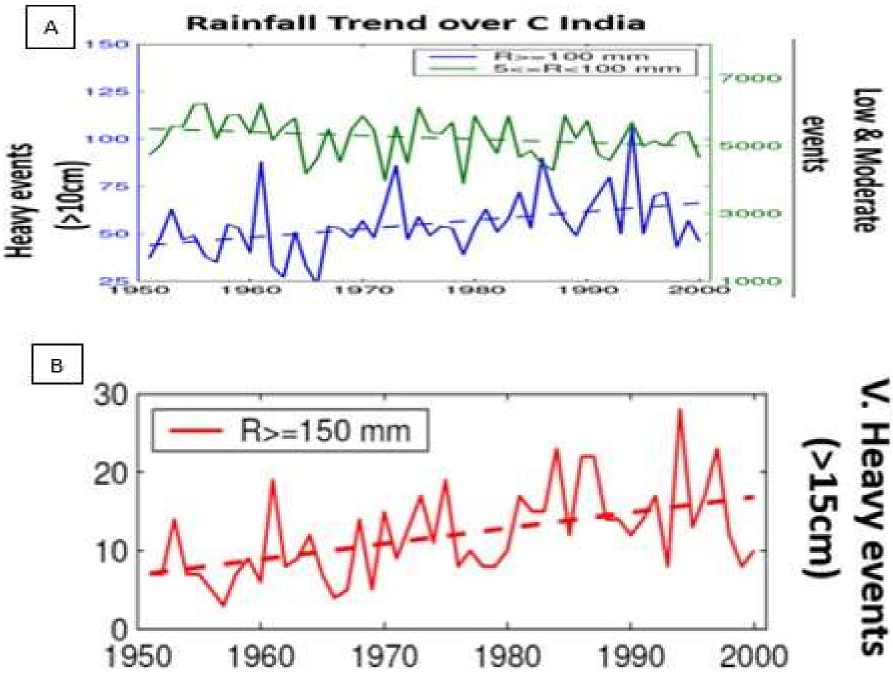

Figure 2. 11. Temporal variation (1951 to 2000) in the number (N) of heavy (R ≥ 100 mm/day), moderate (5 ≤ R < 100 mm/day) daily rain events and very heavy events (R ≥ 150 mm/day) during the SWM (Source: Goswami et al.,2006)

2.5.3. Weakening of Tropical Easterly Jet Stream (TEJ) suppress monsoon circulation??

In boreal summer (Figure 2.12), two jets, one over the Indian Ocean (IO) (with a speed of 23 m/s), form due to the meridional thermal contrast between the Asian landmass and the IO, which is exacerbated by the Tibetan Plateau heating due to its elevation (Koteswaram, 1958), and the other over the tropical eastern Pacific (TEJ), which is responsible for North American monsoon. The IO TEJ develops and retreats during the onset and withdrawal of the Asian summer monsoon respectively by altering the strength of the lower level westerlies and the upper level easterlies, creating a significant seasonal migration (Raman *et al.,* 2009). The strength of the jet stream can be used to predict long-term changes in the atmosphere's general circulation (Pielke *et al.,* 2001). According to Sathiyamoorthy (2005), the strength of TEJ lessened at 200 hPa over the African region but was not noticeable over the Indian monsoon zone. Pattanaik and Satyan (2000) identified that the intensity of IO TEJ have strong correlation with the ISMR and are primarily attributed to the tropical SST anomalies over the tropical central-eastern Pacific. That is ISMR enhances during a stronger TEJ and decreases during a weaker TEJ (Pattanaik and Satyan, 2000; Kanamitsu *et al*., 1972; Tanaka, 1982). Also, the TEJ can be intensified by a La Niña event and extenuate by an El Niño event (Arkin, 1982; Tanaka, 1982; Chen and van Loon, 1987; Huang *et al.,* 2019).

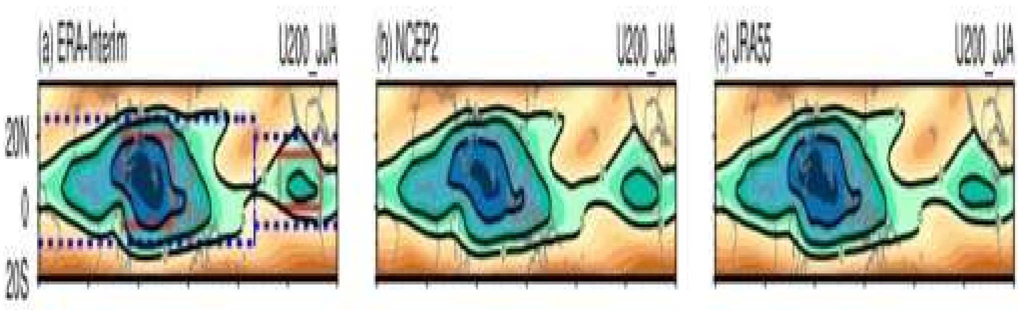

Figure 2. 12. Climatological zonal wind at 200 hPa in boreal summer (June–August) from 1995 to 2014 (Source: Huang *et al.,* 2020)

The high tropospheric heat gradient between the tropical Indian Ocean and the Tibetan anticyclone zone feeds the TEJ (Figure 2.13). Temperatures in the upper troposphere over the Tibetan anticyclone region and the Indian Ocean regions have been cooling and warming, respectively, since 1950 (Abish *et al.,* 2013). The upper tropospheric meridional temperature gradient decreases as a result of this warming and cooling trend, and the strength of easterly thermal wind at the TEJ's core region decreases. Also the El Niño like warming pattern over the tropical Pacific play a crucial role in future weakening of the TEJ which result in suppressed rainfall over the tropical eastern IO (Huang *et al*., 2019). Deterioration in the wind strength at the core of TEJ over the South Asian region (Rao *et al.,* 2004) and decrease in horizontal extent over Atlantic Ocean and regions of West Africa (Sathiyamoorthy, 2005) favors tropical cyclone formation during monsoon season over Indian Ocean due to the reduction in the vertical wind shear (Rao *et al.,* 2008). He came to the conclusion that the weakening of the TEJ and associated decrease in easterly shear is linked to a decrease in the temperature gradient between the equator and 20º N with latitude belt 40-100ºE, i.e., air temperatures in the equator are comparatively higher than those in North latitudes and are much more prominent at 500mb. Chen and van Loon (1987) claimed that inter-annual variability in Walker and Hadley circulation will bring variations in TEJ as it energetically maintained by these circulations (Chen, 1980). Krishnan *et al.,* (2013) concluded that, the intensity monsoon Hadley cell and the related LLJ have diminished dramatically in response to global warming during the last 50 years. The enormous warming of Indian Ocean increases the deep convection which leads to weakening of TEJ (Joseph and Sabin, 2008).

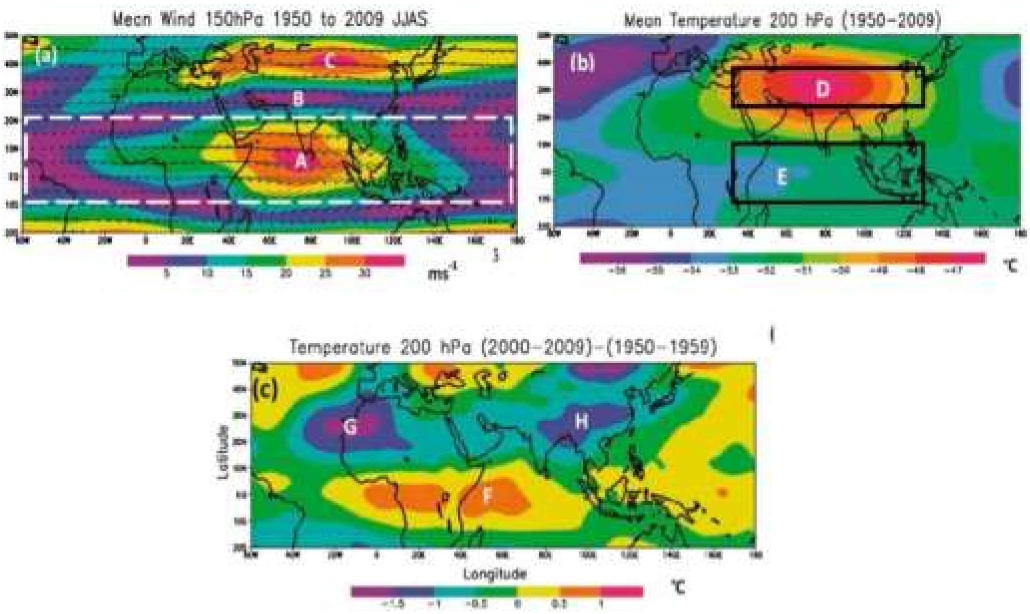

Figure 2.13. The climatological mean of wind (at 150hPa) and temperature (200hPa) for the period 1950–2009(JJAS). "A": TEJ. "B": Tibetan anticyclone and "C": Asian subtropical jet stream. (Source: Abish *et al.,* 2013)

2.5.4 Influence of LLJ

During the boreal summer monsoon season, a strong cross-equatorial low level jet stream (LLJ) with a core at 850 hPa exists across the Indian Ocean and south Asia. The LLJ is generated by cross-equatorial flow influenced by summer hemisphere differential heating between 20^{o}N and 20^{o}S latitudes, resulting in pressure gradient heat lows across the Indian subcontinent and Mascarene highs (Krishnamurthi and Bhalme, 1976). The north-south pressure gradients caused by differential heating of land and ocean drive LLJ, while the Coriolis force modifies the flow. In 1966, Joseph and Raman established the existence of a westerly LLJ (during July) with a core about 1.5 km above mean sea level with wind speeds of 40–60 kt across peninsular India, as well as substantial vertical and horizontal wind shears. The LLJ is fundamental in the monsoon because it transports moisture from the southern Indian Ocean to the Indian subcontinent and regulates the establishment of the monsoon inversion layer over the western central Arabian Sea (Roxy *et al,* 2017; Sathiyamoorthy *et al.,* 2013). The strength, position, and vertical structure of LLJ are indeed factors in the amount of moisture carried to the Indian land mass (Sandeep and Ajayamohan, 2015). The interplay of the LLJ with the Western Ghats orography results in a large amount of rainfall along India's west coast. Because it is the primary source of moisture and momentum flux, changes in the monsoon LLJ will have a considerable impact on monsoon transport and precipitation.

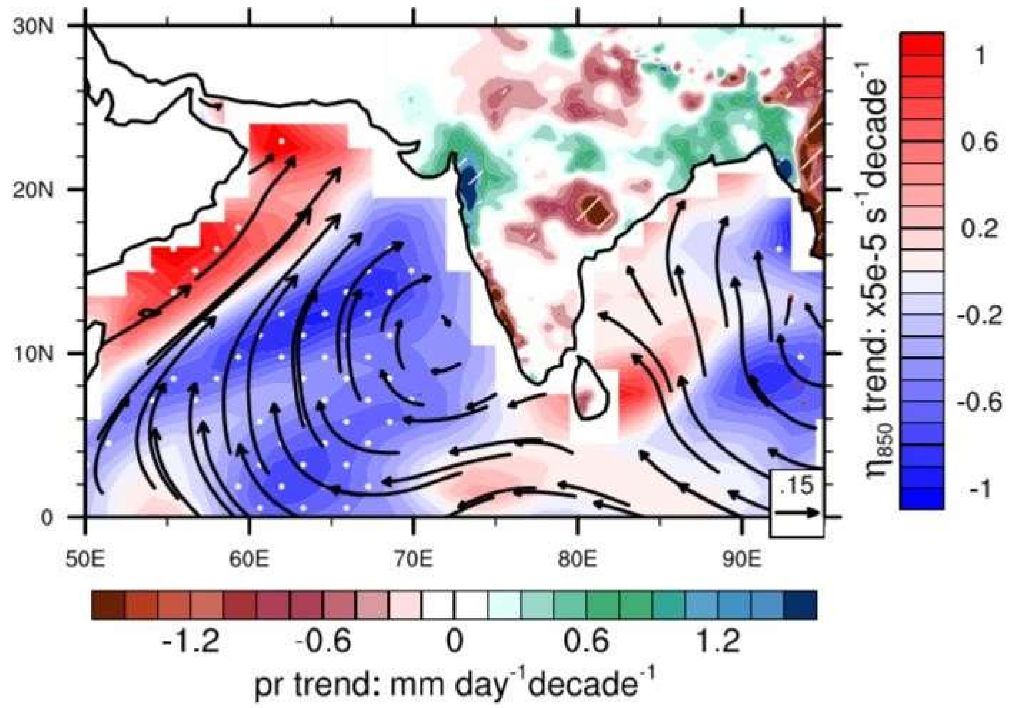

Figure 2.14 Composite variation of LLJ during monsoon months
(Source: Subrahmanyan ,2016)

Subrahmanyan (2016) pointed out the intensification of LLJ is seen from 1997 to 2012 after the onset, predominantly in July and September (Aneesh and Sijikumar, 2016). The broadening of the tropical belt (Fu, 2006; Seidel and Randel, 2007), the poleward shift of mid-latitude jet streams, and mid-latitude storm paths are the principal effects of global warming on large-scale atmospheric circulation (Fu and Lin, 2011). A shift in the LLJ's strength and location has a negative impact on the circulation and precipitation patterns in South Asia (Sandeep, 2014).The monsoon and west African LLJ exhibits a northward shift, while the Caribbean and south American LLJ shows a westward expansion (Torres-Alavez, 2021). Intraseasonal rainfall variability is linked to a diminishing tendency in the LLJ strength (Joseph and Sijikumar 2004), which causes a dry (wet) trend across the southern (northern) half of India's western coast. Further studies reveal that LLJ shifts towards pole-ward is due to increased land-sea contrast which strengthens the cross equatorial sea level pressure over IO (Sandeep and Ajaymohan, 2015). The LLJ appears to be strengthening (weakening) north (south) of 15°N over the Indian Ocean, causing the absolute vortices over the southern (northern) Arabian Sea to weaken (strengthen). The poleward shift in LLJ has the potential to reduce SSTs across the northern Arabian Sea, affecting ISM intra-seasonal variability. Changes in the LLJ over the Arabian Sea are linked to an increase in the number of extreme occurrences across central India (Roxy *et al.*, 2017).

In recent decades, LLJ has become more strongly linked to pIOD (positive IOD) over EEIO and AS. As a result, a moisture convergence zone formed over India's northeast regions. In case of nIOD (negative IOD), LLJ get strengthen (weaken) during June (July, August, September) due to enhance (suppressed) moisture transport to Indian region (Hrudya *et al.,* 2021).

2.5.5 Influence of ENSO and IOD

As part of the world's largest warm pool, the tropical Indian Ocean and its interaction with the atmosphere play a critical role in determining regional climate (Schott *et al.,* 2009). The Indian summer monsoon, El Nino and Southern Oscillation (also known as ENSO), and the Indian Ocean Dipole all seem to be factors that influence the Indian Ocean. Until IOD was identified, ENSO was thought to be the most important external driver of ISMR. Recent studies shows that ISMR are complementarily affected by IOD and ENSO (Ashok *et al.,* 2001). According to him, the ENSO and IOD induce anomalous circulation cells, either act as counter or supported phenomenon over Indian region depending upon the phase and amplitude.

ISMR is linked to Hadley circulation on a regional basis as well as Walker Circulation on planetary (Soman and Slingo, 1997; Dai and Wigley, 2000), which are altered by the influence IOD and ENSO respectively. Ashok *et al.,* (2001) concluded that whenever ENSO and ISMR shows low (high) correlation, the IOD relationship with ISMR will be high (low) i.e., IOD acts as a modulator which subsequently influence the correlation between ISMR and ENSO. The maximum correlation is seen between ISMR and monthly Nino 3.4 during August-November (Gershunov *et al.,* 2001).

El Nino and Southern Oscillation (ENSO)

ENSO is a large scale coupled atmospheric-ocean aperiodic oscillation in Pacific Ocean that occur at every 2-7 years. El Niño is defined as a coupled climatic phenomenon distinguish with large scale anomalous warming of central to eastern tropical Pacific Ocean, associated with anomalous cooling in the tropical western Pacific region, creating extensive sequel globally meanwhile La Niña is the colder counter part of El Nino. El Nino usually appears in March or April and might last a year or longer, with the highest SST around late December. Between June and September, it normally appears to diminish temporarily when favorable SST anomalies and the South Pacific trade winds fade. Normally during the El Nino years, Indian monsoon is weaker compared to other years. Walker (1924) observed that during El Nino (La Nina) years, the entire Walker circulation shifts somewhat eastward (westward), helping to suppress (enhance) convection across the Indian subcontinent by descending the walker cell branch on the western IO (westward). El Nino does not directly suppress ISMR by descending walker branch, according to many studies (Goswami, 1998; Slingo and Annamalai, 2000), but the change in Walker circulation helps the Hadley circulation to descend across the Indian subcontinent. That is, ENSO affect ISMR by altering the Walker and Hadley circulations.

In addition to it, a new type of El Niño, named El Nino Modoki have been occurring since 1970, is anomalous warming in the central tropical pacific and cooling of SST anomalies in eastern and western pacific ocean (Ashok *et al.,* 2007; Marathe *et al.,* 2015) .The location of its heat source has relevant impact on its domain (Soman and Slingo, 1997; Annamalai and Liu, 2005) particularly ISM.

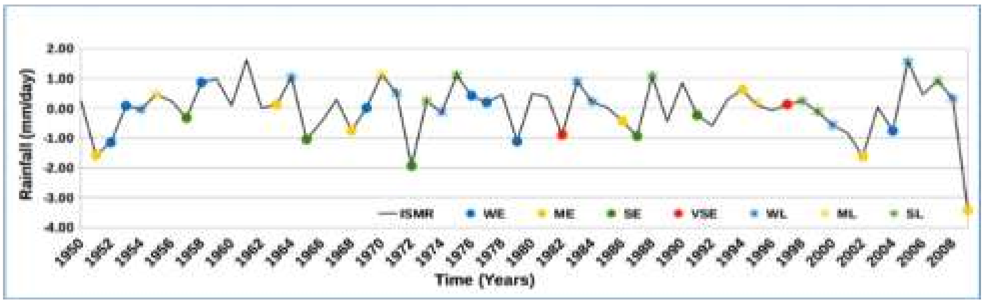

Figure 2. 15 The inter-annual variability of the ISMR over (66.5° E-101.5° E; 6.5° N-39.5° N) during the 1951-2009 period.

