Tropical Sea Surface Temperature Anomalies in Relation to Indian Summer Monsoon, El Niño and Indian Ocean Dipole

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By

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April, 2017

<u>Declaration of Authorship</u>

I, Preenu P N, declare that this thesis entitled, ***Tropical Sea Surface Temperature Anomalies in relation to Indian Summer Monsoon, El Niño and Indian Ocean Dipole*** and the work presented in this thesis is a bonafide record of research work carried out by me under the supervision of **Dr. P.K. Dinesh Kumar**, in the CSIR-National Institute Of Oceanography Regional Centre, Kochi, in partial fulfillment of the requirements for the award of the Ph.D. degree under the Faculty of Marine Sciences, Cochin University of Science and Technology. I further confirm that the subject matter of the thesis has not formed the basis for the award of Degree or Diploma of any University or Institution.

Preenu P N

April, 2017

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Certificate

This is to certify that this thesis entitled, "***Tropical Sea Surface Temperature Anomalies in relation to Indian Summer Monsoon, El Niño and Indian Ocean Dipole"*** is an authentic record of the research work carried out by **Mrs. Preenu P N** under my guidance in the CSIR-National Institute Of Oceanography Regional Centre, Kochi in partial fulfillment of the requirements for the Ph.D. degree of Cochin University of Science and Technology under the Faculty of Marine Sciences, and no part thereof has been presented for the award of any degree in any university.

I also certify that all the relevant corrections and modifications suggested by the audience during the pre-synopsis seminar and recommended by the doctoral committee of the candidate have been incorporated in this thesis.

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"Two roads diverged in a wood, and

I took the one less traveled by,

That has made all the difference."

.....Robert Frost 1920.

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Preenu P N

Preface

Oceans have a greater control on the weather and climate of the Earth. The immense amount of heat energy stored by the oceans act as a giant flywheel to the climate system. The interaction between oceans and atmosphere through the air-sea interface is important in this connection. The Tropic of Cancer (23.5°N) and the Tropic of Capricorn (23.5°S) may be taken as the northern and southern boundaries, respectively of the tropical oceans. This region receives maximum solar radiation throughout the year which in turn heats up the oceans. The Sea Surface Temperature (SST) thus has an important role in relation to weather and climate.

The Indian Summer Monsoon, El Niño-La Niña commonly known as El Niño Southern Oscillation (ENSO) and Indian Ocean Dipole (IOD) are the major climatic phenomena of the global tropics. These phenomena have their footprints both in the atmosphere and oceans. Monsoon is found to interact with the tropical oceans, particularly the Indian and Pacific oceans at inter-annual time scales. Recent research has shown that the main contributors to the inter-annual variability of Indian summer monsoon rainfall (ISMR) are the large scale forcing from ENSO and IOD.

During an El Niño event, the area of maximum SST in the Pacific Ocean shifts eastward, bringing more precipitation over the central and eastern Pacific Ocean. During these periods the eastern Indian Ocean, Indonesia, and north Australia are in the subsiding part of the Walker circulation that has shifted eastward from its climatological position. These changes impact the Indian summer monsoon. IOD also affects the strength of the summer monsoon over the Indian subcontinent. It has been found that in temporal epochs when the correlation between ISMR and ENSO is low, the correlation between ISMR and IOD is high, and vice versa. All these climatic phenomena are associated with anomalies in surface wind and SST. A major concern currently is the phenomenon of global warming. Recent research has shown that ENSO and the related changes in the SST of the global oceans are found to have a role in modulating global warming.

In this thesis some aspects of the large scale ocean atmospheric interaction in relation to climate, particularly that of India have been studied and presented.

The major objectives of the study are:

I. To investigate the role of El Niño/La Niña and ISMR in Global Warming, through their role in changing the SST of the global oceans during 1870-2010 and causing the hiatuses in global warming.

II. To study the inter-annual variability of onset dates of monsoon over Kerala, during 1870-2015 and its relation to SST anomalies and to generate a long time series of the date of Monsoon Onset over Kerala to be used for Climate change studies.

III. To study the genesis of tropical cyclones over west Pacific Ocean and its relation to the date of Pre Monsoon Rainfall Peak (PMRP) and Date of Monsoon Onset over Kerala (DMOK) of the period 1959-2014.

IV. To study the inter-relation among the three phenomena IOD, El Niño/La Niña and ISMR at the inter- annual time scale.

This thesis consists of 7 chapters. Chapter-1 gives a general introduction of the current status of research on ENSO, IOD and ISMR and their relation with one another through ocean- atmosphere interaction.

Chapter-2 gives details of the data sets and methodology used in the study. Hadley center SST data set (HadISST) (1870-2014), Outgoing Long wave Radiation (OLR) data of NOAA (1979-2014) and the dates of monsoon onset over Kerala (DMOK) 1870-2014 from various sources, are the major data sets used for the present study. Tropical cyclone data of the west Pacific Ocean were collected from the Annual Tropical Cyclone Reports published by Joint Typhoon Warning Centre (JTWC), Guam. Details of the statistical techniques such as linear correlation, lag correlation, composite analysis, Hovmöller analysis are also given.

Chapter-3 presents the possible roles of El Niño/La Niña and Indian summer monsoon in the observed global warming. Global warming of the atmosphere and the ocean during 1871 to the present day has not been a continuous increase in the surface temperature of the atmosphere, but the warming has occurred in epochs lasting for a few decades 1910-1940 and 1970-1998. There were epochs with no global warming or even slight global cooling (hiatus) eg 1880-1910, 1940-1970. It was examined among natural causes whether El Niño/La Niña and the Asian Summer Monsoon have roles in global warming and its hiatuses. The global SST time series constructed using an El Niño/La Niña index - SST relation is able to reproduce the gross changes in the rate of global warming observed in the atmospheric surface temperature (global warming) from 1870 to date.

Chapter-4 deals with the variability of the date of Monsoon onset over Kerala (DMOK) and its relation to the SST of the global tropical oceans on inter-annual and longer time scales. Over the last 150 years, the DMOK has varied widely, the earliest being on 11th May, 1918 and the most delayed on 18th June, 1972, with a long term (1870-2014) mean date of 01 June and Standard Deviation (SD) of about 8 days. A long time series of DMOK from 1870 to 2014 has been constructed merging the available data sets of DMOK derived by various methods and periods after making needed adjustments or possible use in research related to climatic change.

In Chapter-5 the genesis of tropical cyclones over west Pacific Ocean and its relation to the DMOK and the Pre Monsoon Rain Peak (PMRP) have been investigated. Tropical west Pacific Ocean generates 26 (out of the 79) tropical cyclones produced by global oceans in a year. About 60 percent of these occur during the four months (June - September) when India gets its summer monsoon rainfall. It is known that both these areas have a prominent 30-50 day signal in many parameters of atmosphere and the ocean. Pre-Monsoon Rainfall peak (PMRP) and DMOK are manifestations of this mode in rainfall associated with the Equatorial Trough (ET) over north Indian Ocean. It has been shown that there is intra-seasonal variation of tropical cyclone genesis in the west Pacific Ocean closely related to the timings of DMOK and PMRP.

Chapter-6 is devoted to the study of the inter-linkages at the inter-annual time scale that exist among ISMR (monsoon rainfall of India - June to September), El Niño/La Niña index (Niño3.4) of October to December season representing El Niño/La Niña in their mature phase and DMI (Dipole Mode Index) of September to November season representing the mature phase of IOD, using a linear correlation method. The main growth of the SST anomaly in the

Niño 3.4 region associated with El Niño/La Niña was during the monsoon season, June to September. This growth in Niño3.4 is having a high and statistically significant negative correlation with ISMR. It is also found that Niño3.4 is positively correlated with DMI of the same year and the DMI in turn is negatively correlated with the Niño3.4 of the following year. But while the linear correlation between Niño3.4 and DMI of the same year is very high and statistically significant during the whole period of study, the liner correlation between DMI and Niño3.4 of the following year is statistically significant only in some temporal epochs of ISMR time series.

Chapter-7 presents a summary of the main findings of this thesis. Future scope of studies is also presented. This is followed by a list of references cited in the thesis in alphabetical order of the names of the first authors.

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6. El Niño, IOD and ISMR inter-relationships

7. Summary and Conclusion

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Abbreviations

ATCR	Annual Tropical Cyclone Reports
BoB	Bay of Bengal
DMOK	Date of Monsoon Onset over Kerala
DMR	Deficient Monsoon Rainfall
DMRE	Deficient Monsoon Rainfall with El Niño
ECMWF	European Centre for Medium-Range Weather Forecasts
EIO	Eastern Indian Ocean
EMR	Excess Monsoon Rainfall
ENSO	El Niño Southern Oscillation
EOF	Empirical Orthogonal Functions
GTS	Global Tele-communication system
IITM	Indian Institute of Tropical Meteorology
IOWP	Indian Ocean Warm Pool
IMD	Indian Meteorological Department
IPCC	Inter-governmental Panel on Climate Change
ISMR	Indian Summer Monsoon Rainfall
ITCZ	Inter Tropical Convergence Zone
JJAS	June July August September
LCC	Linear Correlation Coefficient
MDB	Marine Data Bank
MOK	Monsoon Onset over Kerala
NASA	National Space Administration
NCAR	National Centers for Atmospheric Research
NCEP	National Centers for Environmental Prediction
N3.4	Niño3.4Index

NWPC	North West Pacific Cyclones
OISST	Optimum Interpolation Sea Surface Temperature
OLR	Outgoing Long-wave Radiation
PDO	Pacific Decadal Oscillation
PMRP	Pre- Monsoon Rainfall Peak
OND	October November December
SLP	Sea Level Pressure
SON	September October November
SST	Sea Surface Temperature
SSTA	Sea Surface Temperature Anomaly
TOGA	Tropical Ocean Global Atmospheric
TBO	Tropospheric Biennial Oscillation
WWV	Warm Water Volume

To

My parents

Who didn't give me any scope to feel sorry for myself;

My husband

Who encourages me to follow my dreams;

And my sweet kid

Who teaches me what life is all about;

This thesis is dedicated…

Chapter-1

Introduction

1.1 Tropical Oceans

The tropical oceans [taken as the area between the tropic of Cancer (23.5°N) and the tropic of Capricon (23.5°S)] exhibit large-scale sea surface temperature (SST) variability on both seasonal and inter-annual timescales (Deser et al., 2010). These regions are characterized by low level trade winds, north easterlies to the north and south easterlies to the south of the equator. These trade winds converge along the Inter Tropical Convergence zone (ITCZ) which on the climatic time scale is very close to the axis of maximum SST in monthly means.

Figure 1.1 gives the isolines of mean SST of the period from1950 to 2010 for the months January, May and August. The axis of maximum SST coincides closely with the ITCZ in each of these months. The area enclosed by the isotherm of 28^0C is taken as the Warm Pool of the tropical oceans. From the figure it may be inferred that the centre of the Warm Pool is located over the west Pacific Ocean south of the equator in January. Instead of moving north with the northward motion of the sun in the annual cycle, the centre of the Warm Pool moved north-westward across the equator and in May is located over north Indian ocean here it becomes the warmest area of the global oceans (Joseph, 1990a).

This warmest area of the global oceans produces large scale moisture convergence, convective heating of the atmosphere, generation of the cross-equatorial Low Level Jet-stream and the onset of the Asian summer monsoon (Pearce and Mohanty, 1984; Joseph, 1990, b; Joseph et al., 2006).

After the onset of monsoon, SST of north Indian Ocean cools and in August the centre of the Warm Pool is located over the west Pacific Ocean north of the equator from where it moves southwards across the equator to its January location.

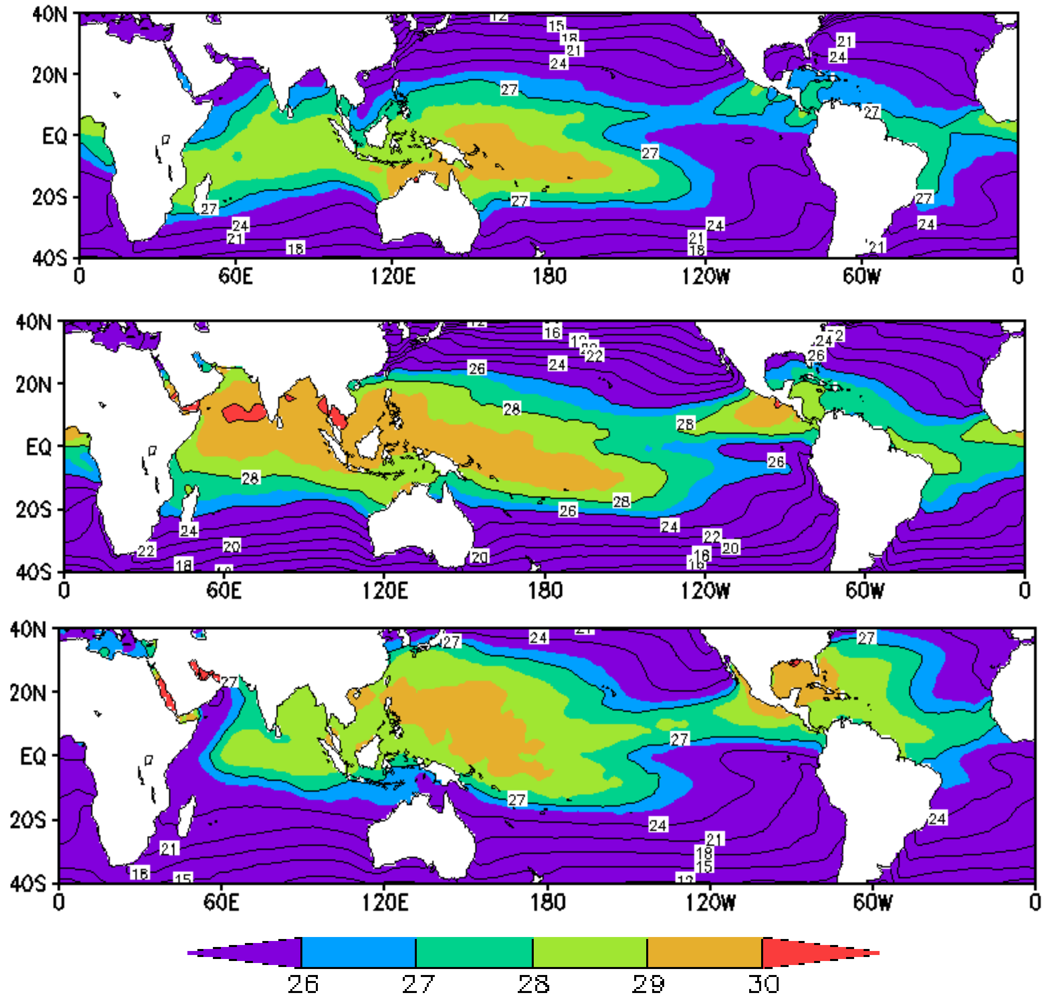


Figure 1.1: Mean SST of the period 1950 to 2010 for the months January (above), May (middle) and August (bottom). Yellow colour area marks the boundary of the Warm Pool with SST of 28^{0}C and more, brown colour 29^{0}C and red colour 30^{0}C

The east-west oriented ITCZ is characterized by maximum precipitation that migrate northwards in spring and southwards in autumn across the equator in the monsoon half of the earth. In this part of the earth, between the boreal winter and summer seasons, the ITCZ undergoes the largest migration from latitudes 10^{0}S to 20^{0}N and back to 10^{0}S, while its migration over other parts of tropical oceans is only through a few degrees of latitude.

1.2 Salient features of tropical oceans

The major climatic phenomena occurring in the tropics are the Indian Summer Monsoon (ISM), El Niño Southern Oscillation (ENSO) and Indian Ocean Dipole (IOD). They have their footprints both in the atmosphere and oceans. Monsoon is found to interact with the tropical oceans, particularly with the Indian Ocean and the Pacific Ocean at inter-annual time scales (Rasmusson and Carpenter, 1983; Webster et al., 1998). Recent research has shown that the main contributors to the inter-annual variability of ISMR are the large scale forcing from ENSO (Rasmusson and Carpenter, 1983; Shukla and Paolino, 1983; Webster and Yang, 1992; Ju and Slingo, 1995; Soman and Slingo, 1997; Webster et al., 1998; Navarra et al., 1999; Lau and Nath, 2004; Kinter et al., 2002; Miyakoda et al., 2003). These studies show that ISMR and ENSO are negatively correlated; this correlation has decreased during recent years (Kumar et al., 1999). The region of maximum positive SSTA in the equatorial Pacific Ocean is located over the equatorial central and east Pacific Ocean during El Niño events, bringing more precipitation over these areas. During these periods the eastern Indian Ocean, Indonesia, and north Australia come under the subsiding part of the Walker circulation. Figure 1.2 gives the SSTA of the October to December season of the El Niño year 1997 and the La Niña year 1998. The mean SST of October to December of the base period 1950-2010 which was used to derive the SSTA is also given at bottom panel of Figure 1.2.