Generally El Nino years are associated with deficit or below normal summer monsoon except certain years (Sikka ,1980; Ramage ,1983; Rasmusson and Carpenter ,1983) while during La Niña events, greater than normal rainfall is seen. Most of the (50%) drought in India are linked with ENSO (Kripalani and Kulkarni, 1996). The majority of Indian region shows a negative correlation with Nino 3.4 Index which indicates that during boreal summer the anomalous warm conditions over the eastern tropical pacific may result in anomalous deficit rainfall over India. However, the El Niño Modoki events reduce the ISM rainfall more effectively than the conventional El Niño (Kumar *et al.,* 2006). It was found that conventional El Nino is linked with negative rainfall anomalies along the monsoon trough, while the Modoki El Nino cause deficit rainfall over peninsular India (Ashok *et al.,* 2007; Amat and Ashok,2018).

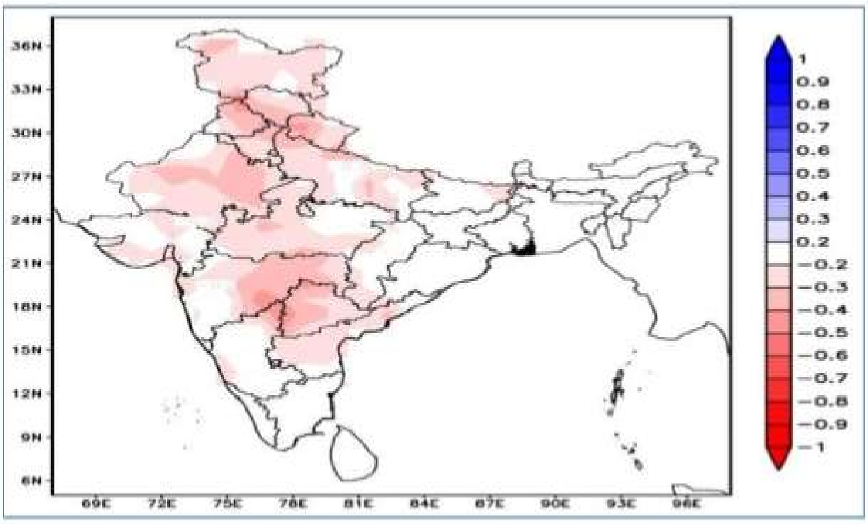

Figure 2.16 Simultaneous linear correlations between Niño3.4 indexes with the JJAS rainfall for the period 1901-2009

SST changes in the eastern tropical pacific during ENSO influences the ISM by changing Walker circulation. According to Webster *et al.* (1998), boundary anomalies caused by ENSO cause alterations in the Walker Circulation. Large-scale circulation changes are noticeable during El Nino (La Nina) years due to an eastward (westward) shift in the Walker Circulation and decreased (increased) equatorial divergence over the tropical Indian Ocean. The anomalous convergence in the tropical Indian Ocean modulates the cross-equatorial meridional circulation, which causes an anomalous divergence over the Indian region which result in drier (wetter) rainfall than normal ISM (Ashok *et al.,* 2001**)**.Also, Goswami and Xavier (2005) pointed out that ENSO influences the meridional TT gradient over the Indian region thereby effectively modulating the strength and duration of the monsoon through the subtropical jet (Shaman and Tziperman, 2007)

Indian Ocean Dipole

The IOD event is considered to be a crucial component of tropical climate variability (Behera *et al.,* 2006) and a key modulator of ISMR variability (Ashok *et al.*, 2004; Webster *et al.*, 1999). The Indian Ocean Dipole (IOD) is an ocean-atmosphere linked phenomenon that originated in the Indian Ocean (Saji *et al.,* 1999; Webster *et al.,* 1999). In the Indian continent, positive (negative) IOD episodes are marked by increased (decreased) rainfall (Ashok *et al*., 2001; Hrudya *et al.,* 2020). It can also modulate rainfall occurrences in central India by producing a convergence zone over the region by moisture movement from the southeast EIO to the BOB (Ajaymohan and Rao, 2008). Positive IOD events are becoming more common (Abram et al., 2008), and the ISMR-IOD link is intensifying over time (Ashok *et al.*, 2004; Krishnaswamy *et al*., 2015), especially during the monsoon withdrawal period (September) (Hrudya *et al*., 2020). The IOD-ISMR relationship is strengthening because of non-uniform warming of the Indian Ocean, i.e., the western equatorial Indian Ocean warms faster than the EEIO (Ihara *et al*., 2008; Cai *et al*., 2013)

2.5 Kerala Summer Monsoon Rainfall (KSMR) and its variability

The State 'Kerala' lies in the extreme southwest of the Indian sub-continent, outlined by Karnataka in the north, Western Ghats (WG) in the east and by the Arabian Sea in the west, between 8º15' N and 12º50' N and 74º 50'E and 77º30'E. Kerala is popularly known as "Gateway of the summer monsoon" since the southwest monsoon typically starts from here by June 1st with standard deviation of seven days (Pai and Nair, 2009). Although the state is 38863 sq.km in area the region is topographically diverse, it varies from hilly area to more or less flat surfaces. The terrain height varies from few meter in the west to few kilo meters in the east with rugged topography having narrow and steep slopes. Based on this, Kerala is divided into three distinct zones: the eastern high lands, the hilly midlands and western low lands.

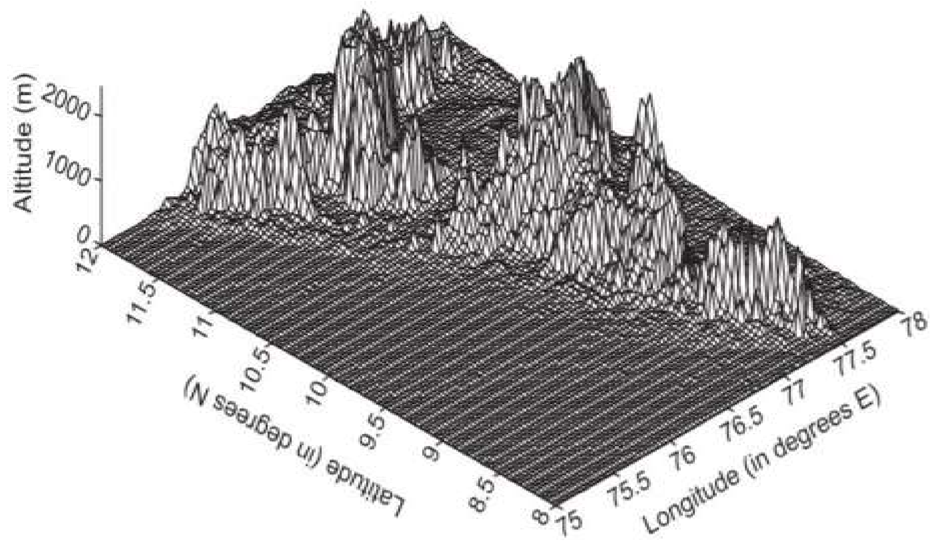

Figure 2.17 Terrain Height of Kerala (Source: Soman and Mohankumar, 2004)

The annual rainfall in Kerala is around 300 cm, with a 15% fluctuation, however it varies greatly on a spatial scale, from 200 cm in the south to 380 cm in the north (Abhilash *et al.*, 2018). The state's annual rainfall pattern shows north-south and east-west gradients. The Western Ghats (with a height of 1-2 km) operate as a natural barrier to the low-level monsoon circulation. The southwest monsoon flow, which supplies moisture for rainfall, is forced upward by the Ghats, resulting in high rainfall on the windward side. In the high lands, there are two pockets of rainfall that exceed 400cm: the northern pocket, which is located to the north-east of Kozhikode and includes the hill station Vaithiri, receives over 450cm of annual rainfall, and the southern pocket, which is located to the east of Kottayam and includes Peermade, receives more than 400cm of annual rainfall. During the monsoon months, the western windward slopes of the hills and mountains receive a lot of rain and serve as a watershed for several rivers. Rainfall on the windward slopes, on the other hand, is not evenly distributed, ranging from heavy to comparatively light rain.

The south-west monsoon (June–September) and the north-east monsoon (October–November) are the two main rainy seasons in Kerala. The pre-monsoon months (March–May), which are marked by thunderstorm activity, have a minor amount of rainfall, whereas the winter months (December–February) are marked by minimal clouding and rainfall (Ananthakrishnan *et al.*, 1979). The monthly rainfall distribution (Figure 2.18) demonstrates that SWM accounts for 65–70% of yearly rainfall, followed by the post-monsoon season (15–20%) and the pre-monsoon season (5–10%). Seasonal rainfall during the monsoon and post-monsoon seasons reflects the presence of regional variation.

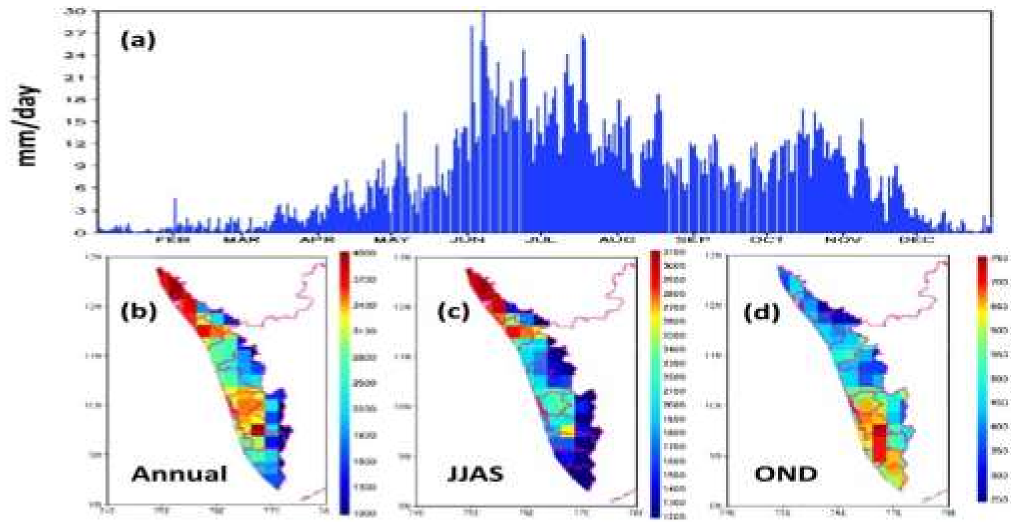

Figure 2.18 Area averaged monthly rainfall (mm/day) over Kerala state and (b–d) spatial distribution rainfall: annual, JJAS and OND seasons (1901-2016) (SOURCE: Abhilash et al., 2018)

Low level westerly jet stream (LLJ) and upper level tropical easterly jet stream (TEJ) are two dynamic components of the ISM system that determine rainfall across the peninsular (Joseph and Simon, 2005; Sathyamoorthy, 2005). Rainfall is modulated on a daily to seasonal scale by the position and intensity of LLJ and TEJ. The Low-Level Jetstream (LLJ) has a significant impact on monsoon rainfall in Kerala (Findlater, 1969). The north-south pressure gradients caused by differential heating of land and ocean, as well as the Coriolis force, are the primary driving forces for LLJ. Krishnamurthi (1985) discovered that a location to the north of the LLJ axis with cyclonic vorticity in the boundary layer can support forced convection due to boundary layer friction. During active spell of SWM season, cyclonic shear area of LLJ is over north Kerala and anti-cyclonic shear zone in south Kerala (Simon *et al.,* 2004). As a result, during SWM, North Kerala receives more rainfall than south Kerala, and during the post-monsoon season, south Kerala receives the most rainfall (Ajithkumar *et al.,* 2015). The wet phase of KSMR is linked to the weakening (strengthening) of deep convection over EEIO, according to Sreenath and Abhilash (2021). They claim that during the wet phase of the KSMR, an unusual low-level easterly flow will emerge from the EEIO and proceed towards the Indian subcontinent, creating oceanic upwelling and lowering the SST across the EEIO and AS.This will enhance the temperature gradient between Indian subcontinent and EIO, thereby strengthens the monsoon flow towards India. As a result, any rainfall deficits or excesses during different seasons may have varying regional effects across Kerala.

Several researches pointed out the decreasing trend in annual and mean summer monsoon rainfall over Kerala (Kumar *et al.,* 1992; Soman *et al*., 1988; Mini *et al*., 2016; Kothawale and Rajeevan, 2017; IMD, 2018) and the decrease is approximately 10-20% (Soman et al.,1988). From 1901-2010, the low level winds (westerlies) over the peninsular India weakens whereas they deepen over the equatorial Indian Ocean (south of Indian peninsula), along with dwindling of upper level easterlies. This suggest the weakening of LLJ over Indian region (Joseph and Simon, 2005). Future projections shows the pole-ward shift in LLJ (Sandeep and Ajaymohan, 2014) which tend to weakens the monsoon circulation. In the upper levels, a weakening trend in the north-south temperature gradient has a significant impact on (weakens) TEJ (Abish *et al.,* 2013). All of these climatic changes will make a substantial difference in monsoon rainfall in Kerala.

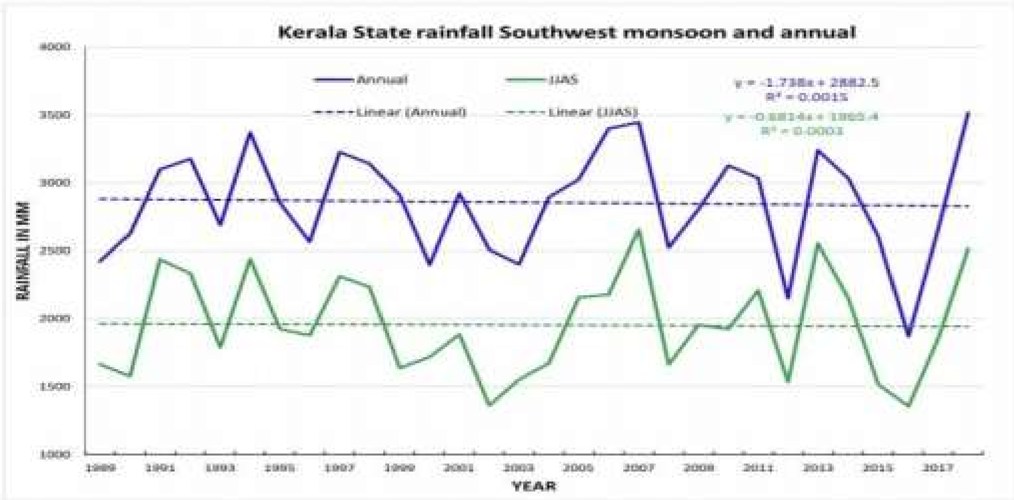

Figure 2.19 Annual and JJAS Rainfall Trend (Source: Imd, 2018)

Based on atmospheric dynamics, this study explores large-scale factors that favors the surplus and deficiency phases of Kerala Summer Monsoon Rainfall (KSMR). Kerala receives roughly 2,000 mm of seasonal rainfall (Krishnakumar *et al.,* 2009), which is determined by orographic factors as well as the strength and position of the LLJ. As a result, changes in lower level (850 hPa) and upper level (200 hPa) circulation patterns are investigated in order to determine the role of the LLJ in controlling the excess and deficiency phases of the KSMR.

MATERIALS AND METHODS

CHAPTER 3

MATERIALS AND METHODS

The study titled “Seasonal and Monthly Rainfall Variability over Kerala in a Warming Climate” was studied during the period. Details regarding the study area, data and methodologies used in the study are conferred in this chapter.

3.1 Area of the study

Variability in summer monsoon was studied based on the rainfall received over Kerala which is a state in the southwestern Malabar Coast of India with geographical area of 38,863 sq.km. It lies between 8^{o}15N & 12^{o} 50’N latitudes and 74^{o} 50’E & 77^{o} 30’E longitudes with 14 districts. It is known as the “Gateway of Summer Monsoon”. It is a narrow strip of land that runs between the Arabian Sea and the Western Ghats on the west and the Western Ghats on the east. The population of the state is 3,33,87,677 (Census 2011, 2011). Majority of the population are directly or indirectly depends on agriculture for their livelihood. The state's agricultural area is 2.19 million acres, accounting for 56 percent of the total land area. The net irrigated area is 0.38 million ha, accounting for 17.5 percent of the cultivable area, with the rest being rain fed. Therefore agricultural operations of the state mainly depends in SW Monsoon.

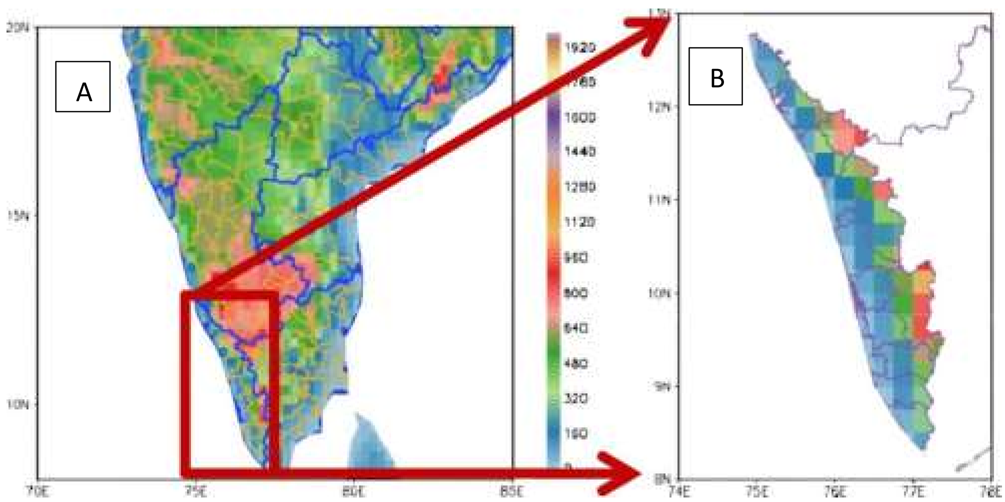

FIGURE 3.1.(A) Location map at a resolution of (0.250 × 0.250), (B) Kerala’s zoomed topography

Rainfall

The rainfall data used in the study were collected from India Meteorological Department (IMD) and National Oceanic and Atmospheric Administration (NOAA). IMD provide high resolution (0.25⁰ x0.25⁰ lat. /long.) daily gridded rainfall data set over India from 1901-2018 (Pai *et al.,* 2014) and is available on request from http://www.imdpune.gov.in . For the present study, data for the period 1948 to 2019 were used. By screening out other outside areas with a mask file constructed at the same resolution (0.25 0 x 0.250 horizontal resolutions), the gridded rain gauge observations for Kerala regions were retrieved. Also GPCP v2.3 (Global Precipitation Climatology Project) were used in the study. They provide consistent precipitation analysis (Adler *et al.,* 2003) by combining data from rain gauge stations, satellites, and sounding observations to estimate monthly rainfall on a 2.5 × 2.5 grid resolution from 1979 to the present. and can be downloaded from www.psl.noaa.gov .

Sea Surface Temperature

Extended Reconstructed Sea Surface Temperature (ERSST) was used in the study. It is a global monthly sea surface temperature (SST) dataset which are obtained from the International Comprehensive Ocean- Atmosphere Dataset (ICOADS) (Huang *et al.,*2017) and the data are available from 1854 to present ,can be downloaded from www.ncdc.noaa.gov. In this study, ERSST v5 was used from 1948 to 2019.

Other physical parameters that are used in the study includes outgoing long wave radiation (OLR),wind (at 850 and 200hPa),vertical velocity (omega), specific humidity, air temperature and pressure which were downloaded from NCEP/NCAR Reanalysis data set from 1948 to 2019.

Nino 3.4

The Nino 3.4 index was designed to indicate the phases of El Nino Southern Oscillation, ENSO (El Nino and La Nina) events in terms of understanding the influence of Pacific Ocean SST (Bunge and Clarke, 2009). It is defined by five consecutive 3-month running mean SST anomalies in the Nino 3.4 zone of the Pacific Ocean (50N-50S and 1200W -1700W).The ENSO events are defined when the Nino 3.4 SST exceeds +/- 0.5° C for a period of six months or more. The Nino 3.4 data are obtained from www.origin.cpc.ncep.noaa.gov.in .

Dipole Mode Index

The intensity of the Indian Ocean Dipole (IOD) is measured using the Dipole Mode Index (DMI) .It's the difference between equatorial Indian Ocean SSTs in the west (500 E to 700 E and 100S to 100N) and the east (90ºE to 110ºE and 10ºS to 0ºS).If SST difference is positive, it is known as positive IOD and if negative, known as negative IOD (Saji *et al.,* 2003). The DMI data was downloaded from www.psl.noaa.gov.in from 1948 to 2019.

Vertical Integrated Moisture Flux (VIMF)

Vertically integrated moisture transport is calculated by vertically integrating the moisture fluxes of the u and v component. In this study VIMF are derived using the NCEP/NCAR reanalysis data. The vertical moisture flux convergence over the intervals 1000–850, 775–700, and 600–500 hPa is computed using finite centered differences on a lat./long. grid. (Darand and Pazhoh, 2019).

$$VIMFC = -\frac{1}{g} \cdot \int_{psurf}^{ptop} \left(\frac{\partial uq}{\partial x} + \frac{\partial vq}{\partial y} \right) \cdot dp \qquad (1)$$

In the equation, q represents specific humidity, u and v denote the wind's x- and y-components, p is pressure, psurf signifies surface pressure, ptop indicates pressure at the top of the atmospheric layer, and g symbolizes gravitational acceleration.

Tropospheric Temperature (TT)

The TT anomaly has a strong and positive relationship with the Indian monsoon rainfall (Parthasarathy et al., 1990; Singh and Chattopadhyay, 1998). The average temperature of the troposphere between 700 and 200 hPa is defined as TT.

3.3 Technical Program

To understand the seasonal and monthly KSMR trend, standardized rainfall anomaly for the period 1901-2019 were done. The long term and decadal trends of KSMR were analyzed. The study focuses the monsoon variability during excess and deficit years. To identify excess and deficit years, percentage departure analyses were carried out from 1948-2019.To understand the influence of physical parameters (such as temperature, specific humidity, vertical velocity, perceptible water, sea level pressure (SLP), outgoing solar radiation (OLR), sea surface temperature (SST)) composite analyses were carried out. The physical mechanism that drive behind the excess and deficit years were identified and analyzed in this study.

Also the impact of Indian Ocean and Pacific Ocean SST over KSMR were analyzed by considering the Dipole Mode Index and Nino 3.4 Index. Monthly rainfall trend and their relationship with IOD were also included in this study. Systematic representation of technical program is given below:

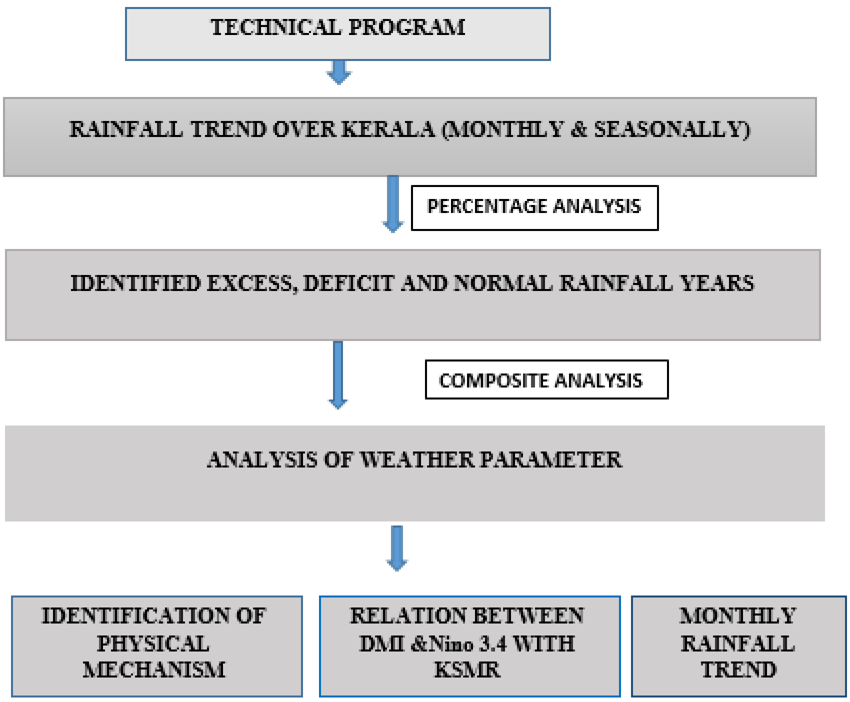

Figure 3.2 Schematic Representation of the technical program

3.3.1 Standardized Anomaly Index

For regional climate studies, Standardized Anomaly Index (SAI) is used analyzing fluctuations. Daily rainfall data during June, July, August and September were collected from IMD and are converted to monthly data sets. (Babatolu *et al.,* 2013). Monthly rainfall is expressed as a standardized deviation xi from the long-term mean (i.e. base period mean), which is determined as

$$x_{\mathrm{i}} = (r - r_{\mathrm{i}})/\sigma \qquad (2)$$

where r is the mean annual rainfall (JJAS), r_i is the long term mean and σ is the standard deviation. When x_i is positive (negative) it shows normal to wet (dry) conditions. By subtracting the long-term mean and then dividing it by the standard deviation gives the standardized series of ISMR. Deficit (excess) monsoon years are those in which the standardized rainfall anomaly is less than (greater than) 1 (+1).

3.3.2 Percentage Departure Analysis

Percentage departure criteria is used to categories the rainfall in excess, deficit and normal. If percentage departure of realized rainfall to normal rainfall is +20% or more (-20% or less) it is considered as excess (deficit) rainfall.

$$Percentage\ Departure = \frac{Observed\ Rainfall - Predicited\ Rainfall}{Predicted\ Rainfall} \qquad (3)$$

3.3.3 Composite Analysis

To understand the nature of Climatic parameters, composite anomaly of excess and deficit years for individual parameter were plotted using Open GrADS. To get the composite anomaly of individual parameter, first take the average of the variable taken over specific selected time period with common characterizes (excess / deficit years). Then subtract the above average with the climatology (1901-2019) of the variable gives the composite anomaly of each variables.

3.3.4 Statistical Techniques

i. Trend Analysis: Important elements of climate research includes "detection, estimation and predication of trends" and connecting it with statistical and physical parameters are significant (NCAR, 2014). Simple linear regression were done in the study to understand the long-term trend of rainfall, IOD, and El Nino.
ii. Correlation: finding connection among the data is one of the key theme of scientific studies.
iii. Coefficient of Variance: it is ratio of standard deviation to means. The highest the coefficient value, the greater will be the level of dispersal.

3.4 Software Used

Software used in the study includes Open GrADS , Python, SPSS and Microsoft Excel. The Grid Analysis and Display System (GrADS) is a bilateral desktop tool used to access, manipulate and visualize the earth data. Composite anomalies of weather parameters are drawn using GrADS software. Statistical tools such as correlation, coefficient of variance, standard deviation was done in using SPSS (Statistical Package for the Social Science) software, used for data mining and text analytics application. Linear and scatter plots were drawn using Microsoft Excel.

RESULTS

CHAPTER 4

RESULT

The monsoon rainfall and associated global circulations plays a vital role to sustain the Indian economy. Important statistical parameters like Mean (M), Standard Deviation (SD) and Coefficient of Variance (CV) of south monsoon season of Kerala was computed for the period 1948 to 2019. The mean rainfall of Kerala during June, July, August and September is 617.3mm, 658.1mm, 421.1mm and 258.8 mm respectively. Abnomalities in SWM rainfall performance have a considerable impact on the national economy.

4.1 Kerala Summer Monsoon Rainfall Trend

Trend analysis shows that the SWM rainfall over Kerala shows a significant decreasing trend for the past 120 years (Figure 4.1). The blue line represents the SWM rainfall and the red and green lines gives its linear trend and running average respectively. The mean rainfall of Kerala during SWM season (from 1901 to 2019) is 2022.355mm with a SD of 386.9mm. The CV of rainfall is 19.12%, indicating it is highly stable**.** From 1951 onwards, there was a considerable decline. During the 120-year period, yearly rainfall decreased by 198.13 mm, compared to the normal of 2141.2 mm. This study limited to 1948 -2019, the rainfall as declined to 113.1mm against the normal rainfall of 2014.1mm.

For better understanding, standardized rainfall anomaly index for the period 1901 to 2019 was done. From figure 4.2, it is evident that KSWM rainfall exhibits decreasing trend with standard deviation 386.8mm. The blue color represents normal or wet rainfall years whereas red color indicates dry or deficit rainfall years. The maximum rainfall was reported during 1924 followed by 1961 which was around 3451.3mm and 3229.3mm respectively and the minimum rainfall was received during 1918 which was nearly 1104.3mm. For the last four decades (1981-2019), the number of dry years are greater than wet years.

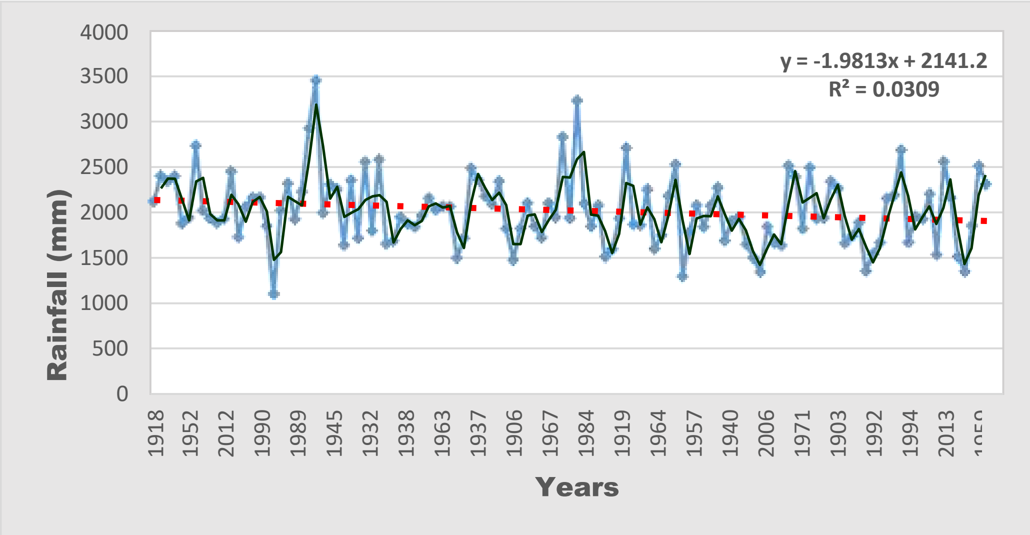

Figure 4.1 Long term variability of Annual Rainfall (1901-2019)

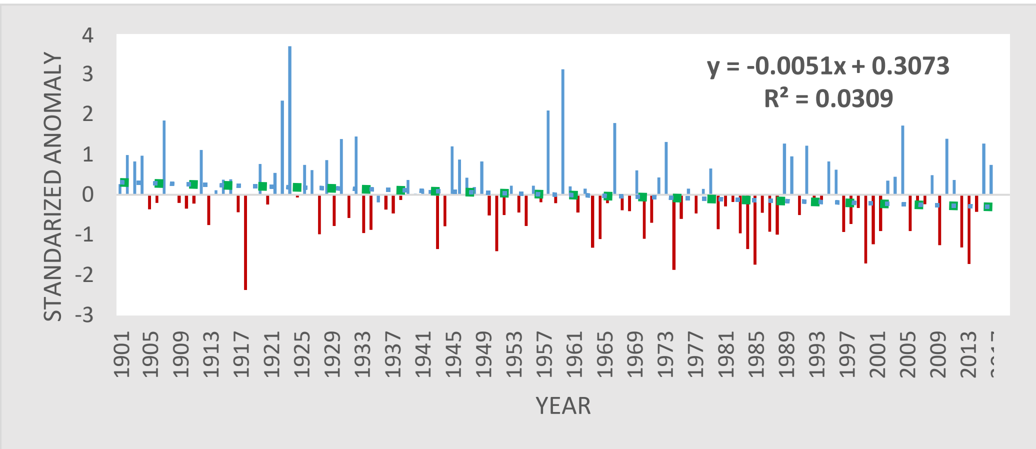

Figure 4.2 Standardized rainfall anomaly for the period 1901-2019

The decline in monsoon rainfall is related with the weakening of southern oscillations, as well as the decline of the meridionial SST gradient from the equatorial region to the South Asia coast thereby weakens the monsoon circulation, reducing moisture transport and decreasing the number of tropical cyclone systems, poleward shift of LLJ, and more warming of the equatorial Indian Ocean than the northern latitudes. (Mooley *et al.,* 1985; Naidu *et al.,* 2009; Rao *et al.,* 2004; Krishna, 2009; Chung and Ramanathan, 2006). In this study, monsoon variability was studied during excess and deficit rainfall years which were identified using percentage departure criteria for the period 1948 to 2019.

4.2 Percentage Departure Analysis

Based on the criteria if the percentage departure is greater or less than 20% it is considered to be an excess or deficit year respectively (Figure 4.3).The blue line represents the percentage departure and the red and green line gives the linear trend and running average respectively. The percentage departure of multiple years has been discovered to have exceeded the threshold criteria. As a result, extreme drought and flood occurrences as a result of climate variability, as well as the rising frequency of drought and flood events in recent decades, notably since 1960, are a major source of concern. From 1948-2019, there are 9 excess (1959, 1968,1975,1991,1994, 2007, 2013, 2018) and 12 deficit (1952, 1965, 1966, 1972, 1976, 1986, 1987, 2002, 2003, 2012, 2015, 2016) rainfall years. Although KSMR shows a decreasing trend in SWM seasons, two recent years (2018 & 2019) exhibit exceptionally heavy rainfall spell which is greater than the annual average of many stations within a short period of time (~ 10 days) during July and August. This may be due to weakening of monsoon low pressure systems and moistening of the tropical troposphere (Hunt and Menon, 2020)

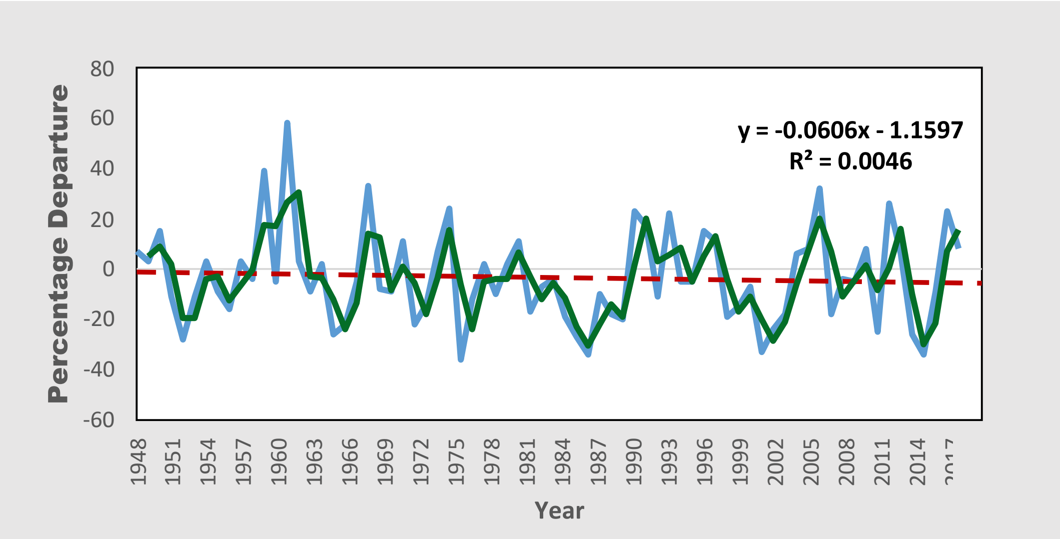

Figure 4.3 Long-term rainfall departure from normal (1948-2019)

Table 4.4 Percentage Departure of Rainfall (1948-2019)

YEAR	ACTUAL RAINFALL (mm)	PERCENTAGE DEPARTURE (%)
1948	2183.7	7
1949	2094.7	3
1950	2342.9	15
1951	1823	-11
1952	1477.7	-28
1953	1825.1	-11
1954	2107.4	3
1955	1849.7	-9
1956	1720.1	-16
1957	2107.4	3
1958	1948.7	-4
1959	2831.2	39
1960	1939.2	-5
1961	3229.3	58
1962	2101.6	3
1963	1848.5	-9
1964	2079.8	2
1965	1508.9	-26
1966	1593.3	-22
1967	1937.3	-5
1968	2711.4	33
1969	1870.6	-8
1970	1860.7	-9
1971	2254.6	11
1972	1596.8	-22
1973	1749.2	-14
1974	2188.2	7
1975	2529.3	24
1976	1297.1	-36
1977	1788.5	-12
1978	2081.1	2
1979	1840.5	-10
1980	2077.5	2
1981	2274.4	11
1982	1689.2	-17
1983	1906.9	-7
1984	1951.6	-4
1985	1650.4	-19
1986	1498.4	-27
1987	1347.2	-34