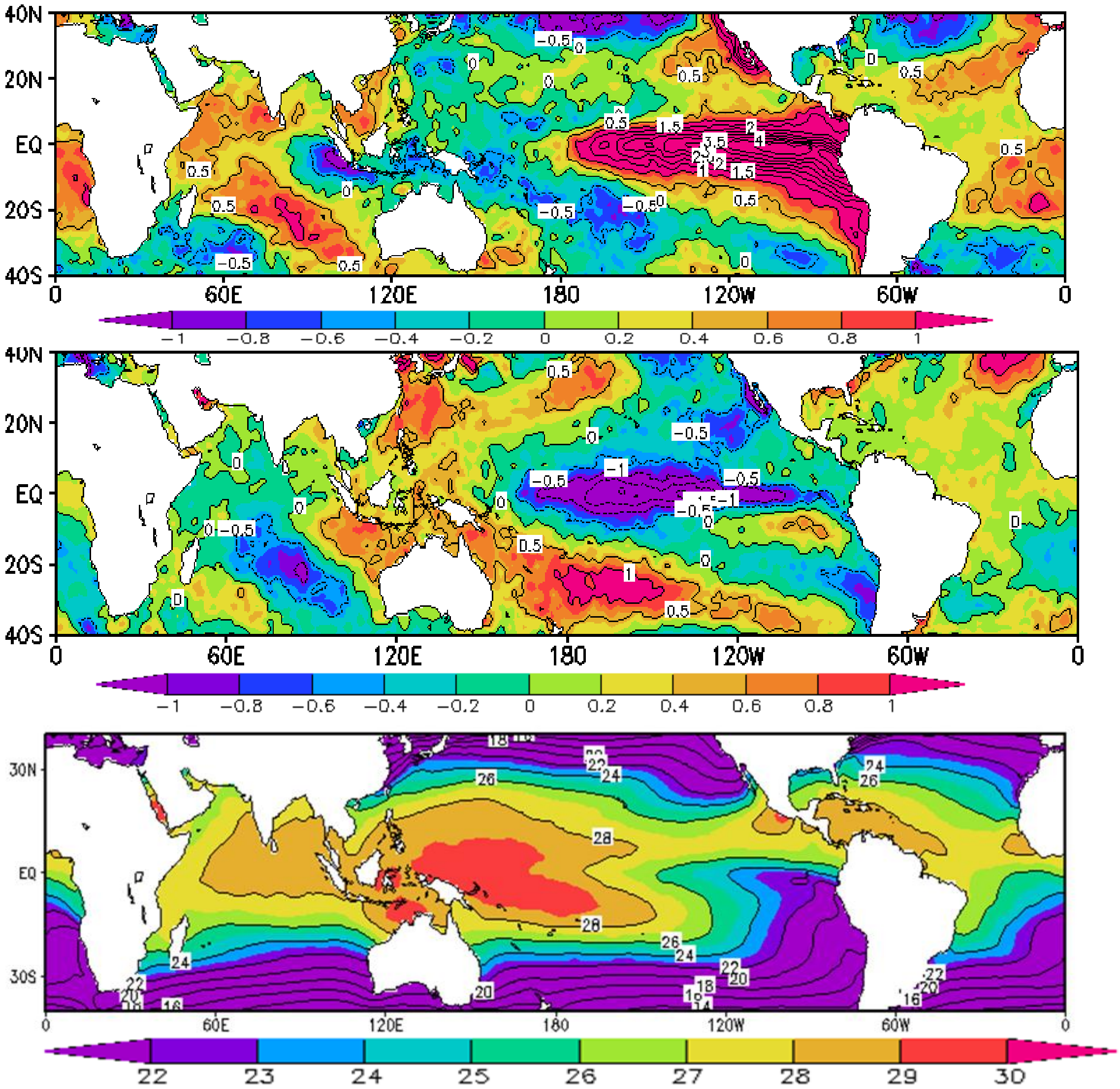


Figure 1.2: SSTA of October to December season of the El Niño year 1997 (above)and the La Niña year 1998 (middle) and mean SST of October to December of 1950 to 2010 (bottom)

The SST cooling in the eastern Indian Ocean is an Indicator of IOD (Saji et al., 1999; Webster et al., 1999) which affects the ISMR. Figure 1.3 is a schematic of the SSTA and deep convection (rainfall) anomalies associated with positive and negative IODs adapted from Saji et al., (1999). It is found that IOD and ENSO have complementarily affected the ISMR during the last four decades (Ashok et al., 2001). Whenever the ENSO-ISMR correlation is low (high), the IOD-ISMR correlation is high (low). A major concern now is the warming trend of SST in global warming and its impact on the global climate. ENSO and the related changes in the SST are found to have significant roles in the global warming (Compo and Sardeshmukh, 2009, 2010). An attempt has been made in this thesis to study some of the major aspects of ocean atmospheric interactions in relation to climate using SST data of the period from 1870 to 2014, with a focus on the changes in the Indian summer monsoon rainfall.

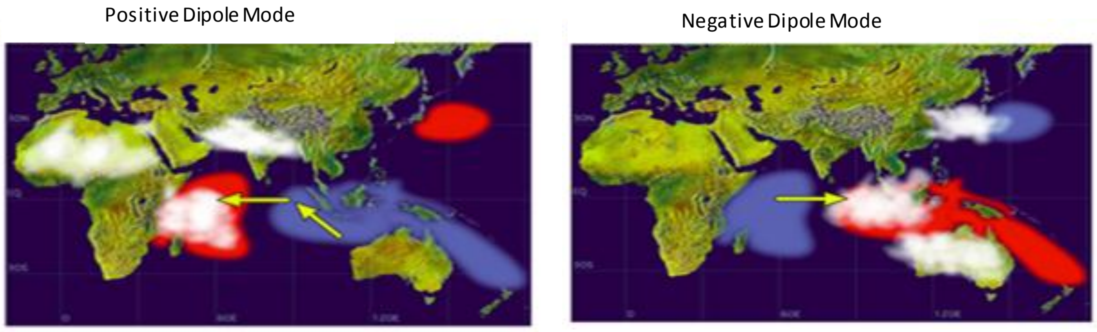


Figure 1.3: Schematic of the SSTA and deep convection (rainfall) anomalies associated with positive and negative IODs adapted from Saji et al., (1999). Positive SST anomaly is marked in red and negative in blue. Areas of associated rainfall are marked in white

1.3 SSTA in Tropical Oceans

SSTA (^{0}C) are defined as the departure of SST from its long term mean. Approximately, 70% of Earth's total area is covered by oceans and changes in SST have an impact on the global climatic variations. The tropical oceans because of their high heat content and capacity for large inter-annual and decadal variations (Gill and Rasmusson, 1983; Fedorov and Philander, 2000), have an important part to play in regulating the global climate (Newman et al., 2003).

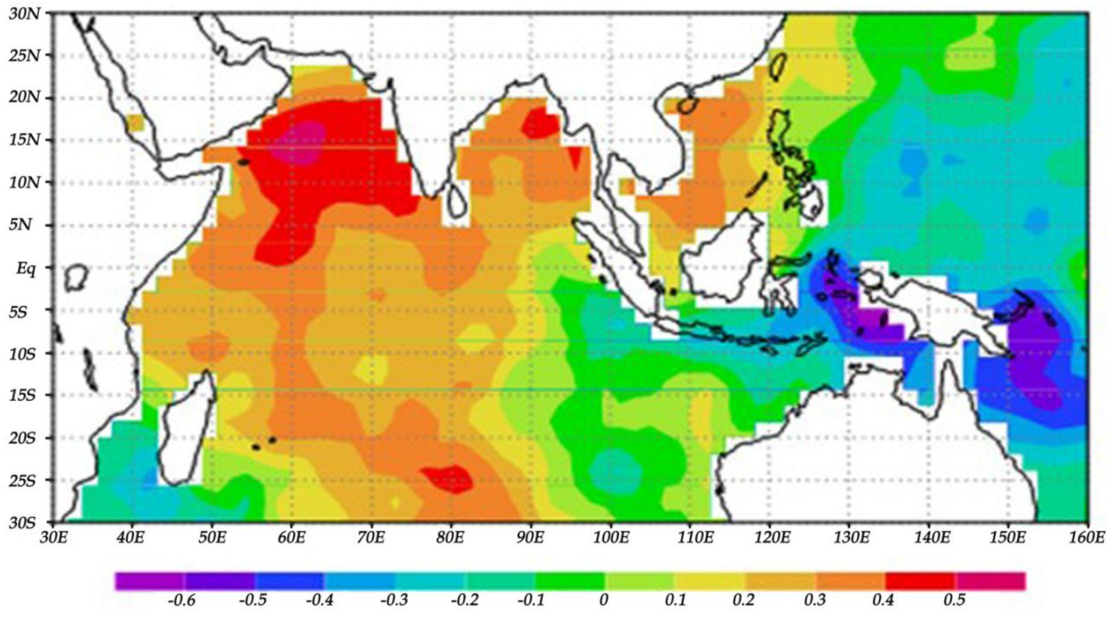


Figure 1.4: SSTA averaged for September to November periods following the five very severe Indian monsoon drought monsoons of 1965, 1972, 1979, 1982 and 1987 of the period 1961 to 1990

Major climatic events in the tropical oceans are found associated with large scale SSTA. For example, El Niño is associated with large warm SSTA over the eastern equatorial Pacific Ocean and also western Indian Ocean. Soon after a season of monsoon drought, the SST of the ocean around India (west Pacific Ocean) warms (cools) and this persists till the following monsoon which has normal or excess ISMR. This phenomenon is called the tropospheric biennial oscillation (TBO) in which the summer monsoons of India and Australia and the SST of the tropical Indian and west Pacific oceans take part. Meehl (1997) has studied the observational aspects of TBO, and Chang and Li (2000) have presented a modeling study. El Niño and monsoon droughts occurring in the same year are associated with larger SSTA (Babu and Joseph 2002). La Niña is associated with SSTA of opposite sign compared to El Niño but with smaller amplitude. The SSTA averaged for the September to November periods following the five very severe Indian monsoon drought monsoons of 1965, 1972, 1979, 1982 and 1987 of the period 1961 to 1990 are shown in Figure 1.4. All these years except 1979 are El Niño years also. IOD, a climatic mode occurring inter-annually in the tropical Indian Ocean is also associated with large SSTA.

1.4 Major climatic events in the Tropical oceans

Tropical oceans receiving maximum insolation is a place where interactions between the ocean surface and the atmosphere are large and associated with Monsoon, El Niño and IOD.

1.4.1 Monsoon

Indian Summer Monsoon is one of the major components of the Asian Summer Monsoon, which brings heavy rains over most of the regions of the Asian Continent. The intensity of rainfall has such a large impact on the resources that the entire economy of India is said to be a gamble on the monsoon rains. The word "monsoon" is derived from the Arabic word mausam, which means seasons. The distinguishing feature of the monsoon is its six monthly wind reversal. The mechanism of monsoon is controlled by several factors such as temperature gradient between land and oceans, moisture from Indian Ocean, the Earth's rotation and solar irradiance (Webster, 1987a) and heavy precipitation occurs over the Indian sub-continent from June to September (Parthasarathy et al., 1992). Monsoon has significant tele-connections with global weather and climate.

The monsoonal region has been defined on the basis of significant change in the wind direction between winter and summer months extending over a large part of the tropics from 25^oS to 35^oN and 30^oW to 170^oE and is give in Figure 1.5. Ramage (1971) has proposed four criteria to determine whether a region can be taken as part of the monsoon area:

(i) A shift in the wind direction at least by 120^0 between January and July.
(ii) An average frequency of the respective prevailing wind direction in January and July exceeds 40%.
(iii) The mean resultant wind speed in at least one of the months exceeds 3 ms^{-1}.
(iv) Fewer than one cyclone-anticyclone alternation occurs every two years in either month in a 5^0 latitude-longitude grid.

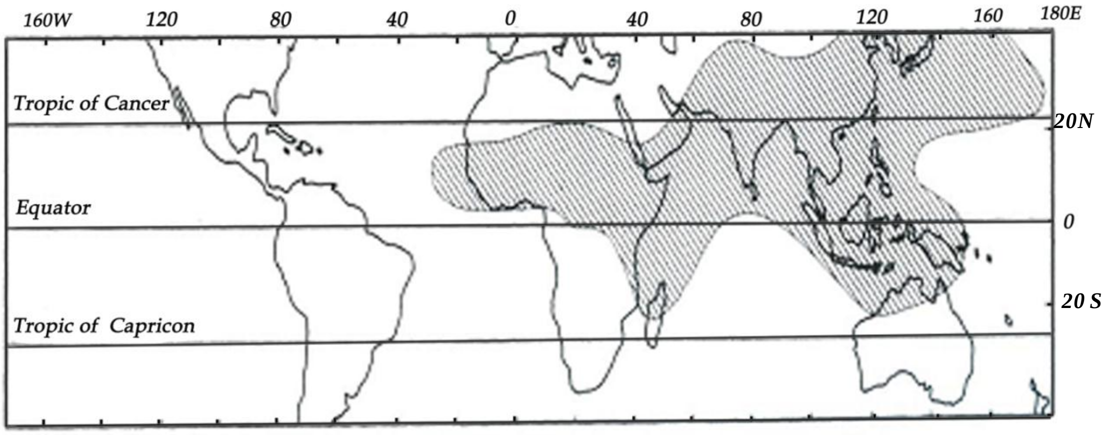


Figure 1.5: Monsoon areas (shaded portions) following the criteria of Ramage (1971)

The summer monsoon onset over Asia has three phases, (a) onset over the southeastern Bay of Bengal (BOB) followed by onset over the South China Sea (SCS) and finally over the southwest coast of India (Krishnamurti, 1981; Joseph et al., 1994, 2006; Wu and Zhang, 1998; Wang and Lin, 2002; Mao et al., 2003; Wang et al., 2004; Mao and Wu, 2007; Yang et al., 2012), (b) peak rainy months in India of July and August (Ananthakrishnan et al., 1967; Mooley and Parthasarthy, 1984; Ding and Sikka 2005; Goswami et al., 2006) and (c) withdrawal in September of monsoon at the end of rainy season (Syroka and Toumi, 2001, 2002, 2004; Saberali et al., ,2012).

The Indian sub-continent, lying in the center of the monsoonal region and surrounded by oceans on three sides is largely influenced by SST of the adjoining seas (Rajeevan et al., 2008). Observational studies (Shukla and Misra, 1977; Shukla, 1987; Joseph and Pillai, 1984; Rao and Goswami, 1988), as well as modeling studies (Shukla, 1975; Washington et al., 1977) have pointed out a significant correlation between SST and ISMR. Model studies (Yamazaki, 1988; Chandrasekar and Kitoh, 1998; Meehl and Arblaster, 2002) have shown that SST plays a significant role in ISMR. There are observational studies which show that positive SSTA in the Arabian Sea during the pre-monsoon season are precursors for an above-normal monsoon rainfall over India (Shukla, 1975; Joseph and Pillai, 1984; Rao and Goswami, 1988; Yang and Lau, 1998; Clark et al., 2000).

1.4.1.1 Monsoon Onset Over Kerala (MOK)

Climatologically, monsoon sets in over the extreme south of India (Kerala) by the end of May or early June, which heralds the main rainy season over the Indian subcontinent. The onset of westerly summer monsoon wind flow is in response to the seasonal transition of land-sea thermal gradient (Holton, 2004; Webster, 2006). It marks the change over from the winter to the summer circulation. Onset of monsoon rains (called the monsoon onset in this study) is preceded by large scale changes in the atmosphere and oceans in the Indo-Pacific region (Kraus, 1955; Ananthakrishnan et al., 1983; Pearce and Mohanty, 1984; Ananthakrishnan and Soman, 1988; Soman and Krishna Kumar, 1993; Joseph et al., 1994, 2006). Accompanying the monsoon onset, there is a rapid increase in the daily rainfall over Kerala which is evident from the Figure 1.6, increased moisture in the atmosphere and strong low level westerly winds (Krishnamurti, 1985).

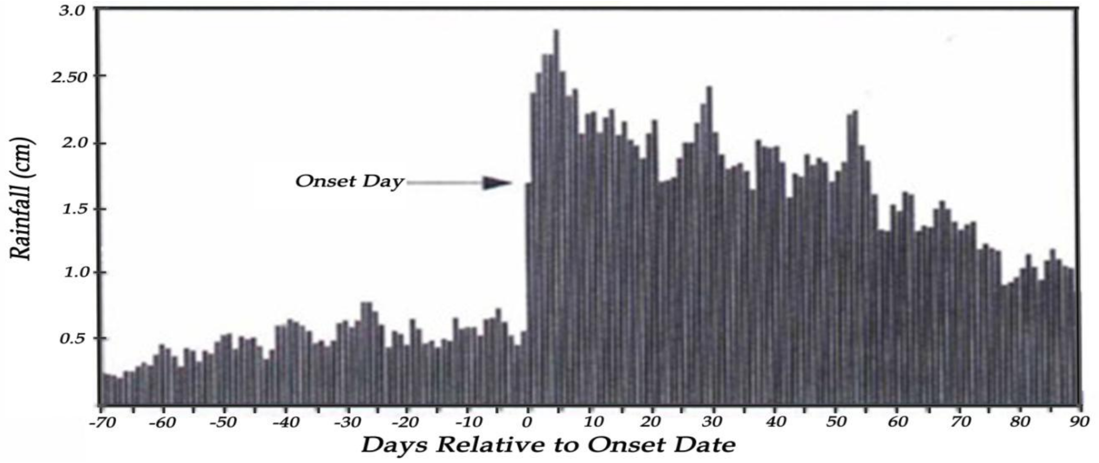


Figure 1.6: The daily mean rainfall of south Kerala as an average of 80 years (1901-1980). Monsoon onset is taken as day-0. Rainfall shows a sudden increase at monsoon onset (from Ananthakrishnan and Soman, 1988)

The dates of onset of the summer monsoon rains over different parts of Asia and the western Pacific Ocean are illustrated by Tao and Chen (1987) in Figure 1.7. Dates of monsoon onset are based on the long-term average pentad (five day non-overlapping) rainfall at several observatory stations of the area. The middle date of the pentad, which shows an abrupt increase in rainfall, is taken as the monsoon onset date for each station. The long term mean date of monsoon onset over Kerala (MOK) is 01 June, with a standard deviation of 8 days. By July 15, the entire country comes under the spell of monsoon rains.