1988	1845.7	-10
1989	1664.7	-18
1990	1638	-20
1991	2515.6	23
1992	2392.5	17
1993	1823.1	-11
1994	2493	22
1995	1933.2	-5
1996	1938.5	-5
1997	2342.9	15
1998	2263.4	11
1999	1661.9	-19
2000	1739.4	-15
2001	1892	-7
2002	1359	-33
2003	1542.6	-24
2004	1669.5	-18
2005	2157.6	6
2006	2193.8	8
2007	2688.5	32
2008	1670.9	-18
2009	1958.9	-4
2010	1928	-5
2011	2209.1	8
2012	1535.6	-25
2013	2561.2	26
2014	2164.8	6
2015	1514.7	-26
2016	1352.2	-34
2017	1855.9	-9
2018	2515.7	23
2019	2309.8	8.13

Table 4.5 Percentage Departure of Rainfall during Excess and Deficit Years

Excess rainfall years	**Actual rainfall (mm)**	**Percentage departure (%)**
1959	2831	39
1961	3229	59
1968	2711	34
1975	2529	25

Deficit rainfall years	**Actual rainfall (mm)**	**Percentage Departure (%)**
1952	1478	-27
1965	1509	-26
1966	1593	-22

1991	2516	24
1994	2493	23
2007	2689	32
2013	2561	26
2018	2516	24

1972	1597	-21
1976	1297	-36
1986	1498	-26
1987	1347	-34
2002	1359	-33
2003	1543	-24
2012	1536	-24
2015	1515	-25
2016	1352	-33

4.3 Decadal wise annual rainfall trend

To understand the rainfall trend in each decade (1948-1957,1958-1967, 1968-1977,1978-1987, 1988-1997, 1998-2007,2008-2019), decadal analysis were carried out. The red line gives the linear trend in each decade and the blue and green line indicates standardized rainfall anomaly and the running average respectively. From 1948-1957, the rainfall shows a decreasing trend with standard deviation of 244.35mm with one deficit rainfall year and the maximum and minimum rainfall was reported during 1950 and 1952 respectively. Similar declining trend is observed during 1958-1967 and 1968 -1977. In 1958-1967, there are two excess and three deficit years with decrease of 175.3mm rainfall from normal. Maximum rainfall was observed during 1961 and minimum rainfall on 1965. From 1968-1977, there are four wet years and two dry years , maximum rainfall is reported during 1968 followed by 1975. From 1978-1987, a steep decrease in rainfall can be seen with standard deviation of 271.6mm. In this period maximum rainfall and minimum rainfall was noted on 1981 and 1987. From 1988, the rainfall shows an increasing trend in each decades. From 1988-1997, there are two excess and zero deficit years and maximum rainfall is received during 1991 and minimum in 1990. In 1998-2007, there exists one excess and two deficit years and maximum rainfall was reported during 2007.From 2008-2019, the rainfall shows an increasing trend with maximum rainfall reported during 2013 with two excess and three deficit years. Maximum variation in rainfall was seen during 1958-1967.

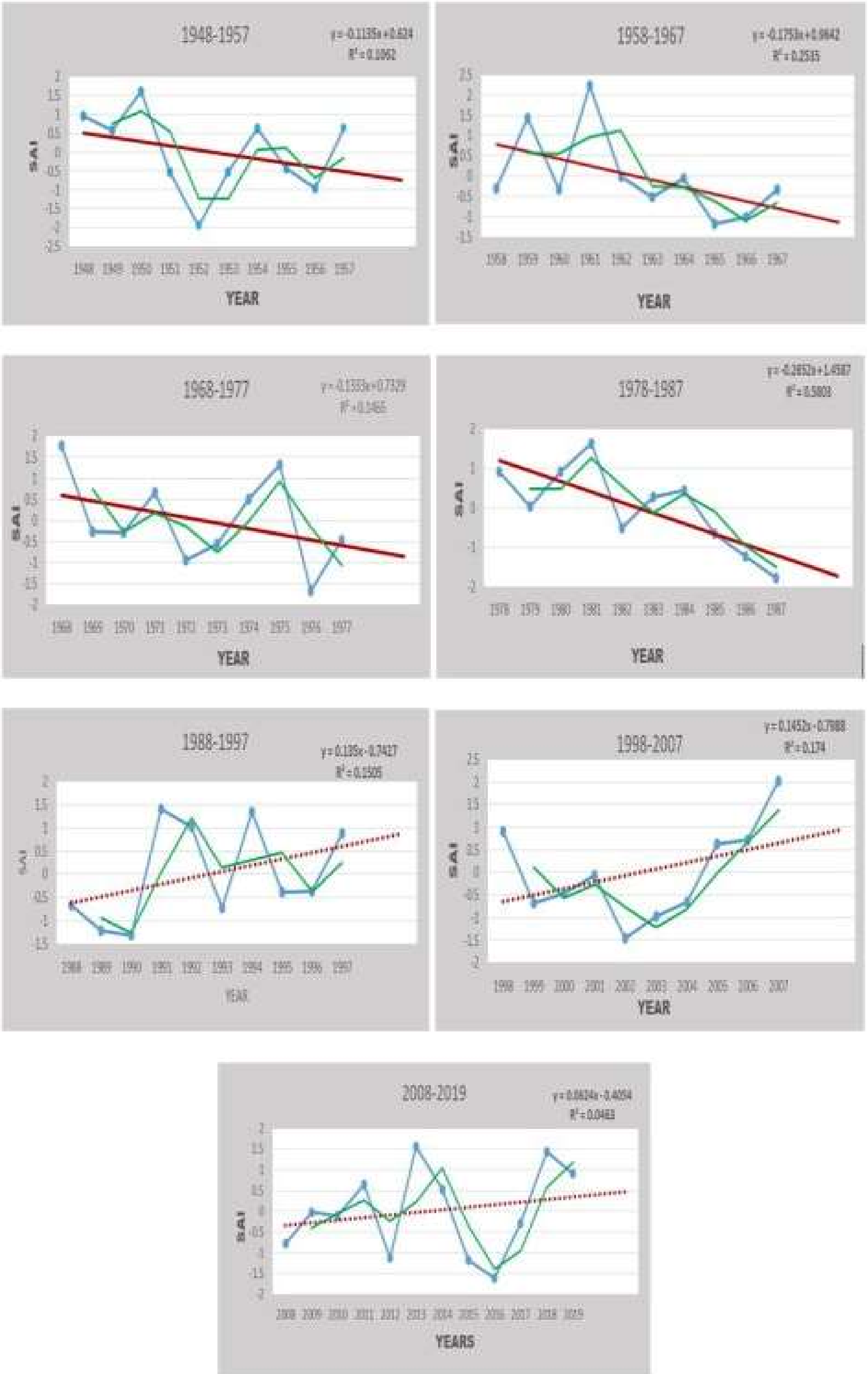

Figure 4.6 Decadal wise rainfall trend from 1948-

Table 4.7 Decade-wise % departure from normal and frequency of excess & deficit rainfall years from 1948 to 2019

Decade	% departure from Normal	No. of excess years	No. of deficit years
1948-1957	-3.4	0	1
1958-1967	3.92	2	3
1968-1977	1.9	2	2
1978-1987	-9.4	0	2
1988-1997	1.7	2	0
1998-2007	5.2	1	2
2008-2019	-2.8	2	3

4.4 Nature of weather parameters during excess & deficit years

I. Air Temperature

Composite anomaly of air temperature for deficit and excess rainfall years for the period 1948 to 2019 were plotted. During excess years, the temperature is maximum over the Pakistan and northwest India (heat lows) between 30-40^{o} N & 65-80^{o}E region than the surrounding ocean. This warming trend may be due to formation more deep clouds. Increased cloud cover may minimize surface warming while speeding up heat within and around the cloud layer (Kumari and Goswami, 2010). Intense heat low act as a suction for moist air along the monsoon trough and result in good monsoon. While during deficit years, this region experience minimum temperature than the surrounding regions i.e., weak heat low, result in deficient or scanty monsoon rainfall. Maximum temperature is observed along the Eastern Ghats.

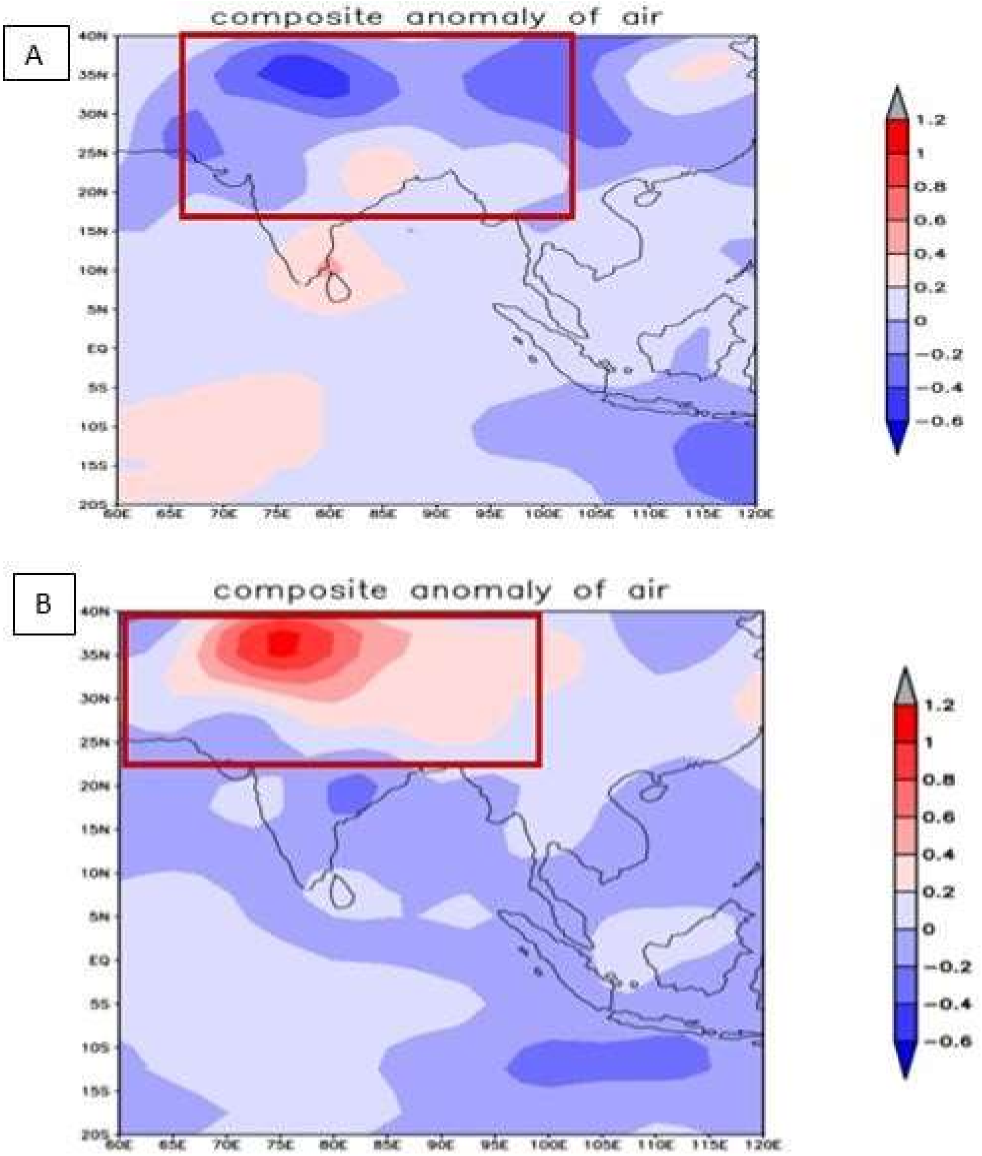

Figure 4.8 Composite Anomaly of Air Temperature during (A) excess years (B) Deficit years

II. OUTGOING LONGWAVE RADIATION (OLR)

OLR data is used as a proxy for rainfall and cloud thickness to interpret precipitation over oceanic regions, with lower values indicating deep and thick clouds. While a greater OLR indicates shallow clouds (with warmer cloud top). During the surplus KSMR years, an OLR dipole structure with a positive pole over the EEIO and a negative pole across the west coast to central peninsular India was discovered. While during deficit years, Positive poles of the OLR oriented throughout southeast to northwest direction across the west coast and central peninsular India.

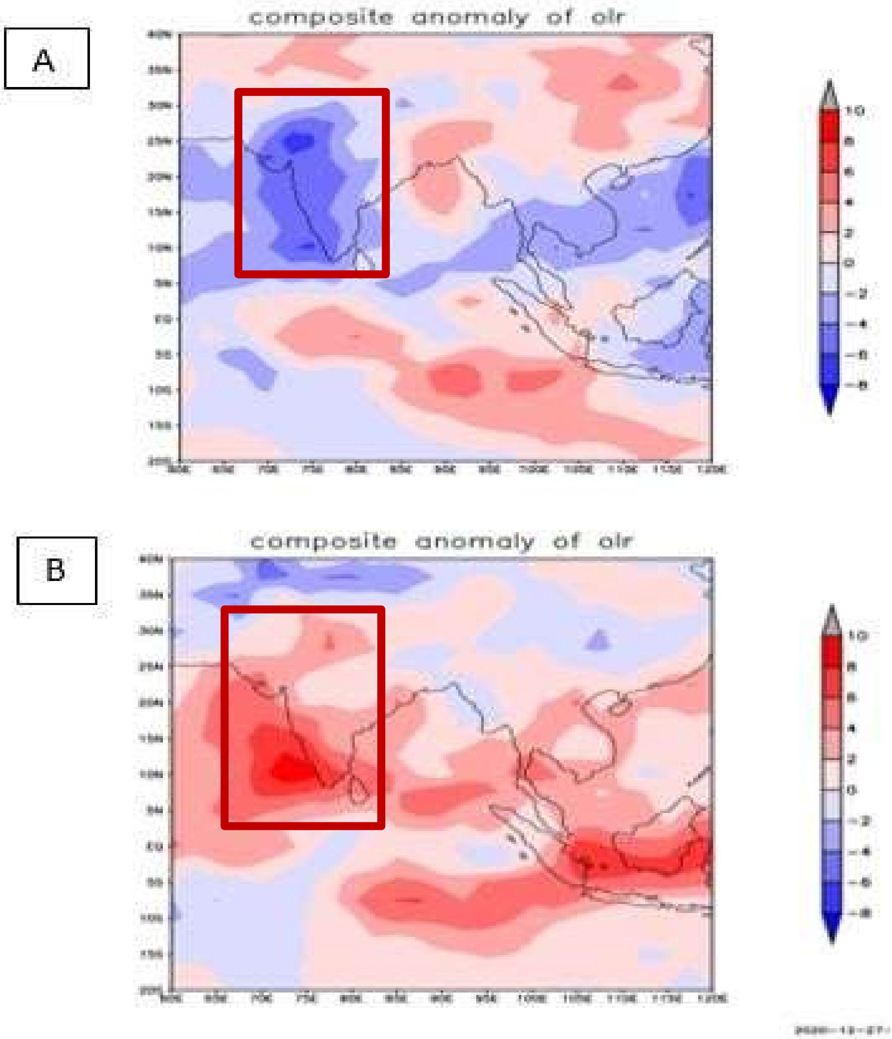

Figure 4.9 Composite Anomaly of OLR during (A) excess years and (B) deficit years

III. VERTICAL VELOCITY

The vertical velocity (omega) related to geopotential because precipitation is associated with large-scale upward motion. The upward motions has negative values and downward motion has a positive value. In excess years, vertical velocity is negative over the northwest region (28ºN-40ºN, 70 ºE-78ºE) and positive over northeast regions (24-31N, 80-105E) whereas during deficit years, positive velocity is over Tibetan Plateau (30-40N, 70-85E)

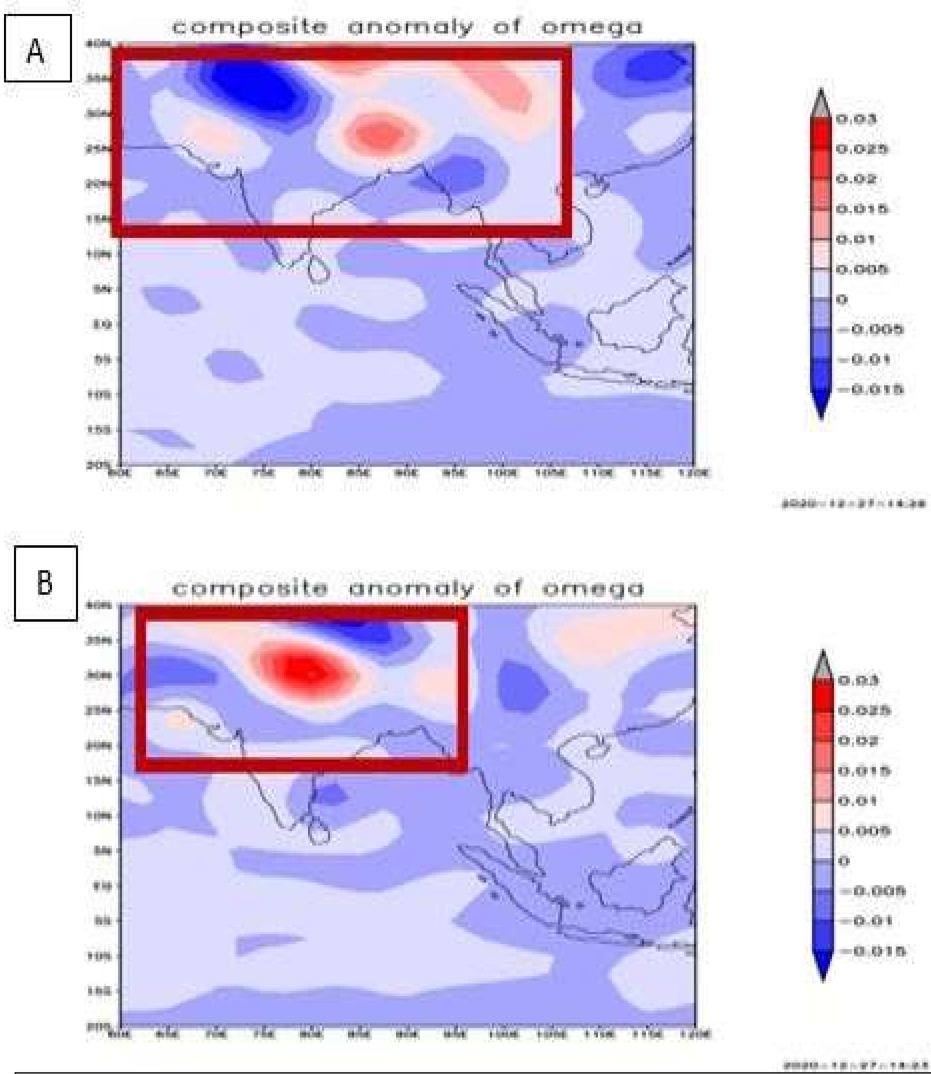

Figure 4.10 Composite Anomaly of vertical velocity during (A) excess years and (B) deficit years

IV. PRECIPITABLE WATER

Precipitable water is the amount of water that is conceivably accessible for precipitation in the atmosphere. During the excess rainfall years, amount precipitable water is high compared to deficit year and higher over 13ºN-30ºN to

60ºE-80ºE and over Arabian Sea and BOB. During deficit period, precipitable water shows negative values over central and southern India than the surrounding regions

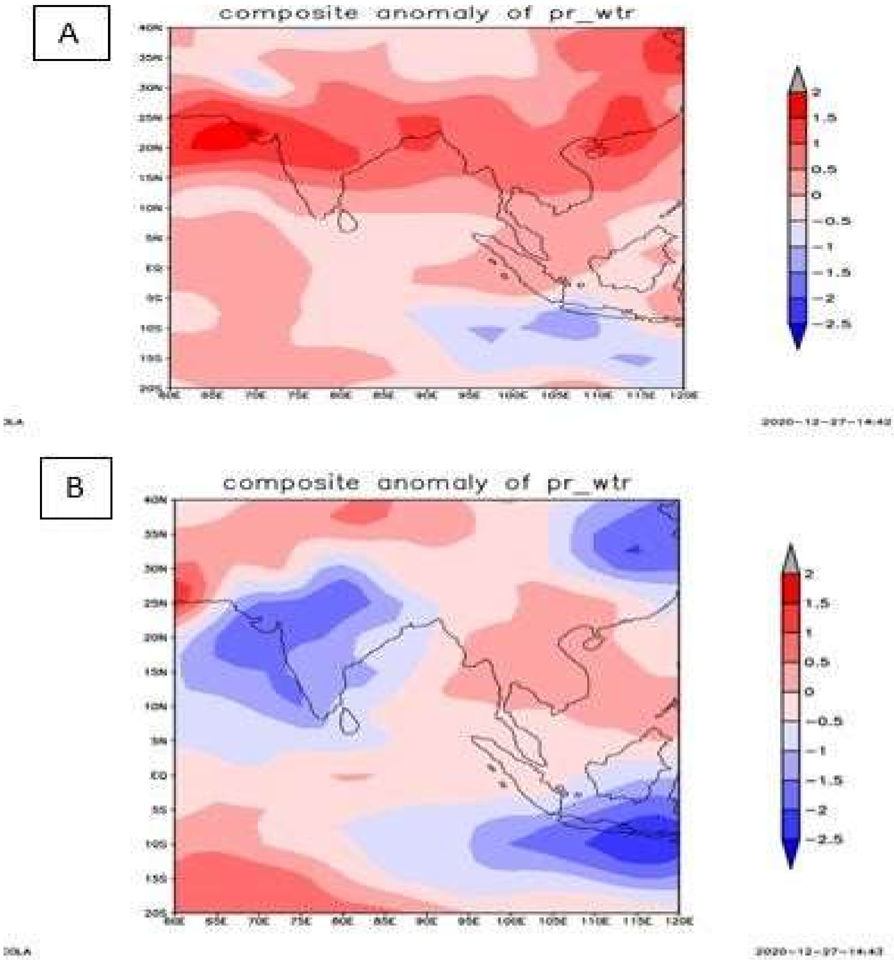

Figure 4.11 Composite Anomaly of pr_wtr during (A) excess years and (B) deficit years

V. SPECIFIC HUMIDITY

From figure 4.12(A) and 4.12 (B), specific humidity is high across Indian subcontinent in excess years whereas it is minimum during deficit years. During excess years, humidity is positive over the peninsular India and are slightly extended towards the west central re gion (10-30ºN, 75-85ºE) while it is low during deficit years.

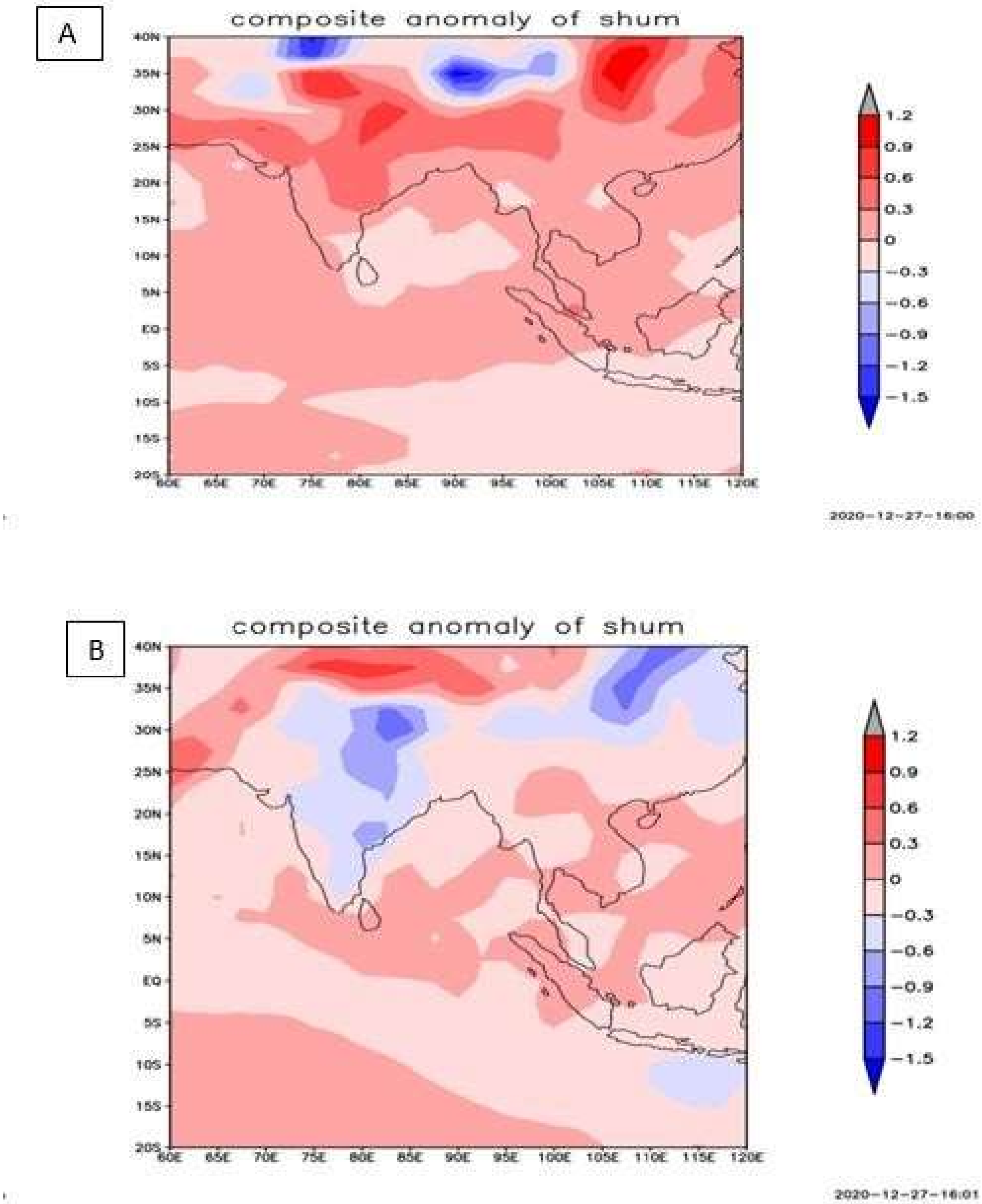

Figure 4.12 Composite Anomaly of Specific Humidity during (A) excess years and (B) deficit years

4.4.5 Sea Level Pressure (SLP)

The intensity of ISMR is controlled by the gradient of sea level pressure (SLP) and the SST distribution over the AS and is further accentuated by BOB. The spatial distribution of SLP is a significant element in the development and maintenance of ISM because it might reflect the sea-land thermal differential (Liu *et al.,* 2019).During excess years, SLP is maximum over the ocean compared to landmass except a small patches of high pressure is formed over central northeast region. During deficit years, the landmass have maximum pressure than surrounding water bodies. High pressure patched are spread across the northeast and west regions. From the figure it is understood that SLP gradient is maximum during deficit years than excess years which implies temperature contrast between land and sea is maximum during excess years than in deficit years.

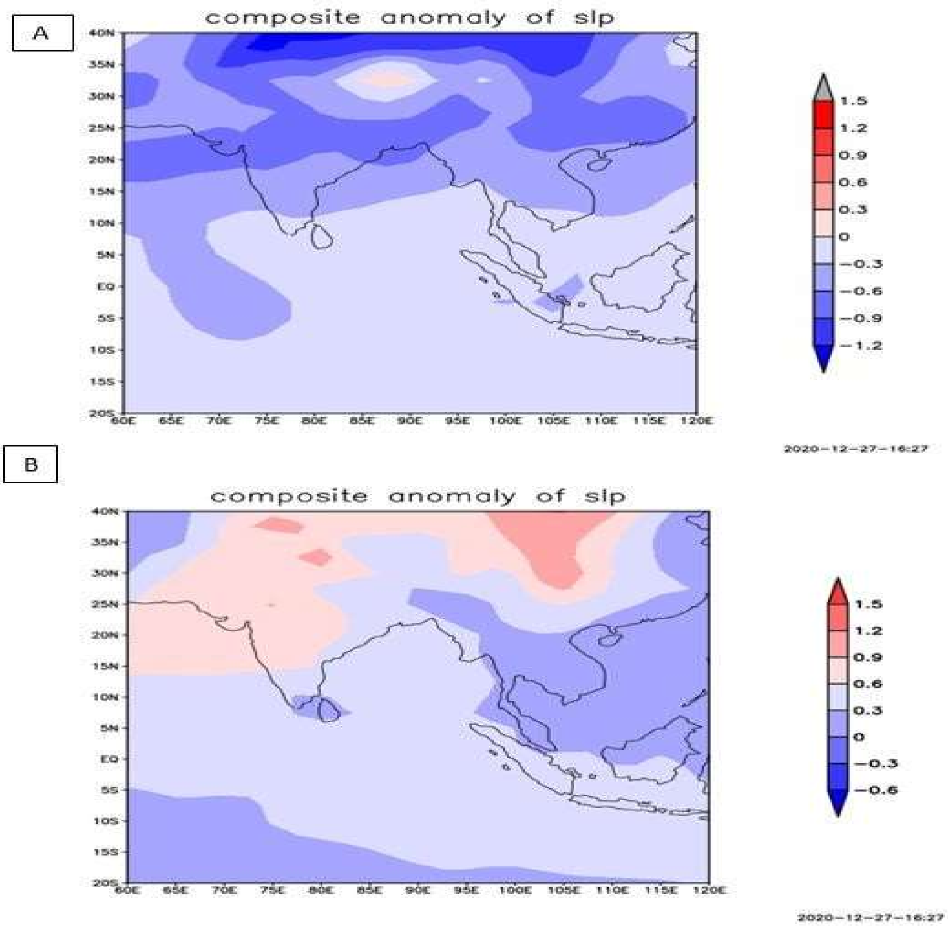

Figure 4.13 Composite Anomaly of Sea Level Pressure during (A) excess years and (B) deficit years

4.5 PHYSICAL MECHANISMS

Kerala summer monsoon rainfall is alter by various atmospheric and oceanic phenomenon. In this section, we discuss about the physical mechanisms that contributes to excess and deficit rainfall conditions.

4.5.1 Indian Ocean Warming

A steady increase in sea surface temperature over Indian Ocean can be visible from 1948- 2019. IO is continually warming, according to several studies, and its warm pool is growing, notably in recent decades. This IO warming can be linked to global warming, sea surface height anomaly and ENSO (Roxy *et.al.* 2014; Rao *et al.,* 2012) thereby reduces land-sea temperature contrast. The impact of IO warming on regional climate is modest. The increased warming of the Indian Ocean enhances rainfall over the Indian Ocean, but the weakening of the Hadley circulation reduces convection over the South Asia (Saha *et al*., 2014 and Roxy *et al*., 2015). According to recent studies, the tropical Indian Ocean's sea surface temperature climbed by 1ºC on average between 1951 and 2015. The ocean heat content in the upper 700m exhibits an increasing trend from 1955. Currently the temperature rise as reached up to 4.2% from the mean value.

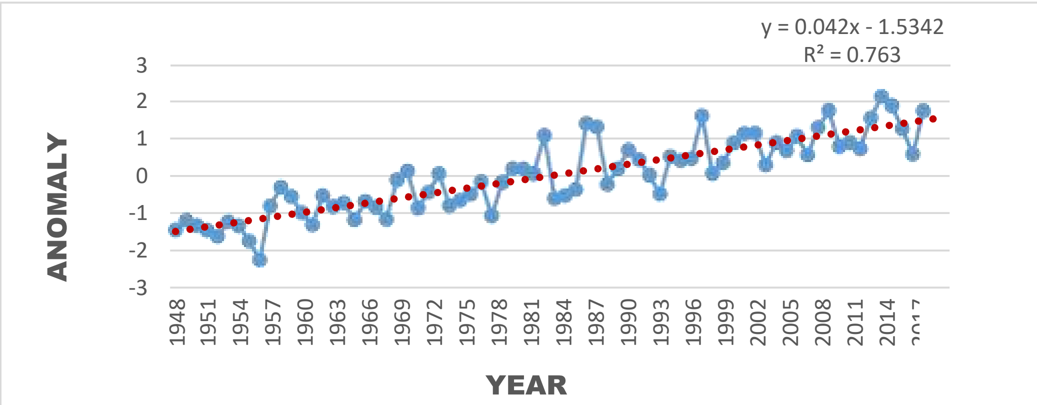

Figure 4.14 Long term SST temperature over the Indian Ocean

4.5.2 Vertically Integrated Moisture Flux and Wind

The decrease (increase) in monsoon rainfall depend upon the moisture transfer from the ocean. We resolve the problem by adjusting land mass and monitoring changes in the vertically integrated moisture flux (VIMF) and wind patterns across the Indian Ocean. If moisture flux is positive, it favors monsoon rainfall. During excess rainfall years (figure 4.15(A)), moisture flux is positive over the south of 14ºN around Kerala i.e., moisture convergence is high over the Kerala region which results in excess rain but in case deficit years (figure 4.15(B)), positive moisture flux (convergence) was seen in northward side Kerala and negative flux is seen Bay of Bengal (BOB) because rainfall over Kerala is influence by the low pressure or depression systems in BOB.

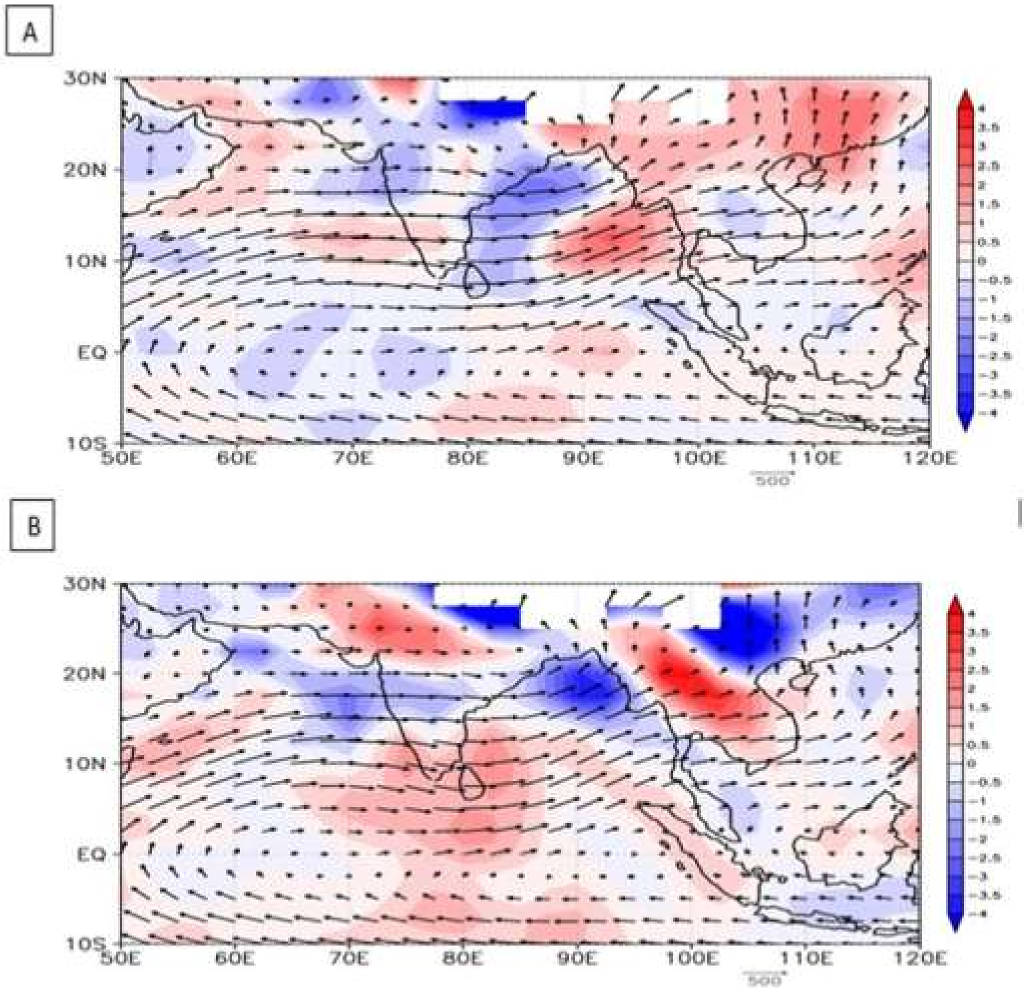

Figure 4.15 Composite of VIMF during (A) excess rainfall years and (B) deficit rainfall years

During excess rainfall years, wind speed (12knots) and cross equatorial flow (figure 4.16(A)) is maximum compared to deficit year. It is because the Low level Jet streams during excess years are stronger compared to deficit years which strength the wind and cross equatorial flow. This low level wind is an important factor that contributes more moisture flux over the Kerala coast. From figure 4.16(C), it is seen that the difference between the wind at 850hPa during excess and deficit years is 3 knots even though it creates huge impact over monsoon rainfall.