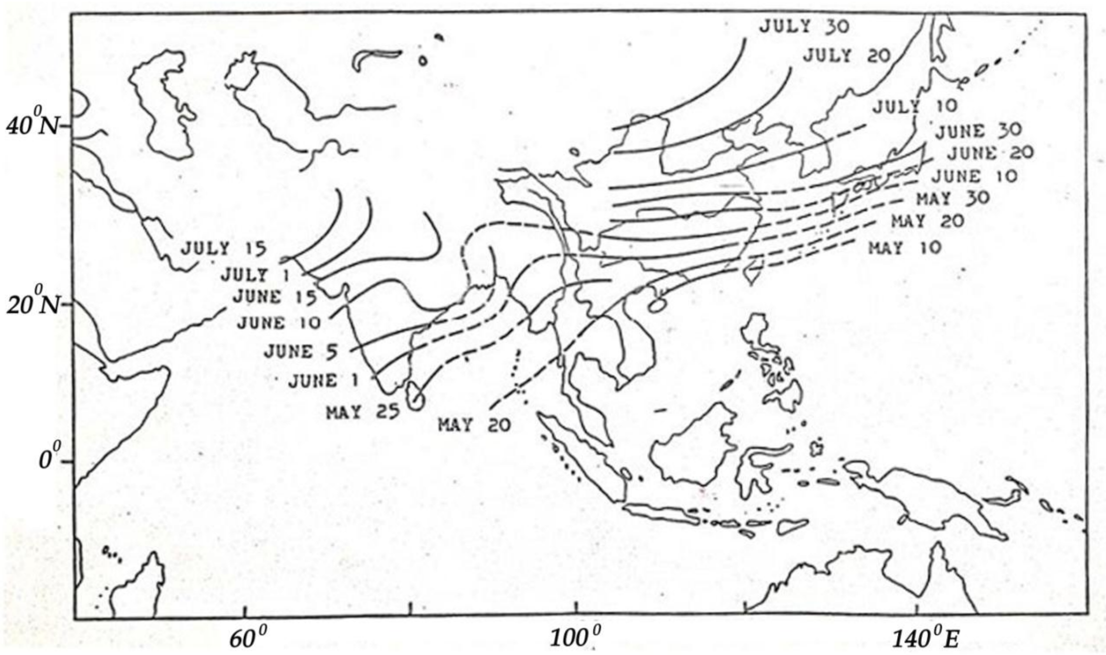


Figure 1.7: Isochrones of mean onset dates of Asian summer monsoon determined from long term rainfall records (Tao and Chen, 1987). The dates of monsoon onset over India are derived by Indian Meteorological Department

There have been many studies of the different aspects of MOK. (i) Monsoon onset and associated changes in ocean and atmosphere were investigated by: Yin, 1949; Koteswaram, 1958; Ananthakrishnan et al., 1968, 1983; Chaudhury and Karmakar, 1981; Krishnamurti et al., 1981; Krishnamurti and Ramanathan, 1982; Mohanthy et al., 1982; Murakami and Nakazawa,1985; He et al., 1987; Lau et al., 1988; Yanai and Li, 1992; Li and Yanai, 1996; Wu and Zhang, 1998; Hsu et al., 1999; Annamalai et al., 2005; Reynolds et al., 2007; (ii) the association of MOK with 30-50 day oscillation: Sikka and Gadgil, 1980; Yansunari, 1980,1981; Krishnamurthi and Subramanyan, 1982; Murakami et al., 1984; Krishnamurthi et al., 1984; Joseph, 1990a,b; Joseph and Pillai, 1988; (iii) forecast of monsoon onset and statistics: IMD 1943; Ramdas et al., 1954; Kung and Sharif, 1982; Ananthakrishan and Soman, 1988, 1989; Mooley and Shukla, 1987; Joseph and Pillai, 1988; Joseph et al., 1994, 2006; Pai and Rajeevan, 2007, 2009.

Joseph et al., 1994 studied the inter-annual variability of monsoon onset and have given a detailed review on studies about MOK. They have observed that prior to and during MOK, dramatic changes occur in the atmosphere and oceans. One of the major parameters associated with the variability of the Date of Monsoon Onset over Kerala (DMOK) is the SST over the Indian and Pacific oceans (Pal et al., 2016). The center of the Warm Pool of the tropical oceans gets shifted from south-west Pacific to the north Indian Ocean during January to May (Joseph, 1990a, b, 2014). The Warm Pool developed over the north Indian Ocean during May increases the moisture content in the atmosphere over north Indian Ocean compared to its adjoining Pacific Ocean, which set the conditions favorable for MOK (Pearce and Mohanty, 1984; Joseph et al., 2006). A general circulation model study has shown that a combination of increased SST and dry intrusion over the equatorial side of the continent play a crucial role in enhancing a convective instability prior to the onset (Kawamura et al., 2002). Based on above study, Minoura et al., (2003) showed that increased land-ocean thermal contrast has influence on the low-level circulations over its adjacent oceans to develop SSTA and surface ocean currents prior to monsoon onset over India.

1.4.1.2 Transition of ITCZ across equator and its relation to DMOK

The area near the equator of low pressure and wind convergence is called the ITCZ. Water vapor condenses as air ascends in the ITCZ forming clouds and rain. From space, the ITCZ can be seen as a narrow band of clouds around the planet in its equatorial region. In the monsoon half of earth during the Southern hemispheric summer (December to February), ITCZ lies south of the equator. Winds from the Northern hemisphere blow across the equator towards the ITCZ. During Northern hemispheric summer (June to September), the ITCZ lies north of the equator and winds blow from the Southern hemisphere across the equator to reach the ITCZ (the summer monsoon). This can be seen in the Figure 1.8 given below.

Figure 1.8: Climatology of global ITCZ of January (blue curve) and July (red curve) as derived by HippsL.E of Utah state university

The time of transition of the ITCZ from south to north across the equator is found related to DMOK (Joseph et al., 1994), where a delayed crossing the equator of the

ITCZ is associated with a delay in DMOK. The ITCZ in monthly climatology has large north south movement over the Indian Ocean and in the adjoining Pacific Ocean (30^0E- 120^0E). In a large part between 180^0E and 360^0E (over Pacific Ocean), the ITCZ lies only in the northern hemisphere both in January and July at low latitudes with only a small north south migration between summer and winter.

1.4.1.3 Warm Pool SST and Associated Convection In Tropical Oceans

Warm Pool of the ocean (the area with SST greater than 28^0C) is located in northern hemispheric winter season (December-February), over southern west Pacific Ocean which shifts to north Indian Ocean and adjoining west Pacific Ocean during the northern hemispheric summer month of May. West Pacific and Indian Ocean are the only regions where SST > 28 ^{0}C exists throughout the year. Warm Pool has important implications, as they are areas of active convection (Lau and Chan, 1986; Ardanuy et al., 1987), wind convergence (Rasmusson and Carpenter, 1982) and areas favorable for the genesis Tropical cyclones (Gray 1975) and production of high precipitation (Taylor 1973; Weare et al., 1981). In most of the months the Warm Pool lies over the Pacific Ocean (140^0 to 180^0E) and is called western Pacific Warm Pool (Sadler et al., 1987). Lukas and Webster (1989) have explained the importance of Warm Pool. The Warm Pool shifts in position to north Indian Ocean before summer monsoon onset, which is followed by dramatic changes in the atmosphere during May. The Warm Pool in the Indian Ocean is found to undergo large seasonal variation (Vinayachandran and Shetye, 1991). In early March, the Tropical Indian Ocean begins to warm up to >29^0C to initiate the onset of monsoon at the end of May (Joseph 1990 b). Large part of Indian Ocean Warm Pool (IOWP) has an OLR < 240 wm-^{2} indicating that this is a region of deep convection. Soon after the onset of monsoon, the Indian Ocean cools, when the Warm Pool centre locates to the west Pacific Ocean. The IOWP creates large scale moisture convergence, which in turn, produces deep convective clouds, which heats the atmosphere in a deep layer and intensifies the low level winds, triggering the onset of monsoon. During an El Niño year, the Warm Pool and associated convection migrates from the west Pacific towards the east along the equator (Gill and Rasmussion, 1983; Donguy et al., 1984).

1.4.2 El Niño Southern Oscillation (ENSO)

ENSO (El Niño Southern Oscillation) is one of the most important manifestations of coupled atmospheric ocean system in tropical area with an irregular oscillation (McPhaden et al., 2006; Deser et al., 2010) of period 2-7 years. It refers to the coherent variations in SST, rainfall, pressure and winds across the tropical Pacific Ocean. ENSO can directly affect all ecosystems and agriculture, while it can intensify cyclones and other severe weather events (Trenberth et al., 1998). El Niño indicates the ocean signature and Southern Oscillation is its atmospheric component and together the phenomena are called ENSO. It swings between opposite extremes in SST over the eastern Pacific Ocean, El Niño (warm phase) and La Niña (cold phase). The warmest SST in El Niño coincides with Christmas season, and so the event is named as El Niño, which means the "*boy*" child. Climatologically, the western Pacific is characterized by warm SST, low sea level pressure (SLP) and heavy rainfall, whereas the eastern Pacific is characterized by colder SSTs, high SLP and poor rainfall. Bjerknes (1966) proposed a theory on the impact of warm SSTA over the eastern and central equatorial Pacific on the Hadley cell. He tested it using the data for 1957-58. These SSTA usually appear in spring and intensify under the effect of the Bjerknes feedback (Bjerknes 1969), a positive air-sea feedback loop in the tropical Pacific. In this positive feedback loop, a positive SSTA in the central Pacific enhances deep atmospheric convection and westerly wind anomaly (Gill 1980). This wind anomaly drives an anomalous eastward flow in the central Pacific that pushes the Warm Pool further to east to strengthen the initial SSTA. This feedback eventually leads to the development of El Niño, which peeks towards the end of the year. Trenberth (1997) defined that if a 5-month mean in SST anomalies in the N3.4 region (5°S–5°N and 120°W–170°W) is > +0.5 °C for at least six consecutive months, then an El Niño event has occurred. Likewise, if the 5-month mean of SST anomalies in the N3.4 region is < -0.5 °C for at least six consecutive months, then a La Niña event has occurred.

1.4.3 Indian Ocean Dipole (IOD)

IOD is a coupled ocean atmospheric phenomenon in tropical Indian Ocean (Saji et al., 1999; Webster et al., 1999; Rao et al., 2002). IOD has significant impacts on the climate over the Indian Ocean (Behera et al., 1999; Ashok et al., 2001; Guan and Yamagata, 2003; Zubair et al., 2003; Saji and Yamagata 2003; Rao and Behera 2005; Dinesh et al., 2016). During a positive (negative) IOD year, the southeastern Indian Ocean is anomalously cool/warm and the western Indian Ocean is anomalously warm/cool. The intensity of IOD is measured by Dipole Mode Index (DMI), which is estimated from SSTA at east and West Indian Ocean (Saji et al., 1999). Based on composite analysis of major IOD events, Saji et al., (1999) suggest that the IOD is initiated in the season May-June, peaks during September-October, and decays by December.

IOD, though considered as an independent mode of variability in Indian Ocean (Annamali et al., 2003; Lou et al., 2008, 2010), has a tendency to co-occur with ENSO events (Annamali et al., 2003; Murtugudde et al., 2000; Xie et al., 2009). Izumo et al., 2010 suggested that negative (positive) IOD tends to favor an El Niño (La Niña) about 14 months later. They suggested a mechanism in which, a negative IOD with positive SST anomalies in east Indian Ocean can have significant influence on the ascending branch of the Walker circulation and to relax the easterlies. The relaxation of easterlies generates positive SSTA in central Pacific, which may get amplified by Bjerkenes feedback and intensify into an El Niño next year.

In this thesis, an attempt has been made to correlate ISMR, El Niño/La Niña index (Niño3.4) between October to December (mature phase El Niño/La Niña) and DMI (Dipole Mode Index) between September to November (mature phase of IOD).

1.4.4 Tropical Cyclone

A Tropical cyclone is an intense vortex (in the atmosphere) over the tropical regions with a cyclonic circulation which leads to organized rainfall causing large scale destruction to life and property. The cyclonic system has high wind speed (> 17 ms^{-1}). It is called hurricane in the north Atlantic Ocean and in north-east Pacific Ocean, while it is known as Typhoon in the north-west Pacific Ocean.

In India tropical cyclones (with maximum surface winds in circulation greater than 47 knots) are called *severe cyclonic storms*, with wind speed > 64 knots are known as *very severe cyclonic storms* and those with >119 knots are the *super cyclones*. The genesis points of cyclones occurring in a 20 year period (Gray, 1975) globally are marked in Figure 1.9.

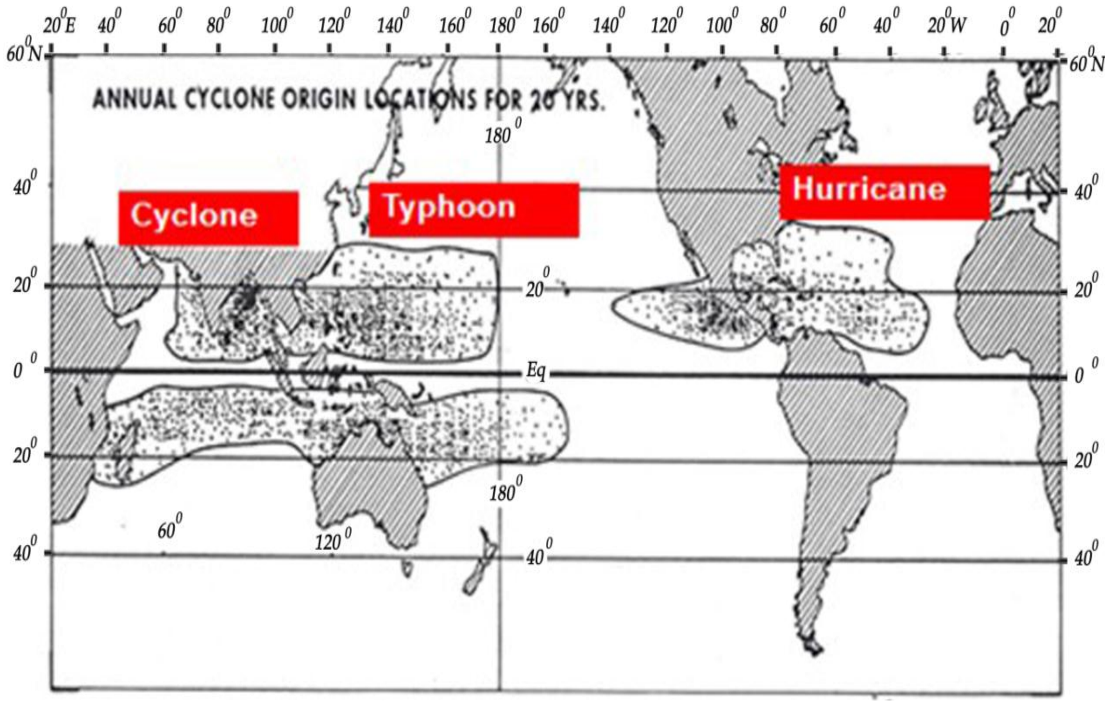


Figure 1.9: Location of genesis points of tropical cyclones over a 20-year period 1958-1977 (Gray, 1975)

Tropical cyclone activity over most of the basins exhibits strong inter annual variations (Landsea, 2000). There are many studies which examined the inter relationship between Indian summer monsoon and cyclones in north west Pacific Ocean (Ramanna, 1969; Saha et al; 1981; Joseph, 1990a; Rajeevan, 1993). There is an inverse relationship between NW Pacific typhoon activity and ISMR on inter-annual time scale (Rajeevan, 1993). Kumar and Krishnan (2005) have shown that the cyclogenesis over NW Pacific was about 1.33 times higher during weak monsoon years compared to strong monsoon years. Pattanaik and Rajeevan (2007) compared the north west Pacific tropical cyclone activity and July rainfall over India to show that north west Pacific cyclone activity induces negative precipitation anomalies during monsoon over India. In the present study, the intra-seasonal variation of tropical cyclone genesis in the west Pacific Ocean with respect to DMOK has been investigated in Chapter-5.