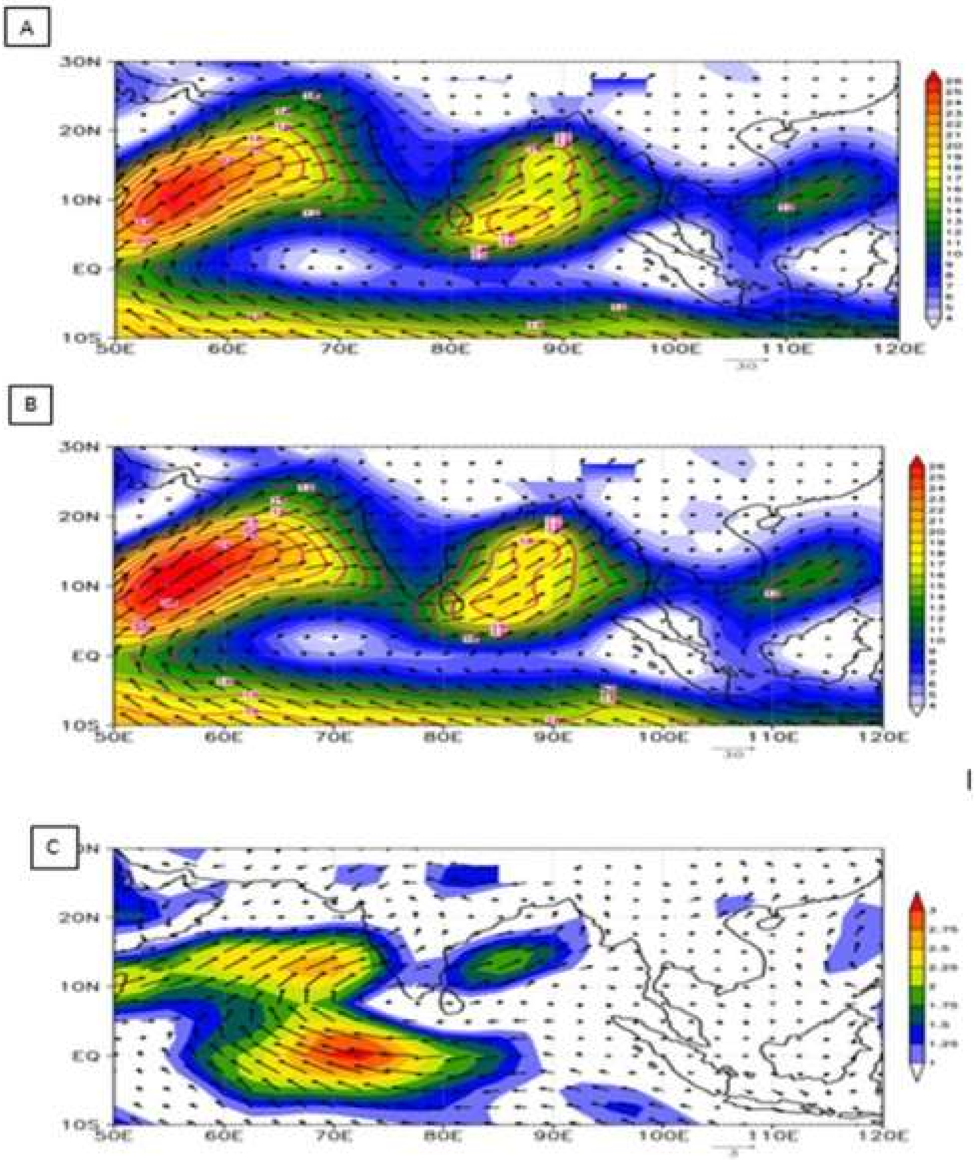

Figure 4.16 (b) 850 hPa wind composite during (A) excess, (B) deficit rainfall years and (C) (Excess – Deficit)

4.5.3 Tropospheric Temperature

It the average temperature gradient over 700-200hPa and this gradient force drives the LLJ. During the excess years, tropospheric temperature gradient is maximum over the land surface than the ocean. This gradient is formed as a result of diabatic heating produced as result of Tibetan high (sensible heat transport) and clouds formation during monsoon (latent heat). During deficit years, the tropospheric temperature gradient between the land and ocean are comparatively weak.

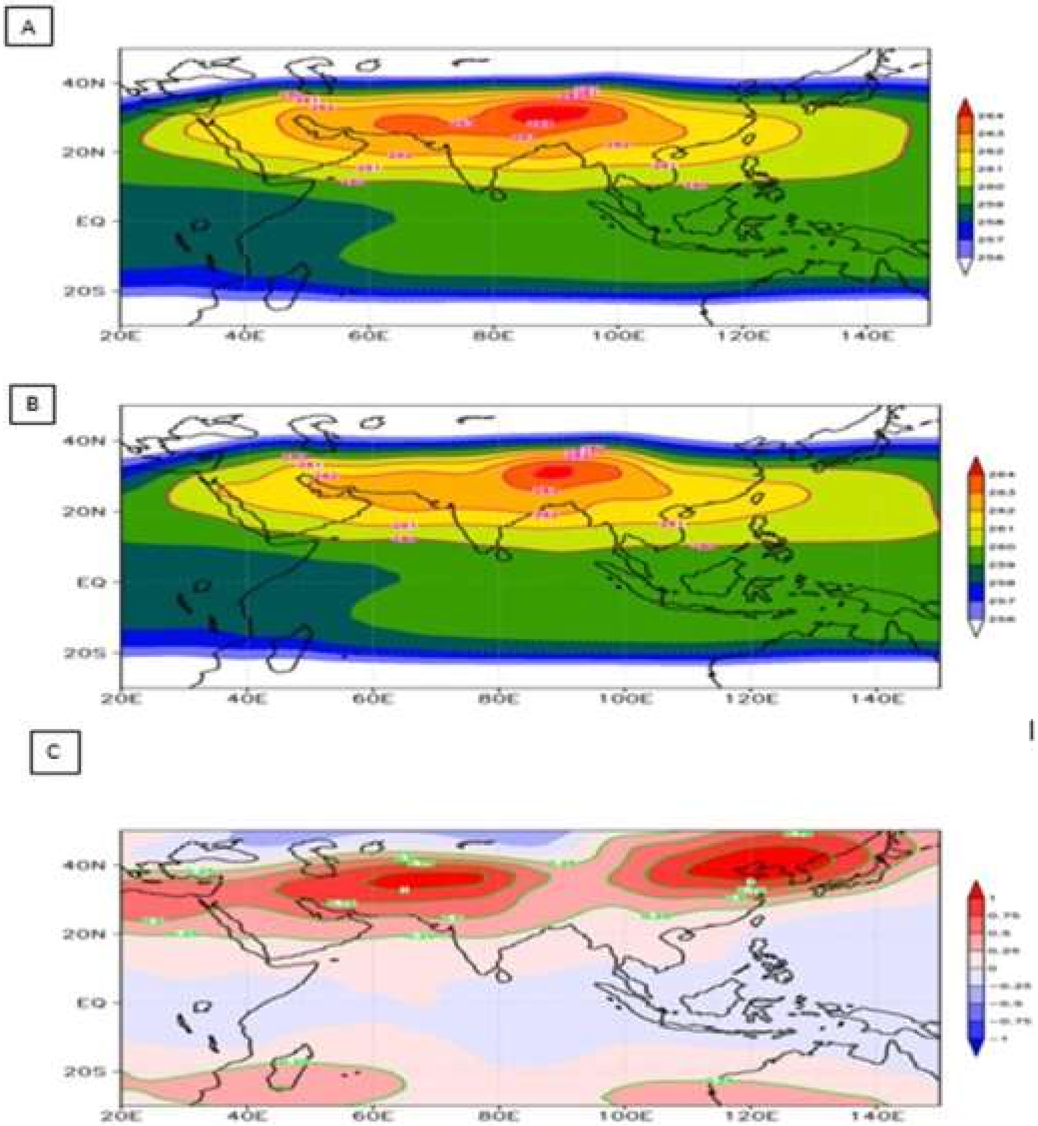

Figure 4.17 Tropospheric Temperature (Average 700-200) composite (A) during excess years (B) during deficit years and (C) (Excess- Deficit) TT difference (Excess – Deficit)

When we take the difference (1ºC), it is visible that the temperature is higher in between 20-40 N and a cooling is seen near the equator. When the troposphere temperature is high, the temperature gradient between the land and ocean increases which strengths the wind speed thereby moisture flux also increases.

4.5.4 Zonal Wind Shear

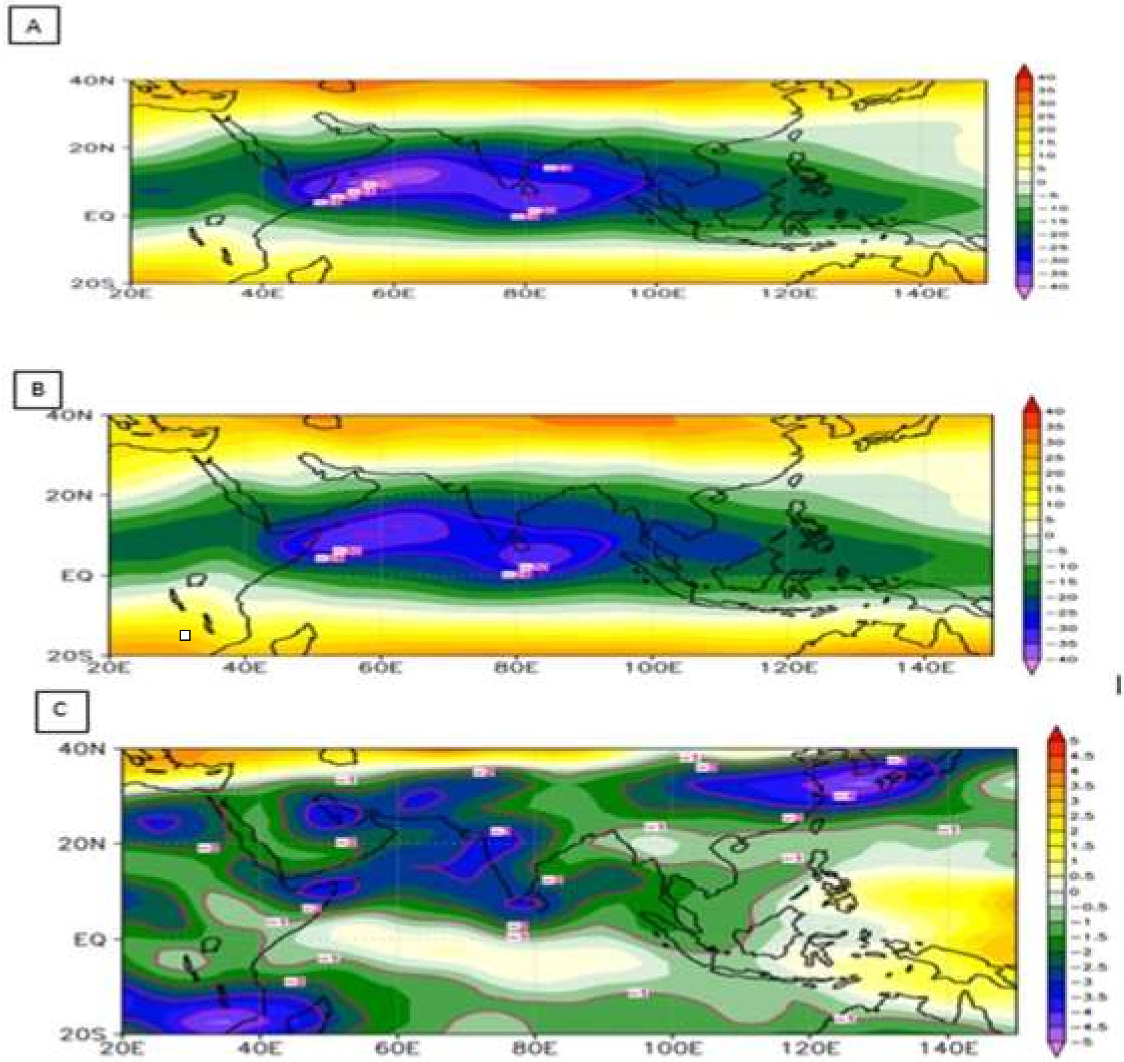

Figure 4.18 Zonal Wind Shear (U200-U850) composite (A) during excess years (B) deficit years (C) (Excess - Deficit)

Monsoon activity is favored by increased wind shear particularly the easterly shear. During excess rainfall years, easterly shear is maximum over the southern part of India (0º-20º N) it's because the LLJ split into two over the Arabian Sea, one heading southeast toward Sri Lanka and the other travelling east through peninsular India. During excess years, former LLJ branch is dominant whereas it is weaker during deficit years and can be seen from the figures 4.18(C). .

4.6 INFLUENCE OF ENSO and IOD

The ENSO and IOD events are generally defined by the Nino 3.4 index and the Dipole Mode Index (DMI). Nino 3.4 is "a measure of how anomalous warm or cool the central to eastern equatorial pacific ocean is compared to normal" (NOAA, 2019) and DMI is anomalous SST gradient between western EIO and southern EIO (BOM, 2020). Time series of Nino 3.4 Index and DMI are shown below (Figure 4.19 and 4.20). El Nino occurrences are represented by positive values (Figure 4.19) while La Nina events are illustrated by negative values. From the graphs it evident that El Nino events shows a constant trend with time.

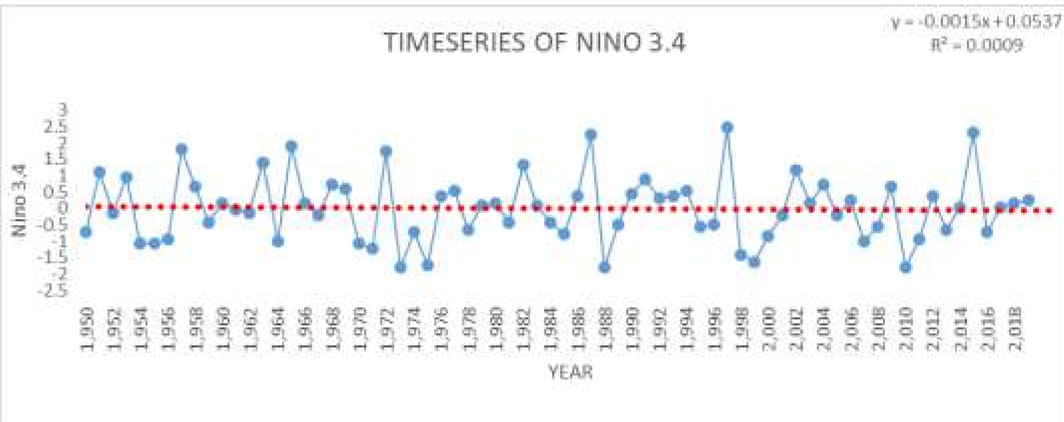

Figure 4.19 Time series 0f Nino 3.4 (1950-2019)

From figure 4.20, it can be seen that IOD events are increasing with time. The processes that triggers IOD events are not completely understood. The role of off-Sumatra anomalous south easterlies (Saji *et al.,* 1999), EIO atmospheric pressure variability (Gualdi *et al.,* 2003), shifts in wind patterns in relation to Pacific ENSO and Indian Monsoon (Annamalai *et al.,* 2003), AS ocean conditions with respect to Indian monsoon (Prasad and McClean, 2004), and the influence of the southern extra tropical region (Lau and Nath, 2003), influence of southern IO trade wind (Hastenrath and Polzin, 2004), influence of ISO- Madden Julian Oscillations originating from the tropical IO, Indian ocean warming (Cai *et al.*, 2013 ; Zheng, 2019) and increased number of tropical cyclones in BOB.

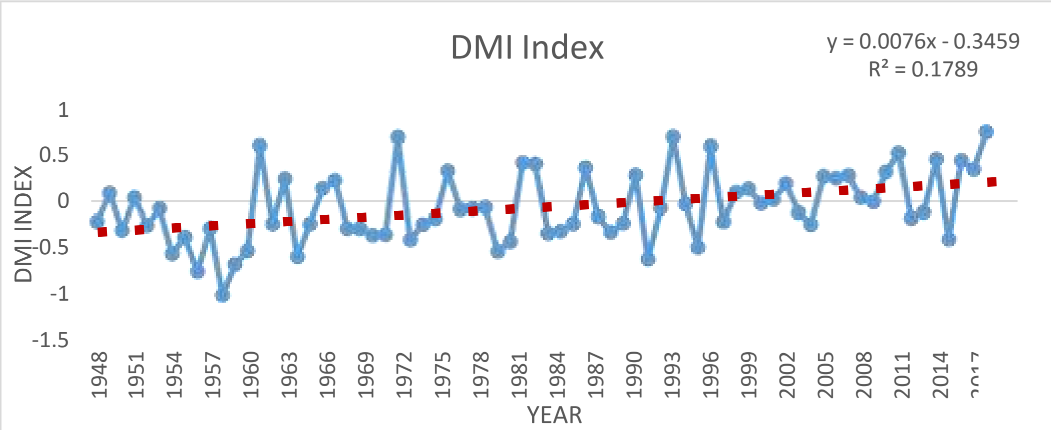

Figure 4.20. Time series 0f DMI (1948-2019)

We then examined the impact of the pacific SST pattern on seasonal rainfall in Kerala. Due to the lack of reliable SST observations prior to 1950, this study is limited to the years 1961–2016. El Nino, La Nina, and Neutral years are determined using NOAA ERSST data sets and the Nino 3.4 SST anomaly. To identify the effect of El Nino over KSMR, scatter plot was drawn. The red, blue and green dots represented as El Nino, Neutral and La Nina years**.** From this it is seen that El Nino events have no significant relation with deficit KSMR.