1.4.5 Global warming - Role of Ocean and El Niño/La Niña

Global warming is the observed rapid increase in Earth's average surface temperature in the past two centuries. It is believed that the concentration of greenhouse gases emitted into the atmosphere since the industrial revolution is the main cause of this warming trend (Meehl et al., 2007). Despite a continuous increase in greenhouse gases, the global annual mean temperature has shown hiatus (no increase or even a small decreasing trend) in 21st century (Easterling and Wehner, 2009; Foster and Rahmstorf, 2011). A hiatus in global warming during 1940 - 1970 may be seen in Figure-1.10 taken from the 4th IPCC report (2007).

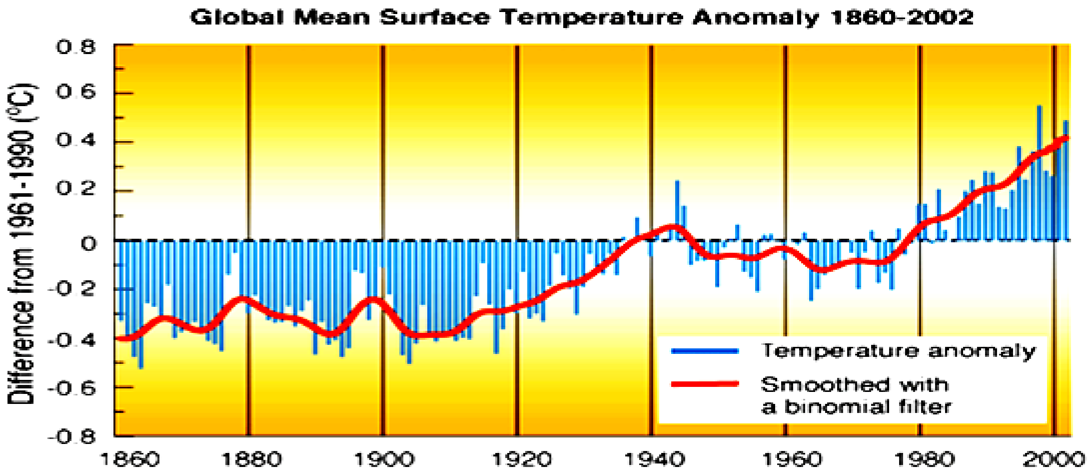


Figure 1.10: Global surface air temperature anomaly 1860-2000. (Fourth IPCC Report, 2007)

Many mechanisms have been proposed for the hiatus in global warming. Some of the main mechanisms suggested for the slow down or reverse trend of global warming are: (i) a minimum in the solar irradiance, related to the number of sunspots (Fröhlich, 2012), (ii) an increase in the stratospheric aerosols in the absence of major volcanic eruptions (Solomon et al., 2011) and the rapid growth in tropospheric short-lived sulfur aerosols (Kaufmann et al., 2011), and (iii) a negative radiative force due to a significant decrease in the stratospheric moisture content (Solomon et al., 2010) and internal cooling (Lean and Rind, 2009; Kosaka and Xie, 2013). Some studies also show significant uptake of heat by oceans, especially the Pacific (Balmaseda et al., 2013; Watanabe et al., 2013), which indicate that the deep ocean is warmed during the hiatus period (Meehl et al., 2011), though this is under debate. While all these studies have addressed only the recent hiatus that occurred since 1998, in the present study the global warming pattern for the past 136 years has been analised. The warming of the atmosphere and the ocean from 1870 onwards has not shown a continuous increase, as it has occurred in epochs lasting a few decades (1910-1940 and 1970-1998). There were also periods of no global warming or even a slight cooling (hiatus) (1880-1910, 1940-1970 and after 1998). In this thesis the possible role of El Niño/La Niña in creating the rapid warming and hiatus phases in Global warming has been examined (Chapter-3).

1.4.6 Scope of the study

The different climatic phenomena generated due to the complex interaction between the tropical oceans and atmosphere can be identified from SST variations. For example, the monsoon on an annual scale and the El Niño with a periodicity of 2-7 years and IOD are associated with large SSTA. The motivation for the present study is to improve the understanding of the relation, between the tropical climate and its changes with SST, and to interpret its role in the global climate change. This study has focused on the tropical area (Indo-Pacific region) where the SST is large and SSTA are prominent. The main objectives of this study are to: (i) investigate the roles of El Niño/La Niña and Monsoon drought in Global Warming and its hiatuses using the SST data for the long period 1870-2014, (ii) to study the inter-annual variability of DMOK during the period 1870-2014 and its relation to SSTA, and to generate a long time series 1870-2015 of the DMOK merging all available data on MOK for use in climate change studies (iii) investigate the genesis of tropical cyclones over west Pacific Ocean and its relation to PMRP and DMOK and (iv) to study the relation among IOD, El Niño/La Niña and ISMR on an inter-annual time scale.

Chapter-2

Data and Methods

Data Sets

The main aim of this chapter is to provide details of different data sets and method of analysis used in this thesis.

2.1 Hadley Center Ice SST (HadISST)

One of the main sources of reanalysis data used in the present study is the Hadley Center Ice Sea Surface Temperature (HadISST) v.1.1 which provides a unique, globally complete analysis of monthly mean gridded (1^0x1^0) latitude/longitude SST from 1870 to 2015. This SST data were taken from the Met Office Marine Data Bank (MDB), which from 1982 onwards also includes data received through the Global Telecommunications System (GTS). In order to enhance data coverage, monthly median SSTs for 1871-1995 from the Comprehensive Ocean-Atmosphere Data Set (COADS) (now ICOADS) were also used where there were no MDB data. HadISST temperatures are reconstructed using a two stage reduced-space optimal interpolation procedure, followed by superposition of quality-improved gridded observations onto the reconstructions to restore local detail (Rayner et al., 2003). The data were provided by the Hadley Centre (Met Office) for scientific use through their web-site http://www.metoffice.gov.uk/hadobs/hadisst/data

2.2 Optimum Interpolation SST (OISST)

National Oceanic and Atmospheric Administration (NOAA) Optimum Interpolation SST (OISST) Version 2 (Reynolds et al., 2007) data (September 1981 to present) were made available by the NOAA Earth System Research Laboratory Physical Science Division (ESRL/PSD) through their Web site at http://www.esrl.noaa. gov/psd/. These are daily SST records (one daily value for each pixel), with spatial resolution of 0.25^0 x 0.25^0, based on the Advanced Very High Resolution Radiometer (AVHRR) infrared satellite measurements.

2.3 Outgoing Long-wave Radiation (OLR)

Outgoing Long-wave Radiation (OLR) data from the National Oceanic and Atmospheric Administration (NOAA) polar-orbiting satellites (Gruber and Krueger, 1984; Liebmann and Smith, 1996) were used as a proxy for convection. The OLR data are available from June 1974 to present, except for most of 1978 when the data are not available due to satellite problems. The data were interpolated in space to remove any missing values. Data from 1979 to 2010 were used in this study. The raw OLR data is mapped onto a 2.5^0 x 2.5^0 global grid and represents the average of twice daily (one daytime and one night time) satellite passes. The data used is the Interpolated OLR data provided by the NOAA-CIRES (Cooperative Institute for Research in Environmental Sciences) Climate Diagnostics Center, Boulder, Colorado, and obtained from their Web site at http://www.cdc.noaa.gov.

2.4 Date of Monsoon Onset over Kerala (DMOK)

The onset dates for north Kerala from 1870 to 1900 were taken from Ananthakrishnan and Soman, (1989) and for the period 1901 to 1980 from Ananthakrishnan and Soman, (1988). These are objectively derived dates, but using daily rainfall data only. The objective dates of MOK from 1971 to 2005 were taken from Pai and Rajeevan, (2009) and for the later years from the annual monsoon reports published by IMD in its journal Mausam. The subjectively derived dates of the Onset of Monsoon over Kerala as derived by IMD are available for the period 1901-2005.

For period 1870-1900 and 1901-1980 the mean difference in onset dates between these two series is 1 day, north Kerala onset date being the earlier one. For declaring monsoon onset, IMD’s subjective criteria stipulates two consecutive days of rainfall over Kerala and the onset is declared on the second day. Thus, there is really no difference in the onset dates of these two series. To get a long series for the DMOK for the period 1870 to 2015, following method was adopted. For the period 1870 to 1900 we have added 1 to the onset dates for north Kerala as derived by Ananthakrishnan and Soman (1989). The subjective onset dates for Kerala as derived by IMD for the period 1901 to 1970 is taken. To this the objective dates of MOK as derived by IMD of the period 1971 to 2015. Merging these three datasets, date of monsoon onset for Kerala for the period 1870-2015 has been constructed.

2.5 Indian Summer Monsoon Rainfall (ISMR)

Parthasarathy et al., (1994) used rainfall data from a network of 306 climatic rain gauge stations well distributed over India to estimate the monsoon rainfall of India of the monsoon season 01 June to 30 September each year of the period beginning 1871. This data base updated each year is available at the website of the Indian Institute of Tropical Meteorology (IITM), Pune (www.tropmet.res.in).

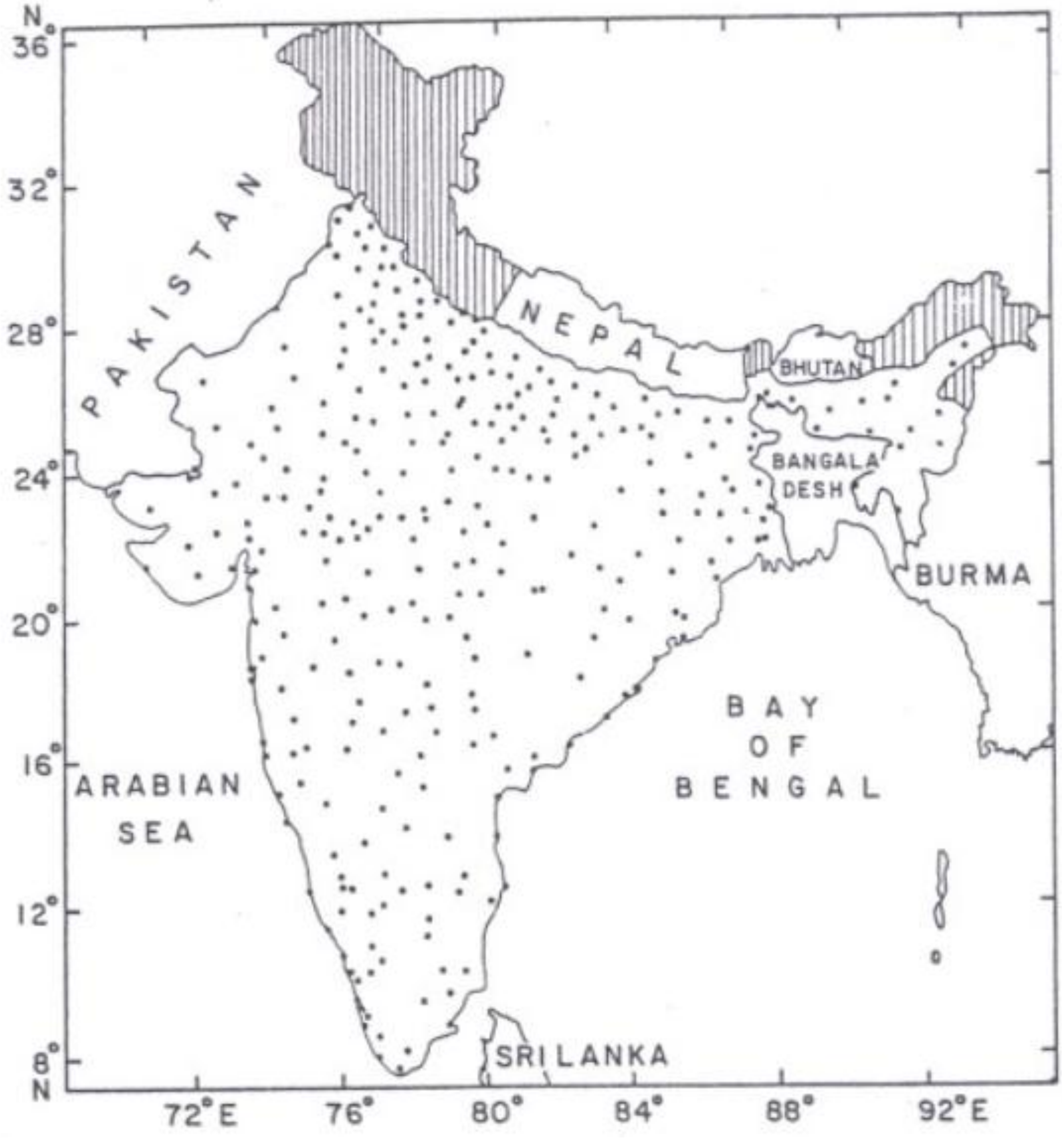


Figure 2.1: Location of the 306 climatic rain-gauge stations used for computing ISMR by IMD, source www.tropmet.res.in. Some hill areas of the Himalayas have not been covered by the network

The long period mean of the ISMR is about 85 cm. Its standard deviation is close to 10% of ISMR. According to the Indian Meteorological Department (IMD), if ISMR is one standard deviation less than the long term mean in a year it is called a drought or dry year, if one standard deviation more a flood or wet year.

2.6 Tropical Cyclones over North West Pacific

The tropical west Pacific Ocean generates 26 (out of the 79) tropical cyclones produced by global oceans in a year (see Figure 1.9). About 60% of these occur during the four months (June - September) when India gets summer monsoon rainfall. In order to study the genesis of tropical cyclones and its association with Indian summer monsoon tropical cyclones over the tropical west Pacific during the period 1959 to 2014 were considered. Data on tropical cyclones in tropical west Pacific Ocean north of the equator were obtained from the Annual Tropical Cyclone Reports (ATCR) issued by the Joint Typhoon Warning Center (JTWC) on Guam. Each year's ATCR contains a summary of the best track position and intensity (at 6-hr intervals) for all cyclones in the area of responsibility of JTWC. The JTWC reports were downloaded from the website https:/metoc.ndbc.noaa.gov/web/guest/jtwc/annual-tropical-cyclone reports

2.7 ENSO index (N3.4)

The monthly ENSO index is based on the SSTA averaged in the N3.4 region (170^0W to 102^0W, 5^0S to 5^0N). Monthly N3.4 index for this study was taken from http://www.cpc.ncep.noaa.gov/data/indices/sstoi.indices. In this thesis strong El Niños were identified using the condition, if 5- month running mean of SSTA in the Niño3.4 region exceeded +1 ^{0}C for 5 consecutive months or more. Similarly, the strong La-Niñas were selected, if 5- month running mean of SSTA in the N3.4 region were -1^0C or less for consecutive 5 months or more.

2.8 Dipole mode index (DMI)

The intensity of IOD was calculated using Dipole mode index (DMI). DMI is the difference between SST of two boxes, one in the western Indian Ocean (50°E to 70°E and 10°S to 10°N) and the other in south eastern Indian Ocean (90°E to 110°E and 10°S to 0°S). The data were collected from JAMSTEC web site http://www.jamstec.go.jp/frcgc/research/d1/iod/HTML/Dipole%20Mode%20Index.html

2.9 Methods of analysis

The analysis and conclusions drawn in this study are based on various techniques and statistical methods to find the variability of the meteorological parameters used in the analysis. Standard deviation was used as a measure to represent the delayed and early monsoon onset years. The Pearson‘s correlation analyses were carried out to find out the relation between two variables and it was tested statistically using Student’s t-test for the desired level of significance. Linear trends of monsoon onset days and its variability were studied during 1870-2014 and the significant differences between different decades were tested using Student’s t-test. Seasonal evolution is an important characteristic of the weather phenomena addressed in this study (Monsoon, El Niño and IOD). Because of the systematic seasonality, composite analyses were used to study different aspects of the above mentioned phenomena. The method of Principle component analysis, also known as Empirical Orthogonal Functions (EOF) analysis was used to study spatial modes of variability of SST. The method gives a measure of variability of SST and how they change with time over Indo-Pacific domain during different seasons.

All data sources and their details are presented in table 2.1 for better summarization.

Table 2.1 Data sets used in the thesis

Name	Source	Coverage
HadISST	http://www.metoffice.gov.uk/hadobs/hadisst/data	Monthly 1^0 X 1^0
OISST	http://www.esrl.noaa. gov/psd	Daily 1^0 X 1^0
OLR	http://www.cdc.noaa.gov.	Monthly/Daily 2.5^0 x 2.5^0
DMOK	various sources (see 2.4)	Over Kerala
ISMR	www.tropmet.res.in	All India
Tropical cyclone	https:/metoc.ndbc.noaa.gov/web/guest/jtwc/annual -tropical-cyclone reports	Over NW Pacific
N3.4	http://www.cpc.ncep.noaa.gov/data/indices/sstoi.in dices	Monthly
DMI	http://www.jamstec.go.jp/frcgc/research/d1/iod/H TML/Dipole%20Mode%20Index.html	Monthly

Chapter-3

Role of El Niño/La Niña and Monsoon in Global Warming and its Hiatuses

3.1 Global warming and its Hiatuses

The increase in surface air temperature in the atmosphere and increase of ocean SST were not continuous, as there were epochs lasting a few decades each (1880 to 1910, 1940 to 1970 and from 1998 onwards) when there was no warming at all or there was even a slight cooling (hiatus) during these periods (Figure 3.1 taken from 5th IPCC Report, 2013).