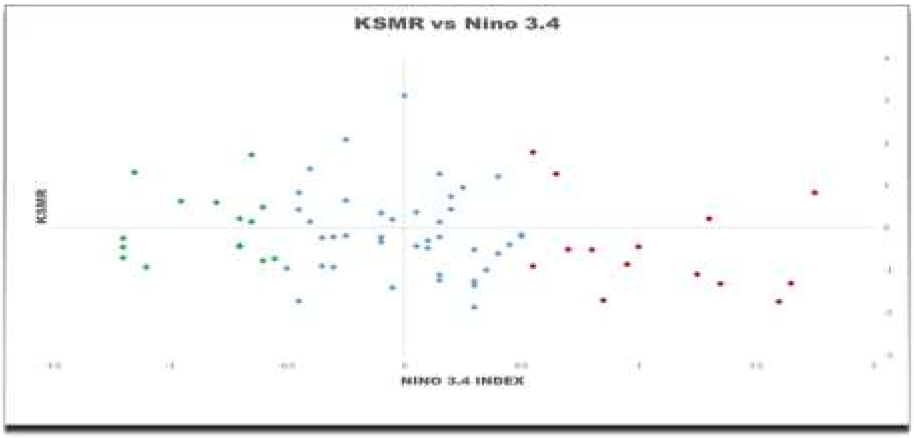

Figure 4.21 Scatter plot to understand the relation of El Nino and KSMR

Table 4.22 Tabular Representation of the scatter plot

	El Nino	La Nina	Neutral
Total	14 (20%)	15 (21%)	41 (51%)
No of excess rainfall years	2	4	5
No of deficit rainfall years	5	0	7
No of normal rainfall years	7	13	29

Also the relationship between KSMR with DMI hold a positive correlation with each other. From the figure 4.23, the red, blue and green dots represented as positive, neutral and negative IOD years. From 1870 to 2020, there are twenty two IOD events are present, out of them there are six positive and sixteen negative IOD events noted. From 4.23, it is seen that IOD and KSMR relationship is strengthening with time.

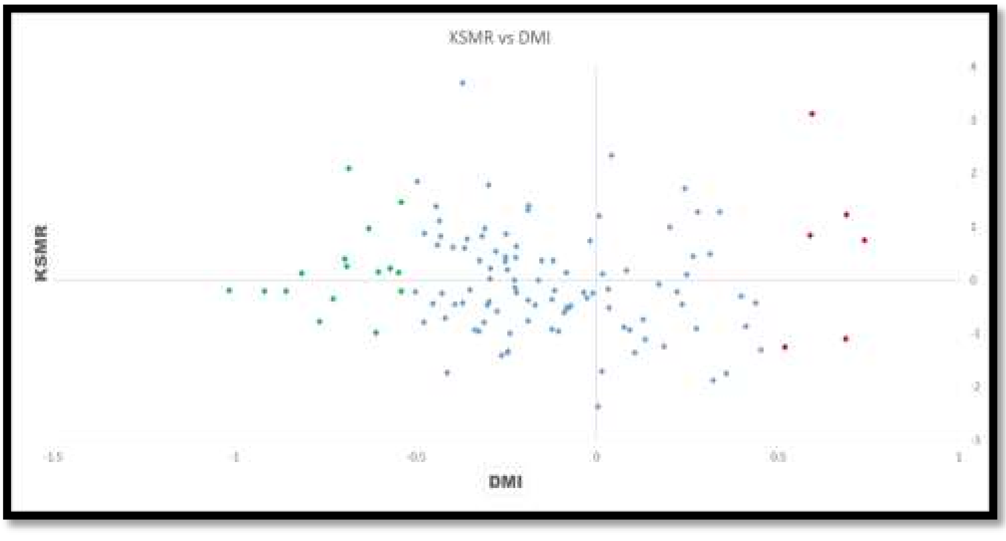

Figure 4.23 Scatter plot to understand the relation of IOD and KSMR

Table 4.24 Tabular Representation of the scatter plot

	POSTIVE IOD	NEGATIVE IOD	NEUTRAL
Total	6 (5%)	16 (13%)	97 (82%)
No of excess rainfall years	2	2	14
No of deficit rainfall years	2	0	11
No of normal rainfall years	2	14	72

Table 4.25 ENSO and IOD intensity levels during years with excess and deficit conditions.

Excess years	Percentage of departure (%)	Nino 3.4	DMI	Strength of ENSO	Strength of IOD
1959	39	-0.25	-0.68	-	Negative IOD
1961	59	0	0.59	-	Positive IOD
1968	34	0.55	-0.29	El Nino	-
1975	25	-1.15	-0.19	La Nina	-
1991	24	0.65	0.28	El Nino	-
1994	23	0.4	0.68	-	Positive IOD
2007	32	-0.65	0.24	La Nina	-
2013	26	-0.4	-0.18	-	-
2018	24	0.15	0.34	-	-
Deficit years	**Percentage of departure (%)**	**Nino 3.4**	**DMI**	**Strength of ENSO**	**Strength of IOD**
1952	-27	-0.05	-0.26	-	-
1965	-26	1.35	-0.244	El Nino	-
1966	-22	0.15	0.134	-	-
1972	-21	1.25	0.68	El Nino	Positive IOD
1976	-36	0.3	0.32	El Nino	-
1986	-26	0.3	-0.24	-	-

1987	-34	1.6	0.35	El Nino	-
2002	-33	0.85	0.014	El Nino	-
2003	-24	0.15	0.18	-	-
2012	-24	0.3	0.52	-	Positive IOD
2015	-25	1.65	0.5	-	Positive IOD
2016	-33	-0.45	-0.41	-	-

4.7 Performance of monthly rainfall

The long term mean and SD of Kerala during June, July, August and September rainfall are 617.3 mm (189.9mm), 658.1mm (217.7mm), 424.1mm (135.7mm) and 258.8 mm (134.06mm) respectively. From 1948-2020, June and July rainfall follows a decreasing trend while August and September rainfall shows an increasing trend with time. During the last 30 years, highest rainfall of June (1096.1mm), July(970.5mm), August(947.6mm), September (529.8mm) are received in 1991,1997,2019 and 2020 respectively.

When we take the cumulative mean rainfall for the 1960-1989 and 1990 to 2019 (figure 4.30), it is noticeable that rainfall received during August and September shows an increase from former period while July rainfall shows a decreasing nature. Kothawale et.al (2013) noted that 50% of June and July deficit rainfall are due to presence or decaying of El Nino events (Boschat and Terray, 2012). After an El Nino incident, the majority of MOK is delayed, resulting in below-normal rainfall in June (Joseph et al., 1994), whereas La Nina years have increased rainfall in August and September. During the El Nino season, August and September rainfall performs better than June and July rain (Boschat and Terray ,2012). It is because the onset of El Nino causes the monsoon wind to diminish and warm SST anomalies in the west EIO leads to shortfall in monsoon rainfall in June and July. The IOD event is another factor that determines monthly rainfall.The positive and negative phase of IOD mode and warmer/ colder SST in the western pole of IO during April-May influences the monthly rainfall.

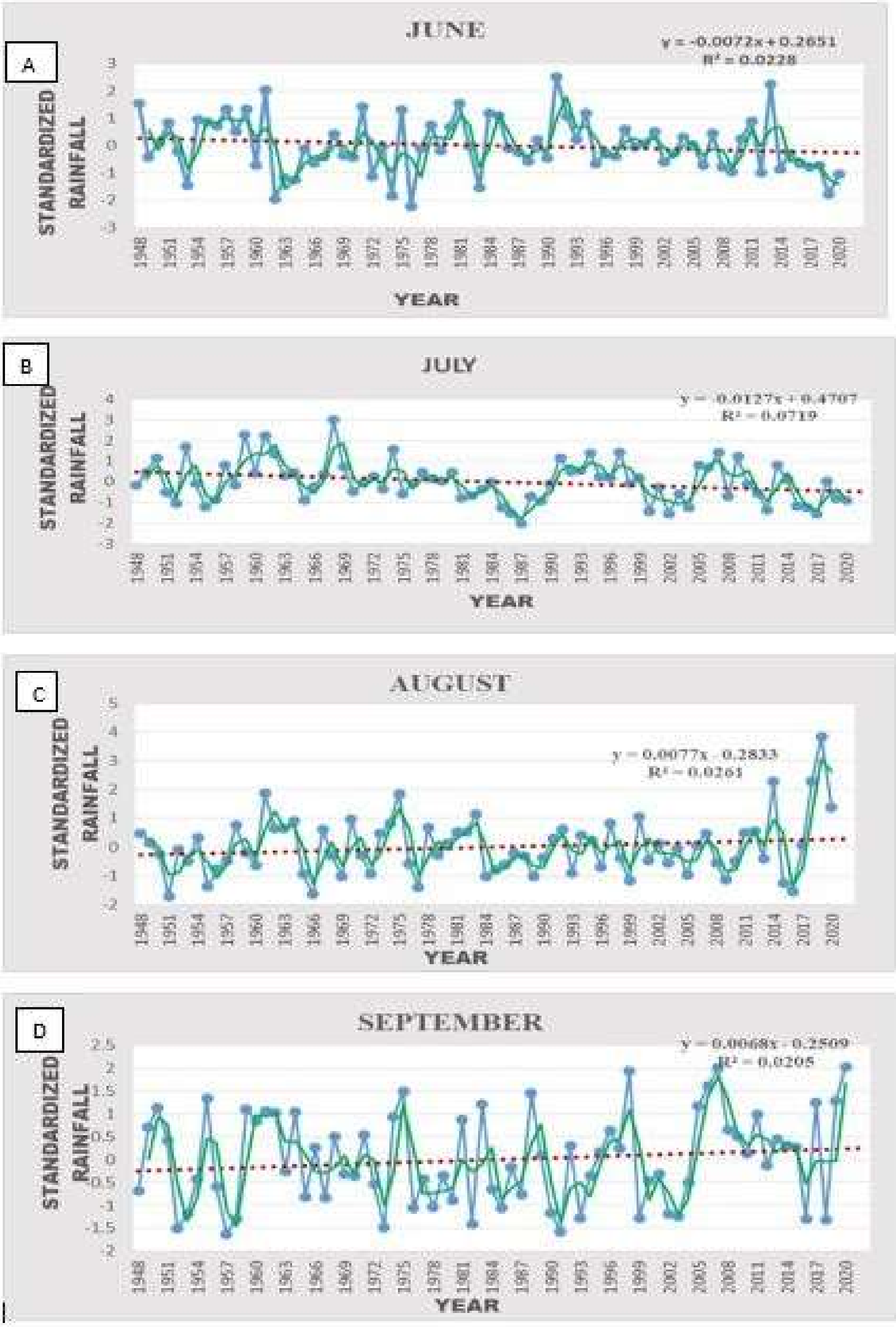

Figure 4. 26 Long term variability of rainfall during (A) June , (B) July, (C) August and (D) September

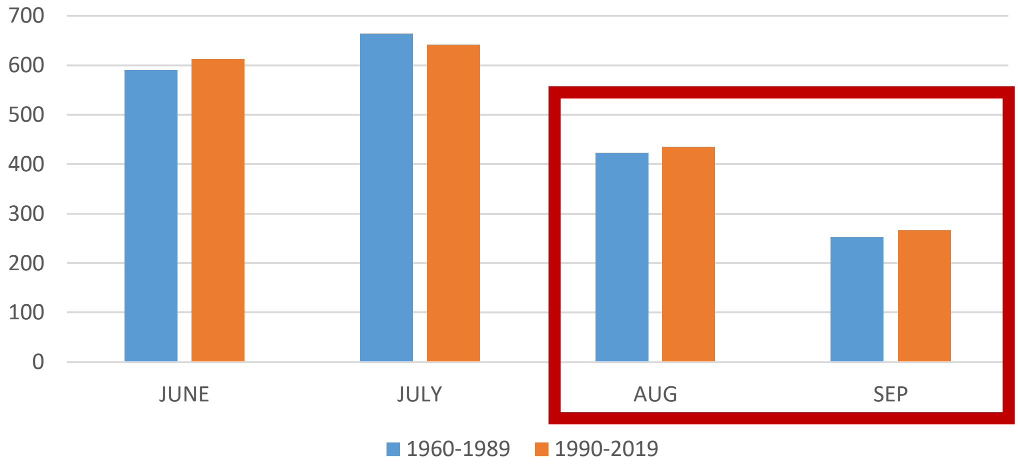

Figure 4.27 cumulative rainfall during 1960-1989 and 1990 -2019

4.7 IOD and Monthly Rainfall

The relationship between monthly rainfalls with IOD was also studied. Rainfall received during June and July shows a significant negative correlation with IOD while a positive relationship exist during August and September. It is because the dipole event normally occurs in May, and the transition to the dipole phase is completed by the end of June/early July, and the IOD index get intensifies from July and attain its peak during October – November due to anomalous strengthening of SST anomaly over EEIO. So we can conclude that IOD influence is more prominent in the second half of the monsoon period (August and September) and KSMR is modulated towards the second half of the monsoon season**.**

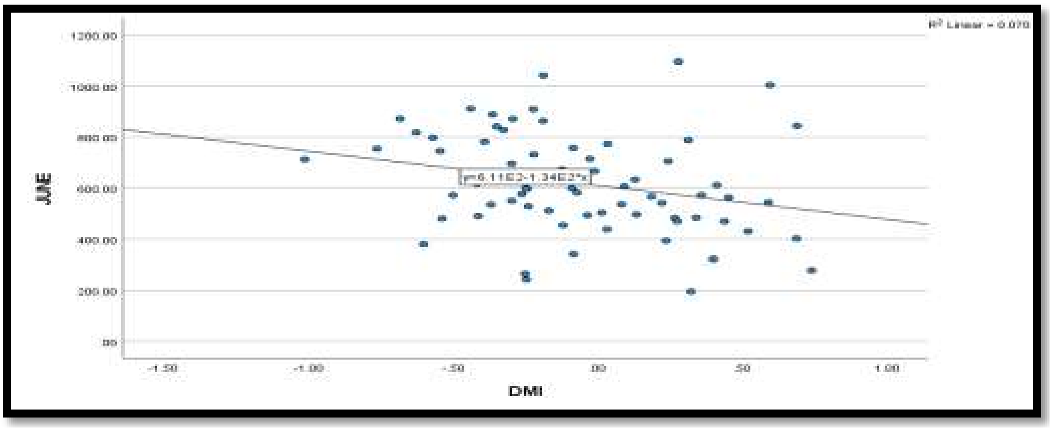

Figure 4.28. Relation between JUNE rainfall and DMI

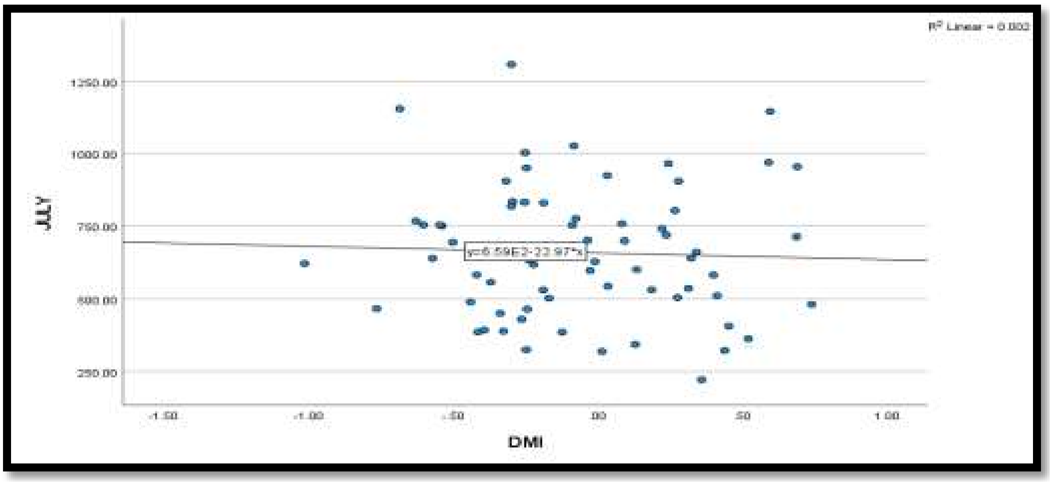

Figure 4.29. Relation between JULY rainfall and DMI

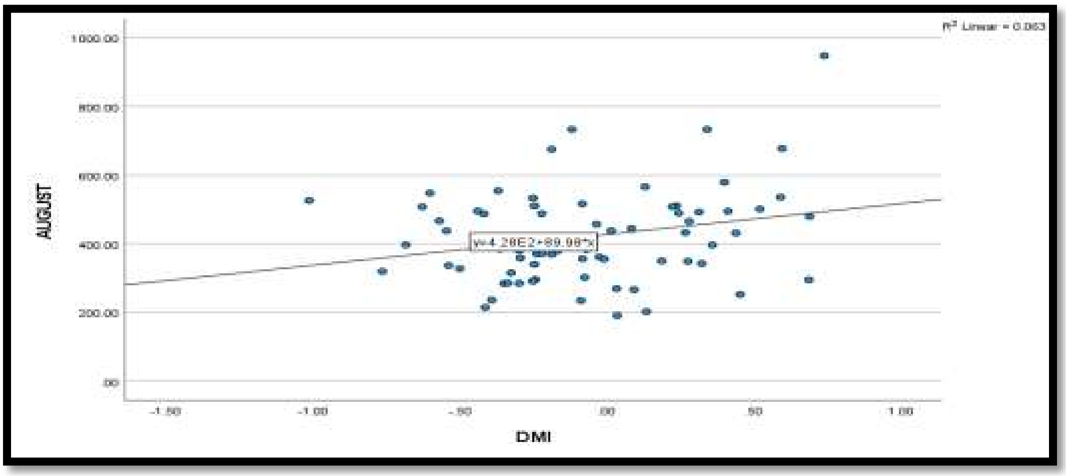

Figure 4.30 Relation between August rainfall and DMI

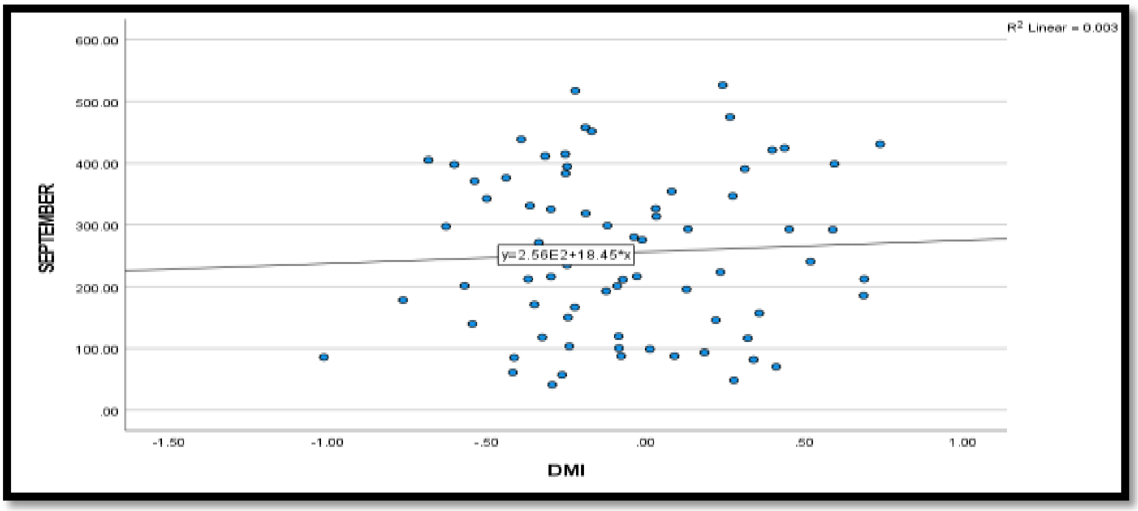

Figure 4.31 Relation between September rainfall and DMI

DISCUSSION

CHAPTER 5

DISCUSSION

Kerala, a state lies in the southwestern coast of peninsular India receives 65-70 percent of its annual rainfall during the south west monsoon season, with the remainder falling during the pre-monsoon and post-monsoon seasons. The annual rainfall of Kerala is ~300cm which is three time greater than the yearly precipitation of India and it's portray a crucial role in agriculture production, power generation and industrial production. The variation in rainfall is associated with prevailing weather phenomenon and monsoon disturbances. Monsoon circulation is influenced by regional and global scale phenomena such as the monsoon trough, off-shore troughs along India's west coast, monsoon lows and depressions movement, upper/mid-tropospheric cyclonic circulations (Rao, 1976), the interaction of the Western Ghats (WG) and monsoon air flow (Sahany *et al.,* 2010; Guhathakurta *et al*., 2014), El Nino, La Nina , Indian Ocean Dipole etc.