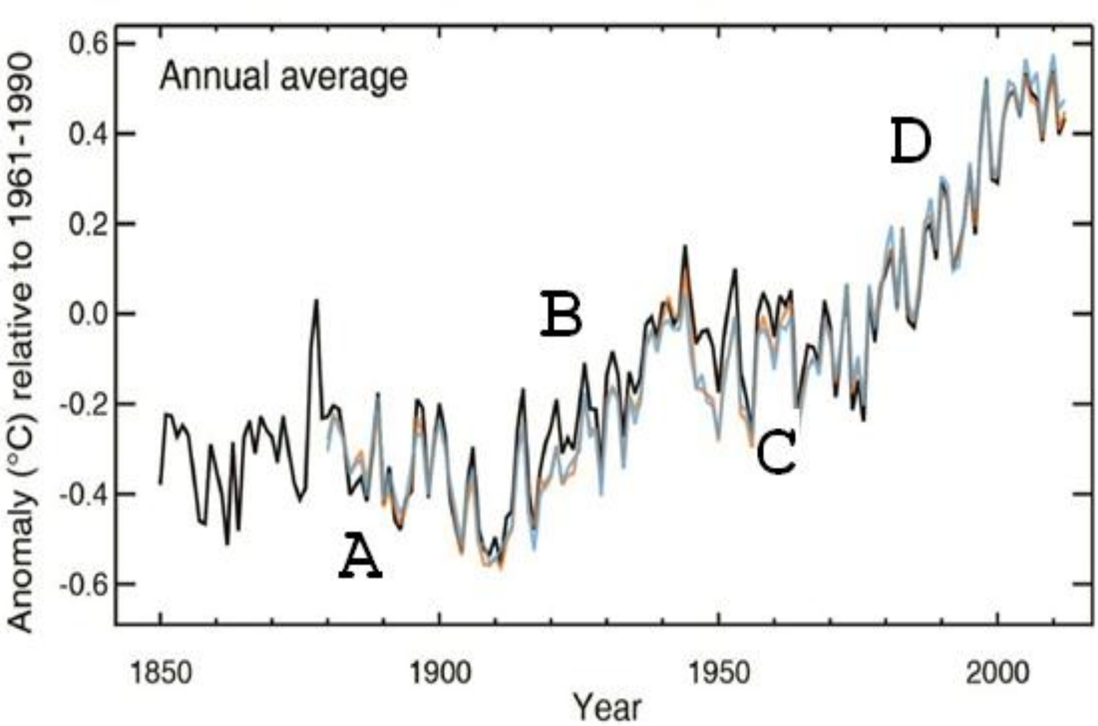


Figure 3.1: Global averaged SSTA (annual) of the period 1850-2012 (5th IPCC Report, 2013). The hiatuses are marked by A and C, whereas warming epochs are marked by B and D.

A recent overview of “Climate change, evidence and causes” by the Royal Society and the US National Academy of Sciences comments on the recent slow-down of global warming from 1998 onwards as “heat is released by oceans into the atmosphere during warm El Niño events, while more heat penetrates to ocean depths during cold La Niña events on a decadal or longer time scales”. Kaufmann et al., (2011) commenting on why global surface temperature did not rise between 1998 and 2008 states that this hiatus in warming coincides with an insignificant increase in the anthropogenic and natural forcing. They found that a decline in the solar insolation from a normal 11 year cycle and a cyclical change from an El Niño to La Niña were quite significant these years. As the recent hiatus in the global warming enters its eighteenth year (2016) scientists are making head way regarding the missing heat during this hiatus (Tollefson, 2014). Oceans are the indicators to explain this anomaly, especially the El Niño of 1997-98 which pumped out substantial amount of heat out of ocean into the atmosphere perhaps enough to tip the equatorial Pacific into a prolonged cold state that has suppressed global temperatures ever since. Trenbreth et al., (2014) analyses based on satellite measurements estimated that aerosols and solar activity accounted for just 20% of the hiatus which leaves the cause for bulk of the hiatus to the equatorial Pacific Ocean. Adding Pacific Ocean data into global climate models can lead to decadal-scale breaks in global warming and when the Pacific Decadal Oscillation (PDO) switches to its positive phase, it heats up the surface ocean and atmosphere, helping to drive decades of rapid warming (Meehl et al., 2011; 2013).

An important question related to the global warming is to what extent ENSO related variations contributed to the observed SST trends. Compo and Sardeshmukh (2010) identified ENSO with the four dynamical eigenvectors of tropical SST evolution that are most important in the observed evolution of ENSO events. This definition is used to isolate the ENSO related component of global SST variations during the 136 year period of 1871-2006. They found that the warming trends over the Pacific, Indian and Atlantic oceans have appreciable ENSO components. In this chapter the possible roles of El Niño/La Niña and also Indian monsoon droughts in creating the phases of rapid warming and hiatus in global warming through the changes in SST between latitudes 30^0S - 30^0N and 80^0S - 80^0N were examined.

3.2 Very dry and wet monsoon years

Very dry monsoon years are selected using the Parthasarathy et al., (1994) derived rainfall series based on the area average rainfall at 306 climatic rain-gauge stations uniformly distributed over India which is shown in Figure 2.1. Year showing a deficiency from the long term average by at least 15% is considered as a very dry year in this study and a year showing excess rainfall of 10% or more is considered as a wet year (Table 3.1). A dry monsoon is followed by a warm SSTA over the Indian Ocean, which persists up to the succeeding monsoon (Joseph and Pillai, 1984). Analysis of historical rainfall pattern shows an inverse relation between El Niño and Indian monsoon rainfall. Dry years in Indian monsoon have generally occurred during strong El Niño events, but such events did not always cause dry monsoons. For example 1997 El Niño was the strongest El Niño of the 20th century, but did not create a dry monsoon. El Niño and dry monsoons occurring in the same year are associated with much larger SSTA in the Indian Ocean (Babu and Joseph, 2002). Thus, frequent dry monsoons can warm the tropical Indian Ocean that is further intensified when it co-occur with El Niño.

3.3 El Niño/La Niña years

The strong El Niños are identified using the condition, if 5- month running mean of SSTA in the N3.4 region (5^0N-5^0S, 120^0-170^0W) exceed +1 ^{0}C for 5 consecutive months or more. Similarly, the strong La-Niñas are selected, if 5- month running mean of SSTA in the N3.4 region (5^0N-5^0S, 120^0-170^0W) are -1^{0}C or less for consecutive 5 months or more (Table 3.1). When El-Niño/La Niña occurs it creates warm/cold SSTA in eastern Pacific Ocean and Indian Oceans.

Table 3.1: Years of very dry and wet monsoons, Strong El Niño and strong La Niña during the period 1950-2010

Years of Very Dry monsoon	Years of Wet monsoon	Years of Strong El Niño	Years of Strong La Niña
1965	1956	1957	1950
1972	1959	1965	1954
1979	1961	1972	1955
1982	1970	1982	1970
1987	1975	1991	1973
2002	1983	1997	1975
2004	1988	2002	1983
	1994		1988
	2007		1998
			2007

3.4 SSTA of composite El Niño, La Niña and Dry monsoon

Since SSTA of tropical oceans are highly seasonal, a composite analysis is very useful to demonstrate the growth and decay of SSTA. The composites of the extreme events of dry monsoon, El Niño and La Niña in the Table 3.1 are given in Figure 3.2.

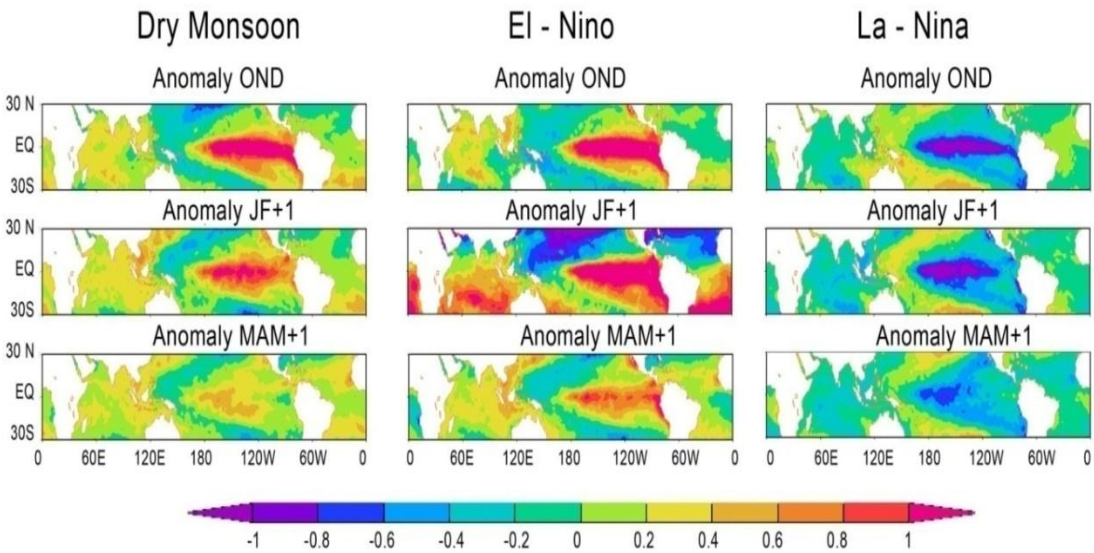


Figure 3.2: SSTA of Tropical Ocean during very Dry monsoons, strong El Niño and strong La Niña during 1950-2008. Composites are made for three seasons OND (October, November, December) during years of strong El Niño/strong La Niña/very Dry monsoon and JF (January, February) and MAM (March, April, May) of the following year

Composites are drawn for three seasons OND (October, November, December), JF+1(January, February+1), MAM+1(March, April, May +1). These months represent the mature state of El Niño and La Niña. From Figure 3.2, it is clear that the dry monsoon SSTA show a similar pattern to El Niño SST anomalies. During an El Niño, warm SSTA occur in the equatorial Pacific Ocean east of 160°E and cold SST anomalies occur in the equatorial west Pacific Ocean, particularly over the tropical northwest Pacific and OND of the El Niño year. During the same period, warm SST anomalies are seen over the equatorial Indian Ocean with a phase lag of 3 to 4 months with the east Pacific warming (Pan and Oort, 1983).

3.5 Warming/cooling after a few cases of El Niño/La Niña

El Niño creates large warm SSTA in equatorial Indo-Pacific basin. This warming is very significant when compared to the normal years, as it warms the surface ocean and the atmosphere above, helping to drive decades of rapid global warming (Meehl et al., 2013). As a consequence, the global surface air temperature may get warmed up by 0.1^0C within 6 months of occurrence of an El Niño (Newell and Weare, 1976; Pan and Oort, 1983; Jones, 1989; Wigley, 2000; Trenberth et al., 2002). However, during exceptional event such as that during 1997-1998 El Niño, the warming was > 0.2^0C. Christy and McNider (1994) and Angell (2000) observed that the entire troposphere had warmed up within a period of 5–6 months, but this lag is slightly less in the tropics and greater at higher latitudes.

In order to study the warming effect caused by El Niño over tropical oceans, four strong El Niño events of this century (1877, 1918, 1982 and 1997) were selected. The El Niño year is taken as El Niño (0) and the following year is taken as El Niño (+1). The average SST of tropical area (30^0S - 30^0N) for El Niño (0) year and El Niño (+1) year were plotted as blue line in Figure 3.3. (left panel) and the dotted red line represents the base period. The El Niño SST minus 15-year base SST were plotted in right panel, shows the increase in temperature during the El Niño years.

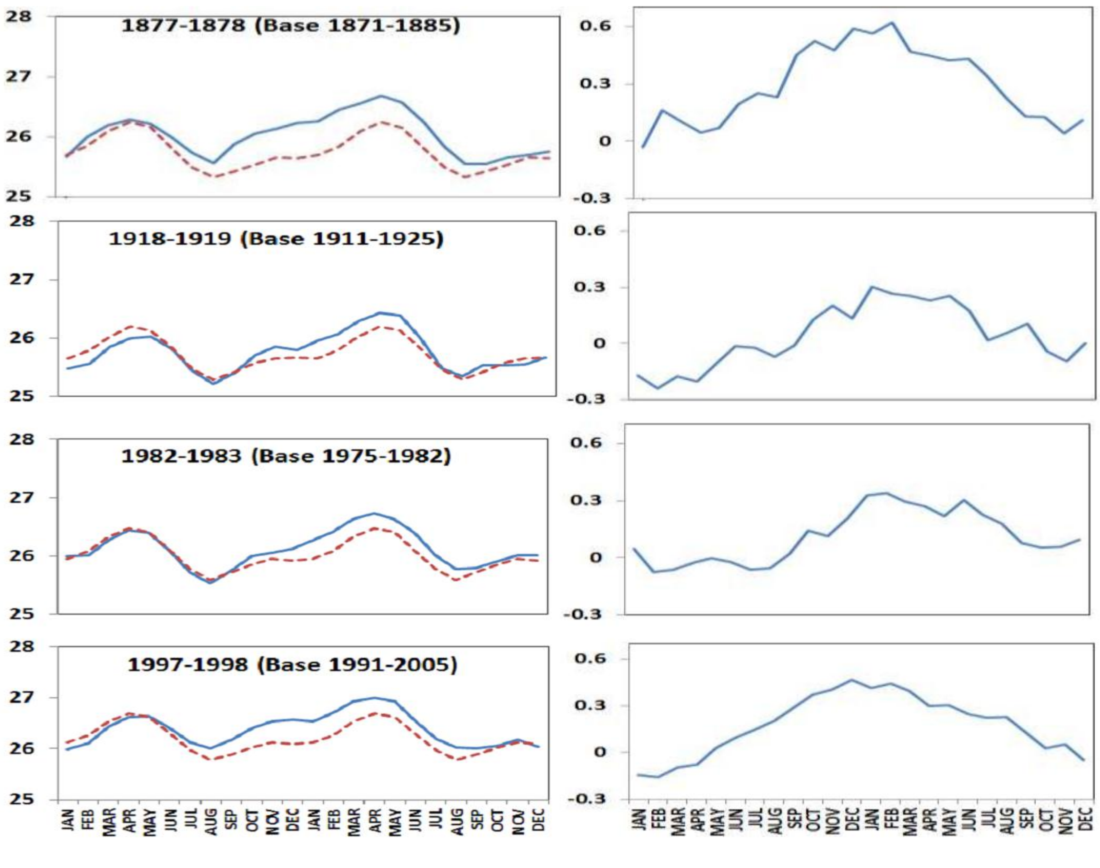


Figure 3.3: SST (^{0}C) averaged over an area between 30^0S - 30^0N for each month of El Niño and the following year. The base period average (broken line) is on the left panel and the monthly anomalies for 4 typical El Niños are on the right panel

Similarly, the four strong La Niña years of this century were selected and similar analysis was done (Figure 3.4). The figure shows that during a La Niña event, the SST decreased in the tropics. Extreme cases (La Niña of 1998) may also lead a cool phase of the PDO in the ocean (Trenberth and Fasullo, 2013). Thus El Niño/ La Niña years contributes an increase/decrease in SST which will reflect the air temperature above.

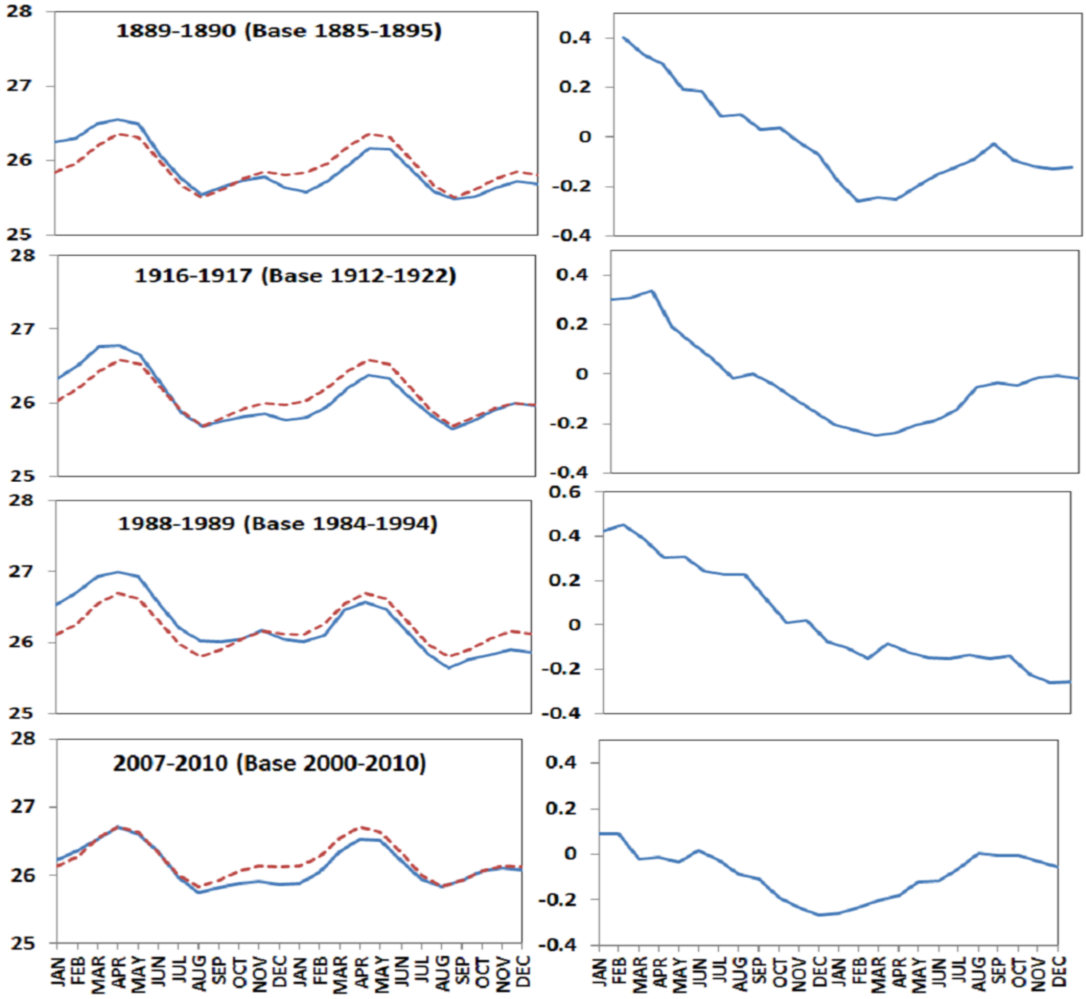


Figure 3.4: The SST (^{0}C) averaged over global area between latitudes 30^{0}S - 30^{0}N for each month of La Niña year and the following year. The base period average (broken line) is on the left panel and the monthly anomalies on the for 4 typical La Niñas are on the right

3.6 Observed warming of the Ocean

SSTs in equatorial oceans between latitude 30^{0}S and 30^{0}N (Figure 3.5a) and global oceans between latitude 80^{0}S and 80^{0}N (Figure 3.5b) for the periods OND (October, November and December) indicate a trend in global warming similar to that seen in Figure 3.1. The warming and cooling superposed epochs (hiatus) can be very well identified in polynomial fit of this time series.

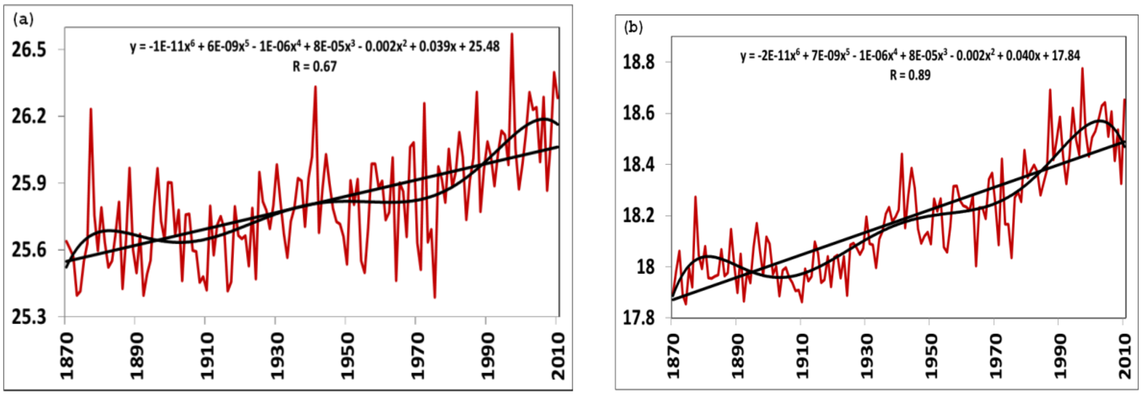


Figure 3.5: Average SST (^{0}C) OND for the area (a) 30^{0}S-30^{0}N and (b) 80^{0}S-80^{0}N (1870-2010)

The warming and cooling trend was more prominent in the equatorial region. El Niño/La Niña which occur in the tropical area may be contributing to warming /cooling effect to this SST variation. The warming and cooling of the ocean is then communicated to the global atmosphere.

Figure 3.6(a) shows that the SSTA of N3.4 area (N3.4 index) is highly correlated with SSTA between 30^{0}S and 30^{0}N (average of 11 years around as base) indicating that the N3.4 index and SSTA of tropical Ocean are strongly related. Using a linear regression equation, the 30^{0}S-30^{0}N SSTA of each year corresponding to the N3.4 index for OND was derived and the cumulative SSTA obtained by adding the SSTA of 1871 to that of 1870 (plotted against 1871) and then adding SSTA of 1872 to this (and plotted against 1872) and so on. Thus 4 epochs marked in Figure 3.1 were recreated in Figure 3.6(b).

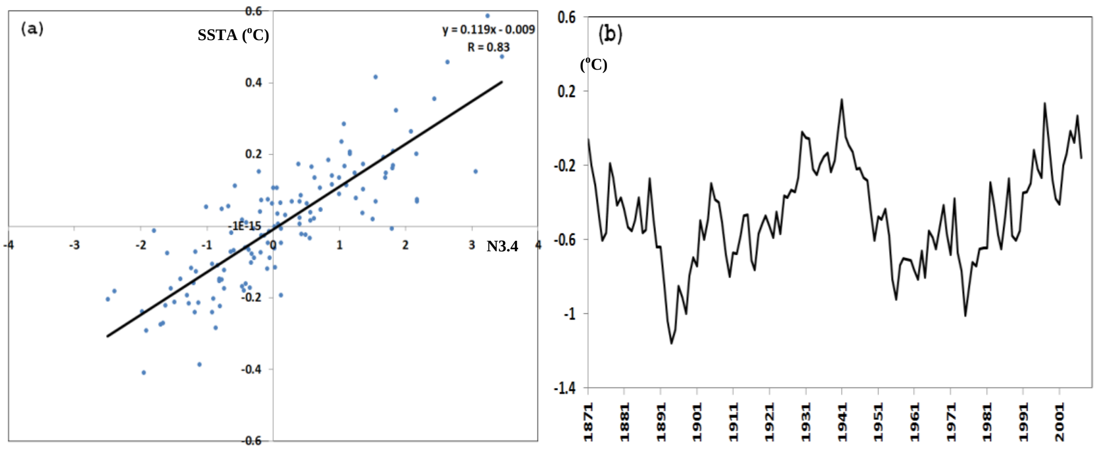


Figure 3.6: (a) Relationship between N3.4 (x-axis) and SSTA (y-axis) for area 30^{0}S - 30^{0}N,
(b) Global warming reconstructed using SSTA for the same area

The same result was obtained for the area 80^0S and 80^0N (Figure 3.7 a and b). The epochs of rapid increase in SST and the hiatus are clearly seen in the figures. This shows that El Niño/La Niña have a great role in creating the phases of rapid increase in SST and the hiatuses. When ocean gets warmed the atmosphere above gets also warmed and when the ocean gets cools the atmosphere above also gets cooled.

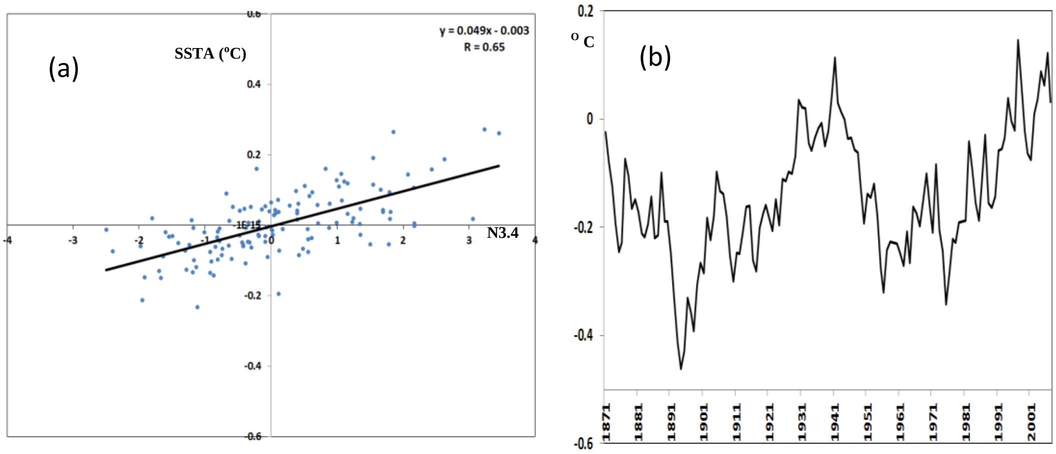


Figure 3.7: (a) and (b) Same as in Figure 3.6 (a) and (b) but for area 80^0S-80^0N

3.7 Decadal variation in ISMR/El Niño

ISMR showed only small long-term trends during the period of recorded rainfall measurements; however, a prominent decadal change was noted (Figure 3.8). During the three-decade long dry epochs of 1901-30 and 1961-90, India experienced drought monsoons on an average once every three years. In contrast, during the three decade long wet epochs of 1871-1900 and 1931-60, the frequency of droughts was approximately one in every 10-15 years. Thus, during the 120 years of 1871 to 1990, alternating dry and wet epochs have occurred regularly. The ENSO phenomenon is considered as the most important external influence on ISMR.

The equatorial Pacific SSTA are negatively correlated with ISMR during monsoon (Rasmusson and Carpenter 1983; Ropelewski and Halpert 1987). Most El Niños are associated with weakening of Indian monsoon and reduction of rainfall. Conversely, the La Niñas are associated with strong Indian monsoon and high rainfall. The thirty year wet epochs of 1871-1900 and 1931-60 coincide with the occurrence of several La Niña and dry epochs 1901-30 and 1961-90 witnessed occurrence of large number of El Niños. These wet/dry epochs of ISMR co-existent with La Niña/El Niño fits very well with the cooling/warming phase of global warming suggesting a natural mechanism for the global warming and its hiatus. But there is a phase shift of a decade between the decadal variations in ISMR and global warming. An epoch of observed frequent droughts was 1960-1990 but the rapid global warming occurred from 1970-1998. Another frequent drought period was 1901-1930, but the rapid global warming was from 1910-1940. The global warming chart (Figure 3.6) produced using the N3.4 index-SSTA regression equation, both for the latitude belts 30^0S-30^0N and 80^0S-80^0N, have been able to show the right timing of the global warming epochs.

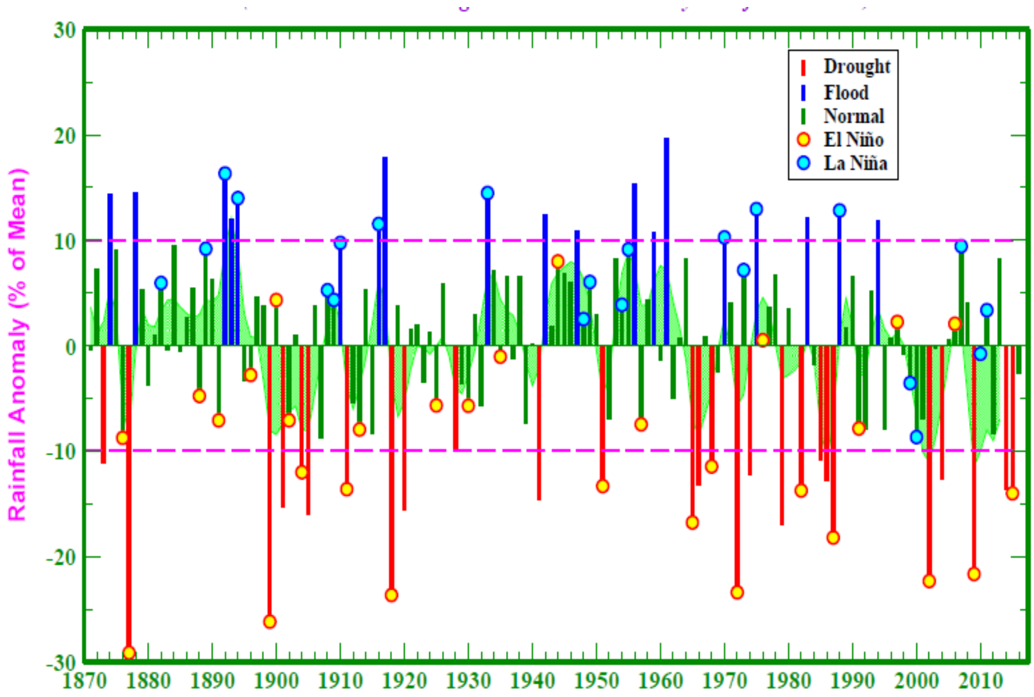


Figure 3.8: ISMR anomaly during 1871-2016. Source: www.tropmet.res.in

3.8 A natural mechanism for global warming proposed

Since the industrial revolution, the greenhouse gas concentration in the atmosphere has been continuously increasing, which has been considered as a cause for the warming trend of globally averaged surface temperature (IPCC Reports, 2007, 2013). However, the long term observations have shown a decadal time scale cooling phases alternating with warming. The rapid global warming periods are found to have more frequent El Niños and dry monsoons. During the hiatuses there were more La Niña and wet monsoons. From the analyses of SST in relation to an El Niño/La Niña Index, it was found that the mean SST of the area between latitudes 30^0S and 30^0N and that between 80^0S and 80^0N showed warm SSTA lasting several seasons from the middle of the El Niño year and cooling lasting several seasons from the middle of the La Niña year. Thus from the present study, observed global warming was found to be the net result of warming due to the increase of greenhouse gases and warming (cooling) due to El Niño (La Niña) events and the associated dry (wet) monsoons.

Chapter-4

Variability of the Date of Monsoon Onset over Kerala and its relation to SST

4.1 Introduction

Monsoon onset is preceded by large scale changes in the atmosphere and ocean in the Indo-Pacific region (Ananthakrishnan et al., 1983; Pearce and Mohanty, 1984; Ananthakrishnan and Soman, 1988; Soman and Krishna Kumar, 1993; Joseph et al., 1994, 2006). Climatologically monsoon sets in over the extreme south of India (Kerala) by the end of May. Accompanying the monsoon onset there is rapid increase in the daily rainfall of Kerala, vertically integrated moisture in the atmosphere and strength (Kinetic Energy) of the low level monsoon flow (Krishnamurti, 1985).

One of the major parameters associated with the date of MOK (DMOK) is the SST over the Indian and Pacific oceans. The centre of the Warm Pool of the tropical oceans gets shifted in the annual cycle from south-west Pacific to the north Indian Ocean from January to May (Joseph 1990a, b, 2014). The Warm Pool of May leads to a build-up of moisture in the atmosphere over north Indian Ocean and the adjoining Pacific Ocean over a month long period and makes conditions favorable for MOK (Pearce and Mohanty, 1984; Joseph et al., 2006).

The long term mean of DMOK is 01 June. This date has varied widely over the years; the earliest was on 11 May in 1918 and the most delayed monsoon was on 18 June in 1972. Joseph et al., (1994) observed that prior to and during MOK dramatic changes occur in the atmosphere and oceans as described earlier by Pearce and Mohanty, (1984) and Krishnamurti (1985). They showed evidence that delays in MOK is associated with El Niño. Analysis of the SST field also showed that delayed MOK is associated with warm SSTA at and south of the equator in the Indian and Pacific oceans and cold SSTA in the tropical and subtropical oceans to the north during the season prior to the monsoon onset (March to May). They hypothesized that such SSTA caused the inter-annual variability of MOK through their action in affecting the timing of the northward movement across the equator of the cloud band associated with the ITCZ. DMOK has been defined using different criteria that include rainfall, outgoing long wave radiation (OLR), lower tropospheric winds etc (Ananthakrishnan and Soman, 1988; Joseph et al., 2006; Wang et al., 2009).

The isoplets of the normal onset of monsoon over India were prepared by India Meteorological Department (IMD) from 1943 onwards and these dates are declared on subjective estimates prepared by operational forecasters. Following quote from Ananthakrishnan and Soman (1988) is relevant in this contest. ‘‘Although the onset of monsoon is associated with changes in the atmospheric circulation features in the lower and upper troposphere, a sustained increase in the rainfall at the observatory stations of Kerala and the island stations over the south-east Arabian sea is an essential feature of the monsoon onset. It is difficult to quantify these precisely and so the experience of the forecaster plays a key role in declaring the date of monsoon onset in individual years.’’