5.1 Rainfall trend (1948-2019)

Based on historical rainfall trend, monsoon is drying up and it is much prominent over Kerala and the Western Ghats (Krishnan *et al*., 2016). From figure 4.1.1, it was observed that KSMR rainfall follows a declining trend and the possible reason for this reductions are declining strength of Tropical Easterly Jet (TEJ), reduced number of monsoon depression, reduction in the low level monsoonal flow, weakening of monsoon westerlies and increased frequency of low pressure systems (Sathyamoorthy, 2005; Rao *et al*., 2004; Krishnan *et al*., 2012; Rajendran *et al.,* 2012, Ajayamohan *et al*., 2010) collectively weakens monsoonal synoptic activity. Fletcher *et al.,* (2018) concluded that majority of rainfall over Kerala is influenced by the moisture flux over Western Ghats. Hunt et.al (2020) identified that moisture flux over Kerala shows a slight decrease due to weakening of monsoon depression but in contrast a significant increased flux was also visible due to large rise in tropospheric humidity. Abhilash *et al.,* (2018) noted that over the peninsular Indian, westerlies are weakening while it is increasing over the equatorial Indian Ocean. Also the upper level easterlies shows a declining nature over the peninsula. This implies most of dynamic factors that contributes monsoon rain are weakening over the Indian peninsular region.

An increasing or decreasing trend in heavy or extreme rainfall leads to flood or drought condition which affects livelihood of vulnerable community. Francis and Gadgil, (2006) concluded that unusual events are expected to elevate in the future and these events contributes a significant part in total monsoon rainfall and its variability (Ramesh and Goswami, 2007).By using percentage departure criteria, excess and deficit rainfall years were identified. Due to the existence and position of an offshore trough over the west coast of India, the development of a depression over the BOB, and its north-west propagation towards central India, results in significant rainfall across Kerala (Rao, 1976; Francis and Gadgil, 2006). Deficit rainfall during summer monsoon are mostly compensated during post monsoon season thus bring annual rainfall to be normal.

5.2 PHYSICAL MECHANISM

The key components of Asian Summer Monsoon are tropical easterly jet stream (TEJ) **(**Koteswaram, 1958), low-level jet stream (LLJ) (Joseph and Raman, 1966; Findlater, 1969) and Tibetan anti-cyclone. An attempt were made to understand the link exist between Kerala rainfall variability with mean climate state and remote tele-connections. The dynamic components that influences the summer monsoon rainfall over the peninsular India are Low level westerly jet stream (LLJ) and Upper level tropical easterly jet streams (TEJ) (Joseph and Simon, 2005; Sathyamoorthy, 2005). LLJ arises as result of thermal gradient between Asian landmass and surrounding ocean (Hoskins and Rodwell, 1995), along with a cross equatorial current presiding from the southern Indian Ocean carries large moisture to the central Arabian Sea (Krishnamurthi and Bhalme, 1976; Ordonez et al.,, 2012**).** Thus the position, strength and severity of LLJ and TEJ changes the amount of rain that falls on daily to seasonal basis as it controls the moisture transport over Indian landmass (Sandeep and Ajaymohan, 2015).

Between excess and deficit years, we examine changes in wind distribution at 850 and 200hPa, moisture flux and tropospheric temperature. During south west monsoon season, southwesterly wind carries moisture from western Arabian Sea results in rainfall over Indian subcontinents and these wind exhibits patterns of convergence and divergence over the oceans. The converging wind triggers convection activities leads to moisture increase (evaporation) in the atmosphere thereby enhanced regional precipitation. During excess years, convergence (positive flux) was much prominent over south of 14^{0}N near Kerala and Indochina peninsular regions. Also large regions of eastern BOB particularly south of Sumatra region experience convergence while a negative flux was observed over northern Bay of Bengal. During deficit years, the positive fluxes are present in northward latitude of Kerala and eastern BOB and negative flux was seen on western BOB. BOB plays crucial role in KSMR. Praveen *et al.,* (2015) pointed out that low pressure system or monsoon depression over the BOB influence the monsoon rainfall that is 60% of monsoon precipitations were due to these systems.

When we consider the wind at 850hPa, it was found that during excess years the low levels wind strengthens (~12 knot) over the Indian peninsular and equatorial IO regions whereas it is weakens during deficit years. This strengthening of wind is associated with the low level jet streams (LLJ). The confluence of LLJ with the orography of Western Ghats produce heavy rainfall over west coast of India. According to Ruchtich *et al*., (2015), strong LLJ conditions results in normal/good monsoon rainfall. Sandeep and Ajayamohan (2014) concluded that variations in monsoon low level flow (northward shift of LLJ) combined with the changes in the wind led in debilitating (strengthening) of vortices over the southern (northern) Arabian Sea. From the analysis, LLJ appears to be strengthening (weakening) over the Indian Ocean at north (south) of ~15^{0}N during excess (deficit) rainfall years. Also this LLJ influences the cross equatorial flow and is more intense during excess years than deficit years.

Another component that influences the KSMR is the tropospheric temperature (TT), the main driving force that influences LLJ. Parthasarathy *et.al*.

(1990) inferred that TT during the pre-monsoon over India influences the subsequent monsoon rainfall. According to Li & Yanai (1996) and Kawamura (1998), the TT and the Asian monsoon have a close association. The current study, however, suggests that inter-annual variability in the summer monsoon is linked to land-sea thermal contrast, particularly between the Tibetan Plateau and the equatorial Pacific Ocean, with the temperature gradient increasing from south to north.During monsoon season, the 700-200hPa experience highest temperature than pre-monsoon period and it is facilitated by diabetic heating formed as result of sensible heat flow from the surface of the Tibetan Plateau along with the latent heat produced from cloud formation over eastern Plateau (Yanai *et al*., 1992; Keshavamurty and Rao, 1992) that is TT is high over Indo-Tibetan regions than the surrounding oceans. As a result of warming of the Tibetan plateau, warm air rises over the landmass, causing a greater inflow of moist air from oceans towards the continent. Kothawale and Kumar (2002) concluded that during deficit years, the surface temperature are above normal and the upper tropospheric temperature are below normal with pronounced cooling during monsoon season due to decreases latent heat release whereas in excess years, the surface temperature are below normal but the upper tropospheric temperature increases with height so we can conclude that monsoon are sustained due to the warming of troposphere rather than land surface gradient. The TT gradient between the Tibetan high and adjacent ocean was high during excess and was comparatively weak during deficit years. Li and Yanai (1996) pointed out that augment (weak) land-sea contrast in the upper troposphere leads to strong (weak) monsoon that is, during excess years, warming effect was seen over land mass and cooling is observed near the equatorial Indian Ocean. When TT gradient increases, it influences the pressure gradient increases thereby strengthens low level winds and brings copious amount of moisture to the land surface.

Yin (1949) and Koteswaram (1958) suggested that mechanism of monsoon depends on upper air circulation. Monsoon atmosphere exhibits barotrophic and baroclinic instabilities. Baroclinic instability are due to lower level convergence (moisture flux convergence over south of Kerala) and upper level divergence (anticyclone) over the Tibetan region. Barotrophic instabilities are due to the existence of horizontal wind shear which result in splitting of jet stream. With vertical and zonal wind shear, Joseph and Raman (1996) discovered the existence of westerly low level jet streams over peninsular India. LLJ appears to be easterly trade winds from the south Indian Ocean at first glance. LLJ splits into two branches over the Arabian Sea. One branches pass south eastward towards Sri Lanka and the other moves eastwards towards peninsular India. During excess years, the eastward branch (easterlies) shows more strength than deficit years i.e., wind shear is maximum result in more the monsoon activity.

5.3 INFLUENCE OF ENSO AND IOD

We also analyzed the impact of SST patterns in the Pacific and Indian Oceans on Kerala seasonal rainfall. Dry and wet ISMR years are associated to El Nino and La Nina, according to Abhilash *et.al*. (2018), however there is no such association in KSMR, implying that KSMR deficit/ excess years do not necessarily corresponds to ISMR excess/deficit years. A similar pattern was seen during 1948-2019. The majority of the extreme JJAS dry seasons coincide with El Nino or Neutral years. On the other side, La Nina years favour Kerala's intense wet season. However there is no significant relationship exist between El Nino and deficit rainfall. As a result, it is crucial to investigate whether any reason or variables have had a significant impact on monsoon performance during El Nino years.

IOD occurrences are growing over time, while KSMR is decreasing. This rise in IOD events is attributed to weaker equatorial westerly winds and eastward oceanic currents, as well as western EIO warming faster than eastern EIO which facilitate to wind and ocean current reversal (Cai *et al*., 2014). Ashok et al., (2001) identified that weakening of ENSO-ISMR relationship is related to the frequent occurrences of strong positive IOD (pIOD) events that neutralizes the ENSO impact. Severe storms over BOB (Francis *et al*., 2007) may generate pIOD episodes by strengthening the meridional pressure gradient across EEIO, which leads to the amplification of upwelling favoring south-easterlies along the Sumatra coast. pIOD (nIOD) events are connected with LLJ relative strengthening (weakening) over the

EIO and AS (Hrudya *et al.*, 2020). Rainfall increases (decreases) across the Indian Region during pIOD (nIOD), and yet these changes in rainfall are linked to SST shifts, low-level circulation, and moisture transfer over EIO.

5.4 MONTHLY RAINFALL VARIABILITY

When we consider the summer monsoon month, rainfall received during June and July shows a declining trend whereas August and September rainfall shows an increasing trend. Kothawale *et.al* (2013) noted that 50% of June and July deficit rainfall are due to presence or decaying of El Nino events (Boschat and Terray, 2012). After an El Nino occurrence, the majority of MOK is delayed, resulting in below-normal rainfall in June (Joseph *et al.*, 1994), whereas La Nina years have increased rainfall in August and September. During the El Nino season, August and September rainfall performs better than June and July rain, according to Boschat and Terray (2012). Because of the beginning of El Nino, which causes the monsoon wind to diminish and warm SST anomalies in the west EIO, results in a shortfall in monsoon rainfall in June and July. Since the relationship between ENSO and KSMR are diminishing, another factor that influences the change in monthly rainfall pattern are IOD events. IOD holds negative correlation with June and July rainfall while it is positive during August and September. The dipole event is normally triggered in May, with the transition to the dipole phase occurring by the end of June/early July. The IOD index intensifies from July and attain its peak during October – November due to anomalous strengthening of SST anomaly over EEIO that is IOD influence is more realized in the second half of the monsoon period (August and September) and modulates the variability in KSMR towards the second half of the monsoon season**.** Ratna *et al.,* (2020) identified that extreme strong IOD is associated with low level divergence over the EEIO and convergence over India which result in excess rain especially during late in the season. However warm SST anomalies in the central eastern Pacific contribute to low level divergence and supressed rainfall over India in June.

SUMMARY and CONCLUSION

CHAPTER 6

SUMMARY AND CONCLUSION

India, an agriculture oriented county whose economy is balanced by the summer (June through September) monsoon rainfall are currently facing serious challenges due to the existence of year to year monsoon variability. By understanding the variations in rainfall pattern, alternative agriculture practices and methods can be identified to sustain its impact. From 1871-2010, Indian summer monsoon rainfall (ISMR) shows a stable pattern (Kulkarni, 2012) but its relationship with El Nino– Southern Oscillation (Ashok *et al.,* 2019; Kumar *et al.,* 2006), Indian Ocean Dipole, Eurasian and Himalayan snow (Hrudya *et al.,* 2020; Zhang *et al.,* 2019; Kripalani *et al.,* 2003), etc. creates marked change in ISMR which makes the seasonal predictions more challengeable.

Evaluation of rainfall trend is an important element in long-term water resource evaluations and planning efforts. There are several studies that emphasis on ASMR variability based on various combination of parameters as predicators. However the summer monsoon rainfall over Kerala shows low correlation coefficient with ASMR i.e., the existing predictors of ASMR has less influence on KSMR. The present study investigate the monthly and seasonal rainfall variability over Kerala for the period 1948-2019 and brings out some of the interesting and significant changes in the rainfall pattern of the study area. The principles seasons of the state are the SW monsoon from June to August and the NE monsoon comes around October to November. Reduction in monsoon rainfall over Kerala is mainly driven by the weakening of upper and lower level Monsoonal circulations. Any shortfall in rainfall during SWM season is compensated by NEM season so that annual water stress is curtailed during most of the years.

Overall KSMR shows a decreasing trend and the number of dry years (43) are greater than wet (29) years. By using percentage departure criteria, deficit and excess rainfall years were identified. Deficit years includes 1952,1965, 1966, 1972, 1976, 1986, 1987, 2002, 2003, 2012, 2015,2016 and excess rainfall were received during 1959,1961, 1968, 1975, 1991, 1994, 2007, 2013. Maximum and minimum variation in monsoon rainfall is observed during the period 1958-1967 and 1948-1957 respectively. The perspectives of weather parameter such as air temperature, outgoing solar radiation (OLR), vertical velocity, precipitable water, specific humidity and sea level pressure during excess and deficit years were individually analyzed using composite anomaly analysis.

The driving factors for the occurrence of excess and deficit years were identified and studied. Excess (deficit) years are characterized by high (low) Tropospheric Temperature (TT) over the Indo-Tibetan plains, developed from high (low) diabatic heating contributed by sensible and latent heat. Corresponding reduction (enhancement) of TT over Equatorial Indian Ocean contributes strong (weak) TT gradient during excess (deficit) years which trigger strong (weak) easterly wind shear and positive (negative) moisture flux convergence over Kerala region. In the lower levels, this is associated with strong (weak) cross equatorial low level jet. Therefore we can conclude that, TT gradient and easterly shear are the dominant factors that modulates the excess and deficit conditions.

It is also found that high rainfall variability in the recent decade can be linked to rapid Indian Ocean warming. Therefore the relationship between ENSO and IOD with KSMR was considered. It was found ENSO and KSMR does not holds any significant relationship but IOD events influences the rainfall during August and September. From the study it visible that KSMR shows a decreasing trend while IOD events are increasing with time. This decrease in KSMR is associated with weakening of the Tropical Easterly Jet (TEJ), pole-ward shift in monsoon Low level Jet (LLJ), enervate of tropospheric temperature gradient and the increase in IOD events are correlated with expeditious warming of Indian Ocean and frequent occurrence of strong cyclonic events over BOB. On monthly scale, the rainfall received during June and July shows a decreasing trend whereas the rainfall during August and September are increasing. This monthly variations in monsoon rainfall is linked with ENSO and IOD events.

Hence it may be concluded that, high rainfall variability of SWM over Kerala can be linked to Indian Ocean Warming along with restraining warming of Indian sub-continent. The possible reason for excess and deficit rainfall are linked to TT gradient and easterly shear.

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SEASONAL AND MONTHLY RAINFALL VARIABILITY OVER KERALA IN A WARMING CLIMATE

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THESIS ABSTRACT

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ABSTRACT

An attempt were made to understand the seasonal and monthly rainfall variability over Kerala for the period 1948-2019. The Kerala summer monsoon rainfall (KSMR) shows decreasing trend with time. When considering individual monsoon months, June and July shows decreasing rainfall trend while August and September exhibits an increasing trend. Excess and deficit rainfall years were identified using percentage rainfall departure criteria. Meteorological parameters such as air temperature, wind, perceptible water, surface pressure, humidity, sea-level pressure, vertical velocity, and precipitation were analyzed using composite analysis. The influence of dominant remote tele-connection patterns which modulates ISMR like El- Nino and Indian Ocean Dipole (IOD) were examined. Though phases of ENSO largely affects ISMR variability, both warm and cold phases of Nino3.4 SST variability has little influence on KSMR meanwhile IOD shows an increasing trend in recent decades, its influence on seasonal KSMR is strengthening. The influence of IOD is more realized in the second half of the monsoon period (August and September) and modulates the variability in KSMR towards the second half of the monsoon season. Physical mechanism behind the excess and deficit rainfall is also identified. Excess (deficit) years are characterized by high (low) Tropospheric Temperature (TT) over the Indo-Tibetan plains due to high (low) diabatic heating contribution from sensible and latent heat. Corresponding reduction (enhancement) of TT over Equatorial Indian Ocean contributes strong (weak) TT gradient during excess (deficit) years which trigger strong (weak) easterly wind shear and positive (negative) moisture flux convergence over Kerala region. In the lower levels, this is associated with strong (weak) cross equatorial low level jet during excess (deficit) years. Hence it may be concluded that, TT gradient and easterly shear are the dominant factors modulates the excess and deficit KSMR and high rainfall variability in the recent decade can be linked to rapid Indian Ocean warming.