The IMD, while declaring the date of the MOK, has been taking into consideration subjectively the following features (Rao 1976): i) the rainfall should be widespread over Kerala and adjacent areas, with large amounts at individual stations; ii) this rainfall persists over several days; iii) the lower-tropospheric westerlies in and around Kerala should be strong and deep; and vi) the air should be rich in moisture (high relative humidity) up to at least 500 hPa. For several decades till the 1980s, IMD was considering all these factors in a subjective way to fix the date of MOK. But during the 1990s and up to 2005, IMD declared monsoon onset when after 10^{th}May a large percentage of the synoptic stations of Kerala and the island stations of southeast Arabian Sea reported rainfall for two consecutive days as per the rain only criteria for monsoon onset given by Ananthakrishnan et al., (1967) without taking into account the other factors.

In 2006 IMD adopted an objective criteria for declaring MOK using criteria derived by Joseph et al., (2006) (as adapted by Pai and Rajeevan, 2009) based on daily rainfall of Kerala, the depth of westerlies in a box (equator to latitude 10^0N and longitude 55-80^0E), the zonal wind speed over the area bounded by latitude 5-10^0N and longitude 70-80^{0}E and the mean OLR in the box latitude 5-10^{0}N and Longitude 70-75^{0}E. In both the subjective and objective methods, a sharp increase in rainfall of Kerala during MOK is important.

Li and Yanai (1996) showed that the onset of Asian summer monsoon is concurrent with the reversal of the meridional temperature gradient in the upper troposphere south of the Tibetan plateau. This reversal is the result of the large temperature increase in May - June over Eurasia centered on the plateau. The Tibetan heat source according to them is mainly contributed by the sensible heat flux from the ground surface. They however did not study the inter-annual variability of the DMOK. On the lines of their study Xavier et al., (2007) found that the meridional gradient of the tropospheric temperature (averaged between 600 and 200 hPa) is proportional to the meridional gradient of deep tropospheric heating and could lead to acceleration of the deep tropospheric circulation. Their objective definition of the large scale monsoon onset (over India and not Kerala) is based on the reversal of GrTT, (Gradient in Tropospheric Temperature as average of 600 to 200 hPa) between a northern box (40-100^{0}E, 5-35^{0}N) and a southern box (40-100^{0}E, 15^{0}S-5^{0}N) denoted by GrTT. The onset date (GrTT onset) is defined as the date when GrTT changes sign from negative to positive.

Fasullo and Webster (2003) used the vertically integrated moisture transport through India to define the date of monsoon onset over India (not Kerala). The authors claimed that their index is indicative of the transition in the large scale monsoon circulation over India. Joseph (2013) has compared the objective dates of monsoon onset over Kerala (Pai and Rajeevan, 2009) with the monsoon onset dates over India as given by Xavier et al., (2007) and Fasullo and Webster (2003) using data of the period 1971-2000 and found that the linear correlation coefficients between them are not large (0.59 and 0.56, respectively).

In a very recent publication by Oronez et al., 2016 objective dates have been derived for MOK using surface wind direction data collected by merchant ships plying over the Indian Ocean during the period 1877 to 2013, excluding several years where wind data was scarce due particularly to world wars. The above MOK data sets have been compared with IMD objective onset dates of the period from 1971 and is given in Table 4.2.

The details of the data sets used for this study is described in chapter 2. The data of the date of monsoon onset for north Kerala for the period 1870-1900 and for Kerala for 1901 -2014 are given in Table 4.1.

The 11 year moving averages for the different series used for the study are given in Figure 4.1(a). The onset dates for north Kerala by Ananthakrishnan and Soman (1988) is available for a period 1870-1980. The subjective onset dates derived by IMD is available for the period 1901-1980. The mean difference in onset dates between the above two series is 1 day, north Kerala onset date being the earlier one. For declaring monsoon onset, IMD's subjective criteria stipulates two consecutive days of rainfall over Kerala and the onset is declared on the second day. Thus, there is really no difference in the onset dates of these two series. The mean and SD of the three series is given in Table 4.2.

To construct a long series for the DMOK for the period 1870 to 2014, the following method was adopted. For the period 1870 to 1900 we have added 1 to the onset dates for north Kerala as derived by Ananthakrishnan and Soman (1989). The subjective onset dates for Kerala as derived by IMD for the period 1901 to 1970 is merged to the above series. To this the objective dates from 1971-2014 of MOK derived by IMD is added. Thus the time series of DMOK for the period 1870-2014 is constructed and is given in Table 7.1.

There were no data set for DMOK spanning 1870-2015 till now. Though this data is merged by most reliable way it can be used for climatic studies in future.

Table 4.1: Date of monsoon onset over Kerala (DMOK); NK-North Kerala; IMD (S) - IMD subjective; IMD (O) - IMD objective. Dates are counted in days from 01 May as follows: 01 May=01; 01 June=32; etc

YEAR	N K	IMD (S)	IMD (O)	YEAR	N K	IMD (S)	IMD (O)	YEAR	N K	IMD (S)	IMD (O)
1870	34	-	-	1901	36	38	-	1932	15	33	-
1871	31	-	-	1902	37	37	-	1933	22	22	-
1872	32	-	-	1903	43	43	-	1934	39	39	-
1873	23	-	-	1904	32	38	-	1935	45	43	-
1874	16	-	-	1905	39	41	-	1936	21	19	-
1875	34	-	-	1906	44	45	-	1937	34	35	-
1876	39	-	-	1907	37	39	-	1938	26	26	-
1877	38	-	-	1908	41	42	-	1939	37	36	-
1878	40	-	-	1909	33	33	-	1940	45	45	-
1879	17	-	-	1910	37	33	-	1941	22	23	-
1880	26	-	-	1911	35	37	-	1942	41	41	-
1881	40	-	-	1912	36	39	-	1943	13	29	-
1882	31	-	-	1913	35	33	-	1944	34	34	-
1883	34	-	-	1914	35	35	-	1945	36	36	-
1884	40	-	-	1915	42	46	-	1946	34	29	-
1885	35	-	-	1916	28	33	-	1947	33	34	-
1886	31	-	-	1917	30	31	-	1948	40	42	-
1887	32	-	-	1918	8	11	-	1949	13	23	-
1888	28	-	-	1919	35	34	-	1950	27	27	-
1889	30	-	-	1920	34	34	-	1951	32	31	-
1890	28	-	-	1921	37	33	-	1952	33	20	-
1891	33	-	-	1922	31	31	-	1953	48	38	-
1892	26	-	-	1923	41	42	-	1954	32	31	-
1893	26	-	-	1924	32	33	-	1955	16	29	-
1894	35	-	-	1925	27	27	-	1956	20	21	-
1895	43	-	-	1926	38	37	-	1957	18	32	-
1896	34	-	-	1927	27	27	-	1958	44	45	-
1897	37	-	-	1928	34	34	-	1959	15	31	-
1898	35	-	-	1929	32	29	-	1960	15	14	-
1899	27	-	-	1930	38	39	-	1961	20	19	-
1900	41	-	-	1931	30	35	-	1962	10	17	-

YEAR	N K	IMD (S)	IMD (O)	YEAR	N K	IMD (S)	IMD (O)
1963	35	31	-	1994	-	28	28
1964	35	37	-	1995	-	39	41
1965	37	26	-	1996	-	34	40
1966	31	31	-	1997	-	40	43
1967	40	40	-	1998	-	33	34
1968	40	39	-	1999	-	25	22
1969	32	17	-	2000	-	32	32
1970	26	29	-	2001	-	23	26
1971	25	27	27	2002	-	29	40
1972	53	49	48	2003	-	39	44
1973	37	35	36	2004	-	18	34
1974	23	26	23	2005	-	38	38
1975	31	31	33	2006	-	-	26
1976	31	31	30	2007	-	-	28
1977	38	30	29	2008	-	-	31
1978	29	28	29	2009	-	-	23
1979	43	42	43	2010	-	-	31
1980	31	32	34	2011	-	-	29
1981	-	30	30	2012	-	-	36
1982	-	30	30	2013	-	-	32
1983	-	44	43	2014	-	-	37
1984	-	31	32				
1985	-	28	24				
1986	-	35	43				
1987	-	33	32				
1988	-	26	33				
1989	-	34	35				
1990	-	19	18				
1991	-	33	33				
1992	-	36	36				
1993	-	27	34				

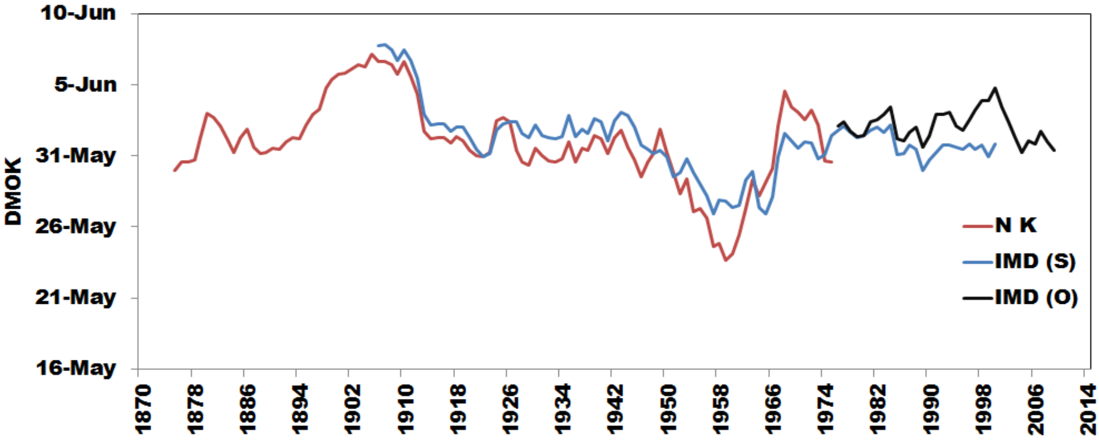


Figure 4.1 (a): 11 year moving averages of dates of monsoon onset over North Kerala (red curve), IMD monsoon onset over Kerala (Subjective) (blue curve) and IMD monsoon onset over Kerala (Objective) (black curve)

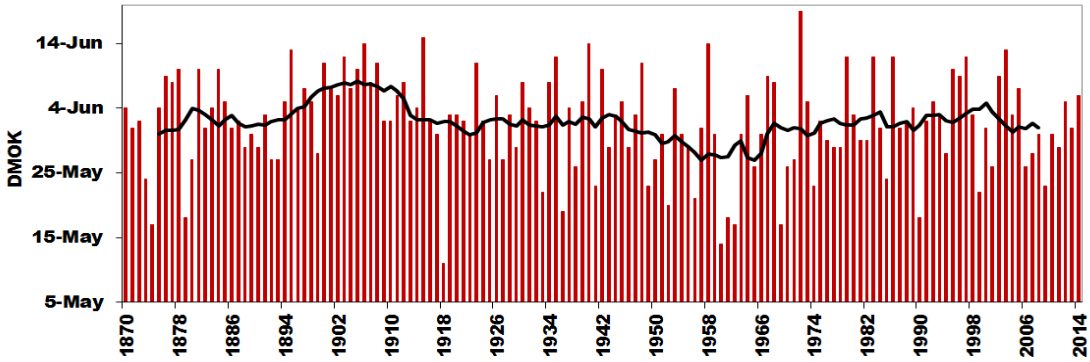


Figure 4.1(b): DMOK for the period 1870-2014 combining the three series given at 4.1(a) is shown in red bars and their 11 year moving average is marked by black line. The statistical properties of above mentioned series is given in Table 4.2

Table 4.2: Mean and Standard deviation of different series. NK-North Kerala; IMD (S) - IMD subjective; IMD (O) - IMD objective

SERIES	MEAN	S D
NK - 1870 to 1900	01 June	6.5 days
IMD (S) –1901to 1970	01 June	7.9 days
IMD (O) - 1971 to 2014	01 June	6.7 days
NK+1 - 1870 to1900+ IMD (S) - 1901 to 1970 + IMD(O) - 1971 to 2014	01 June	7.2 days

The SST gradient in the Indo-Pacific region during the season March-April also reflects a similar pattern of inter-annual variability of DMOK which is presented in Figure 4.1(c). This long time series of DMOK is represented by the red bars in Figure 4.1(b) and their 11-year moving average by the black line.

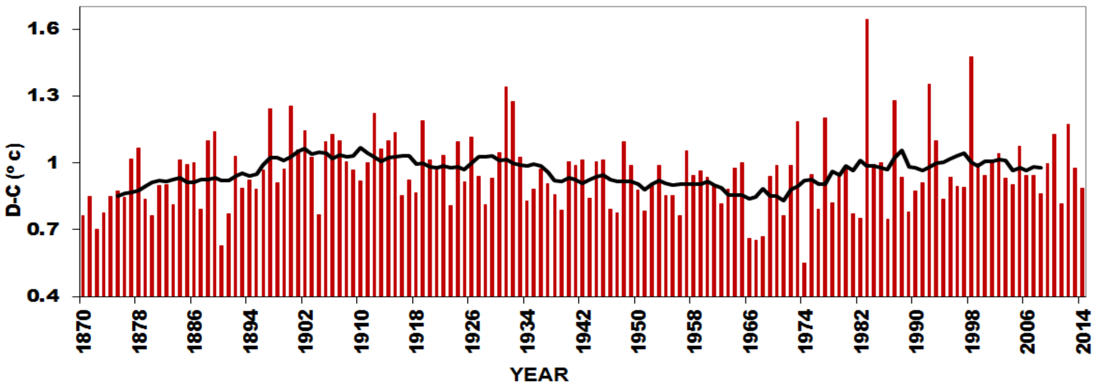


Figure 4.1: (c) SST difference for each year of 1870-2014 between the latitude longitude boxes 30^0E-270^0E, 0^0-20^0S (D) and 30^0E -270^0E, 5^0N-20^0N (C) for the season Mar-Apr and its11 year moving averages

The mean DMOK of 1870 to 2014 is June 1st and the standard deviation (SD) is around 7 days. Those who use this long data set for climate change studies should know that the onset dates of the period 1870 to 1970 have not been corrected for bogus onset occurrences. Onset dates of this period have been derived without using low level wind or satellite data of convection which are needed for the whole Indian Ocean basin to identify the pre-monsoon rain peak or bogus monsoon onset (which occurs once in about ten years) as described in Joseph et al., (2006). In this connection a study by Sabeerali et al., (2012) on the withdrawal dates is relevant and they observed a clear shift prior (posterior) to the 1976/77 climate shift most of the withdrawal dates are associated with a late/early. However in this study no shift was observed in the DMOK.

Ordonez et al., (2016) have derived a method to eliminate bogus onset dates of monsoon using surface wind direction as recoded by merchant ships plying in Indian Ocean. It is compared the MOK dates derived by them (called ship wind data) with the objective MOK dates derived by IMD. The difference between these dates (IMD objective onset dates-ship wind onset dates) is positive in most of the years and large positive in several years as may be seen in Table 4.3, hence unable to use this long period data set provided by the authors to derive a long data set of objective DMOK.

Table 4.3: Difference in the days between the IMD Objective minus Ship wind of each year 1971- 2005. (1971 data is not there in ship wind data.)

IMD Objective onset date minus Ship wind onset date				
-5 and below	-4, -3, -2	-1, 0, +1	+2, +3, +4	+5 and above
	1982	1974,1987	1975,1977	1972(06days)
		1992,2005	1979,1983	1973 (10 days)
			1984,1985	1976 (05 days)
			1988,1990	1978 (10 days)
			1991,1993	1980 (06 days)
			1994,1997	1981 (05 days)
			1998	1986 (06 days)
				1989 (11 days)
				1995 (07 days)
				1996 (06 days)
				1999 (10 days)
				2000 (18 days)
				2001 (08 days)
				2002 (17 days)
				2003 (08 days)
				2004 (21 days)
	1 Year	4 Years	13 Years	16 Years

4.2 Inter-annual variability of MOK and its relation with SSTA

SST plays a vital role in the timing of monsoon onset over Kerala (Joseph et al., 1994, 2006; Joseph 2014). SSTA on either side of equator in Indian and Pacific Oceans were found to be related to the date of MOK. In this study MOK is considered as delayed if it has occurred 8 days or more (one standard deviation) after 01 June. In the same way, if DMOK is eight days or more before 01 June, it is defined as early. Joseph (2014) found the linear correlation coefficient (LCC) between the objective dates of MOK and the SST using 30 years of data (1971-2000). Longer period data were used in this study and found the LCC between IMD's objective DMOK and SST of January-February and March-April for the period 1971-2014 which is given in Figure 4.2.

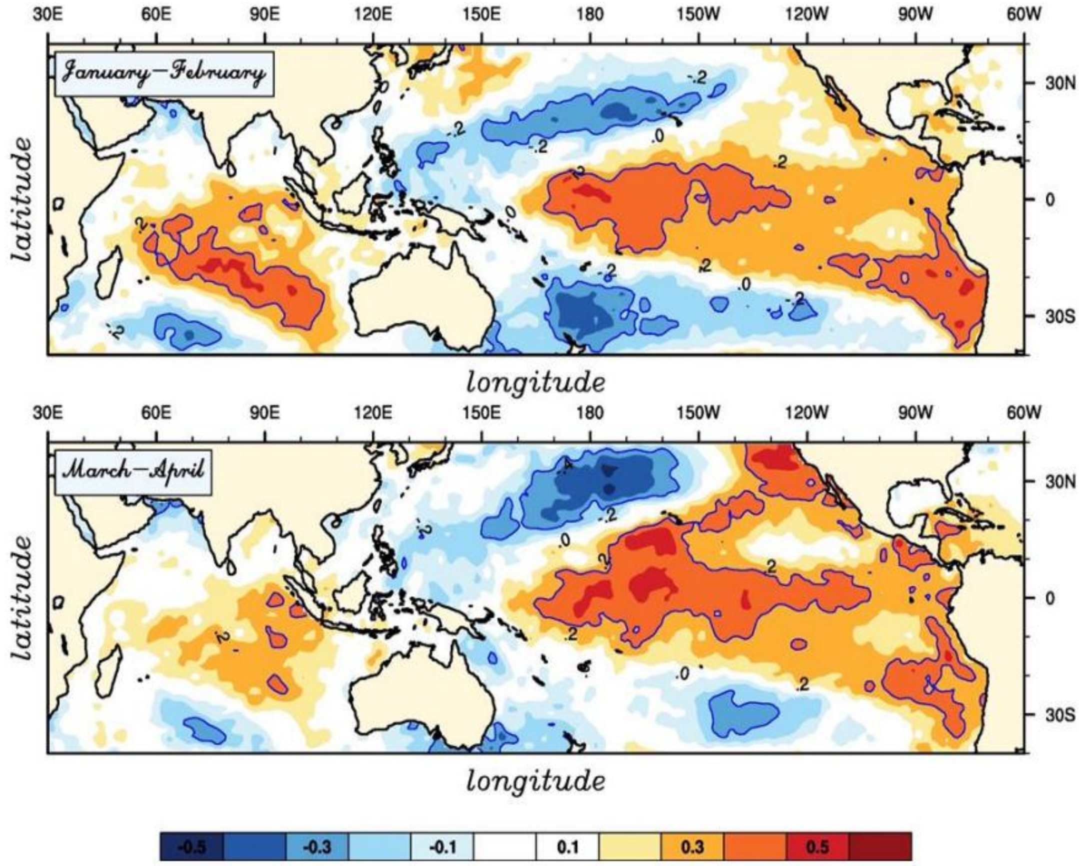


Figure 4.2: Linear Correlation coefficient between HadISST and IMD's objective dates of MOK from 1971- 2014 for January-February (upper panel) and March-April (lower panel). The significant values are marked by thick blue line. Linear correlation coefficient is 0.3 for a significance level of 95 % by t-test and r=0.3 is marked by a blue line

The LCCs are high and statistically significant over large spatial areas marked by the blue lines in the figure which gives correlation coefficient of 0.3 which is significant at 95% level using t-test. MOK is delayed when the SSTA is positive south of the equator and negative north of it. This correlation pattern was also persistent for several months prior to MOK. The LCC patterns over the Pacific and Indian oceans show that delayed MOK is associated with El Niño where as an early MOK is associated with La Niña. The association of delayed MOK with El Niño has been shown by Joseph et al., (1994). The composite SSTA of January-February and March-April of delayed and early DMOK years of the period 1901 to 2014 are given in Figure 4.3.

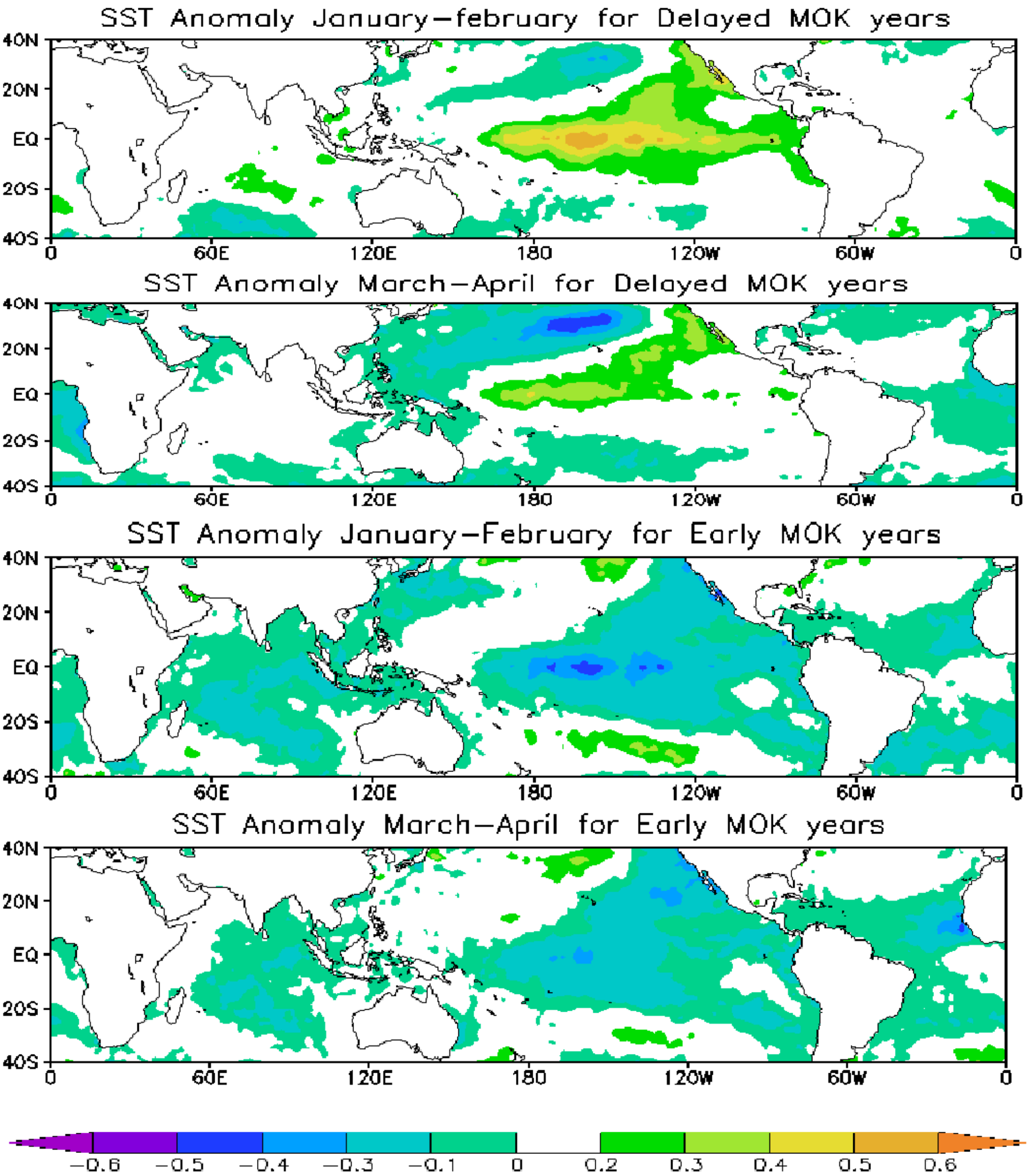


Figure 4.3: The composite SSTA of the Jan-Feb and Mar-Apr of delayed and early DMOK years of the period 1901 to 2014

According to the study by Joseph et al., (1994), the monsoon onset delays occur in El Niño(0) year and more often in El Niño(+1) year. During the period 1870-2014 there were 21 cases of monsoon onset delays of one standard deviation or more and 13 out of these 21 cases occurred in El Niño(0) or (+1) year (Table 4.3). For compositing the SSTA were derived by subtracting an 11-year mean SST centered over the year for each case. The SST anomalies are similar to those derived from the correlation study in Figure 4.3. Further a first EOF mode of SST over Indo-Pacific domain and this mode depict El Niño related SSTA in Pacific Ocean is computed (Figure 4.4).

Table 4.4: Years of delayed MOK during 1901 to 2014 The El Niño year is marked by (0) and the following Year by (+1)

Years of delayed MOK		
1903(+1)	1940(0)	1983(+1)
1905(0)	1942(+1)	1986(0)
1906	1948	1995(+1)
1908	1958(+1)	1996
1915	1967	1997(0)
1923(0)	1972(0)	2002(0)
1935	1979	2003(+1)

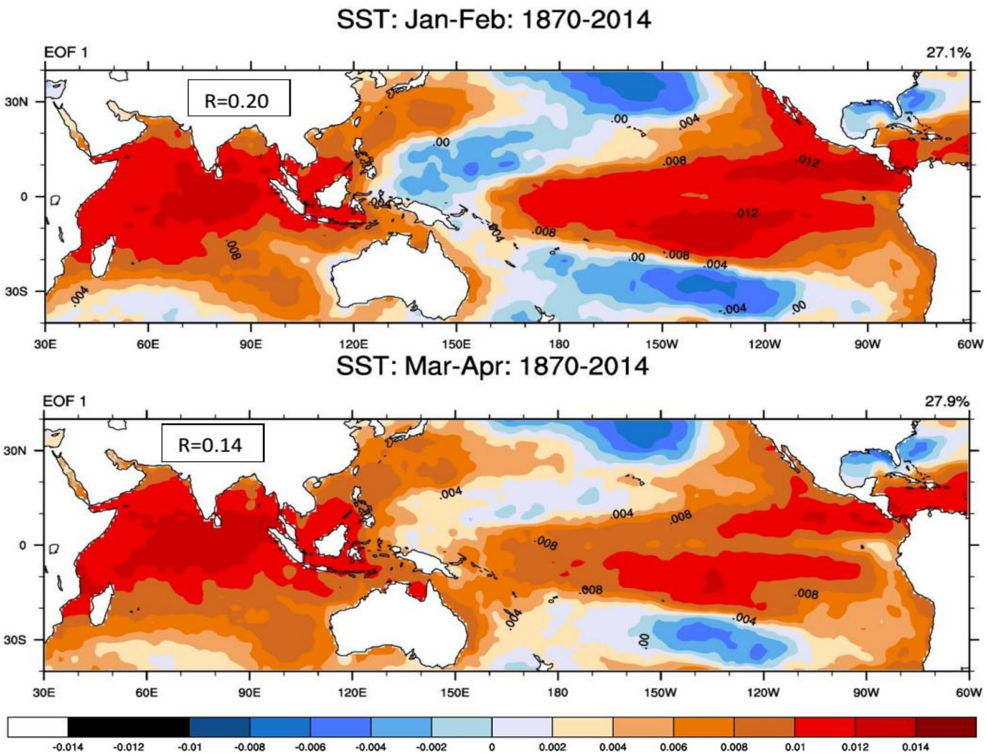


Figure 4.4: EOF1 of SST over Indo-Pacific domain for the season Jan-Feb (upper panel) and for the season Mar-Apr (lower panel). The correlation of time series of first PC (principal component) with monsoon onset dates marked by R

The time series of this anomaly has statistically significant correlation of 0.2 with the monsoon onset dates of 1870-2014, for the Jan-Feb season. For the Mar-Apr season the correlation is smaller. However the correlation between dates of monsoon onset and N3.4 index of Jan-Feb and Mar-Apr seasons are statistically significant at 95% level (Figure 4.5).

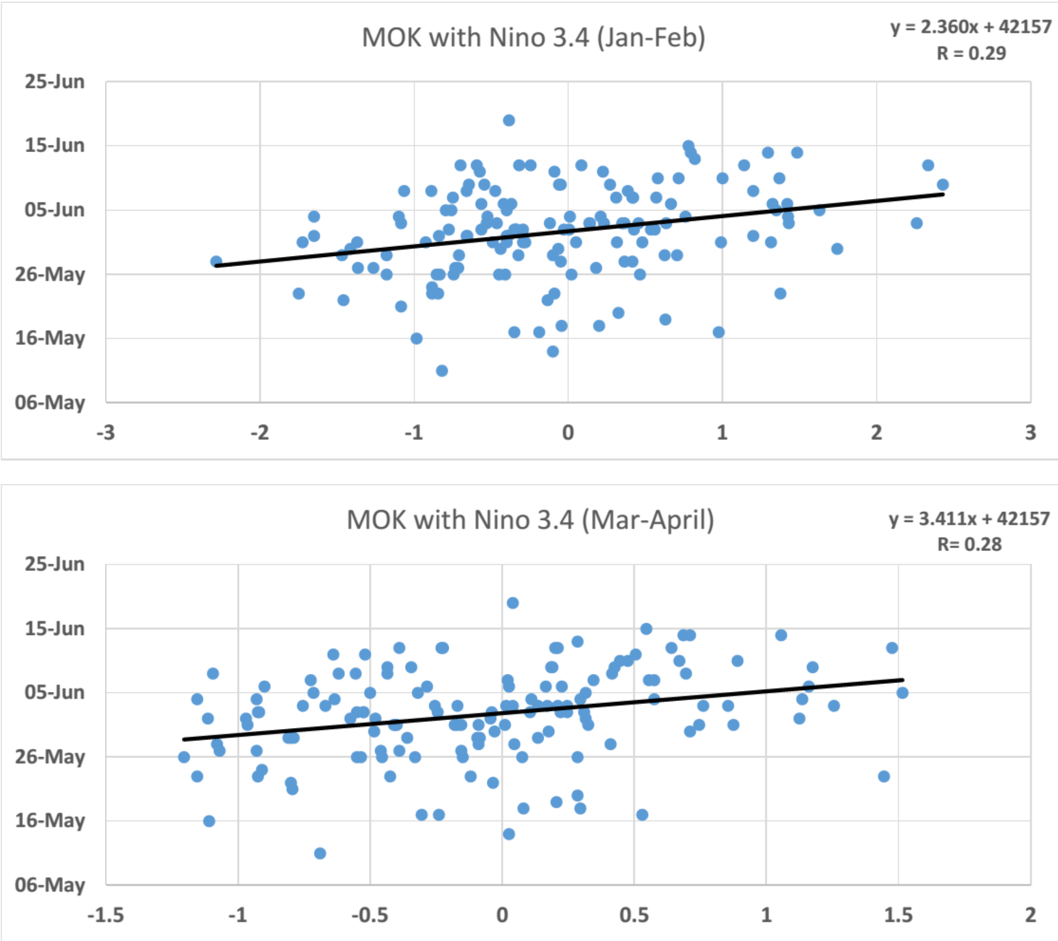


Figure 4.5: Correlation between dates of monsoon onset and Niño 3.4 index of Jan-Feb (above) and Mar-Apr (below)

The spatial pattern of EOF-1 shows that in delayed onset years there is warm SSTA south of the equator in the Indian and Pacific oceans. EOF-1 is similar to the Figure 4.4 showing the correlation between date of monsoon onset and SST.

4.3 Evolution of OLR in delayed and early MOK

As described in section 4.1: the center of the Warm Pool shifts in the annual cycle from south west Pacific Ocean to north Indian Ocean from January to May. But there is a fine structure to the changes in SST and the associated convection (OLR) in northern Indian Ocean prior to MOK. The fine structure in the changes in SST for the MOK of 2003 has been described by Joseph (2014). SST of north Indian Ocean reaches maximum first over the BoB, about 7-8 pentads before the date of MOK, when SST there reaches high values of 31^{0}C or even 32^{0}C in the central BoB. To the south of the region of this SST maximum, in the area of large SST gradient, deep convection develops and grows in area and intensity. The latent heat released in this convective area heats the atmosphere there and generates strong cross equatorial low level wind flow. The convective clouds and the low level winds cool the BoB. When these changes are happening in the BoB, the Arabian Sea has much lower SST with very little convection and only feeble low level winds there which make the SST of Arabian Sea warm rapidly. About 3 pentads before MOK, SST over the central Arabian Sea reaches 31^0C or 32^0C and the SST gradient area south of this region of convection, and strong low level winds develop the northward movement of which brings about MOK.

The evolution of convection (OLR) prior to MOK for the delayed and early monsoon onset years was also studied. Composite pentad OLR charts are made for the delayed onset years (1979, 1983, 1986, 1995, 1996, 1997, 2002 and 2003) and for the early onset years (1974, 1985, 1990, 1999 and 2009) for several pentads prior to MOK with MOK taken as occurring around zero pentad.

Figure 4.6(a) shows the composite mean pentad variations in the OLR of delayed onset years of the period 1974-2014 (see Table 4.4). Zero pentad is that of MOK. Coinciding with pentad -10 (22 - 26 April) there is deep convection (low values of OLR) around equatorial BoB extending to the west Pacific Ocean. This band of convection increases in size and moves north up to pentad -7 (07 - 11 May) and it's left most portion brings rain over Kerala. This was called the Pre- Monsoon Rainfall Peak (PMRP) by Joseph and Pillai (1988) and "Bogus Monsoon onset" by Flatau et al., (2001) associated with strong low level winds. Soon after the PMRP the band of convection moves north eastward to usher in monsoon onset in some years over the Indo-China peninsula at pentads -6 and -5 as shown by Joseph et al., (2006). At pentad -4 (22 - 26 May), the OLR field suggests that Indian Ocean is practically free of convection. At pentad -3 (27 - 31 May) a fresh elongated narrow band of convection formed close to the equator (in the Indian Ocean south of the Arabian Sea). This band of convection grows rapidly in intensity and area as seen by the OLR of pentads -2 (01 - 05 June) and -1 (06 - 10 June), bringing about the monsoon onset over Kerala at zero pentad (11 - 15 June).

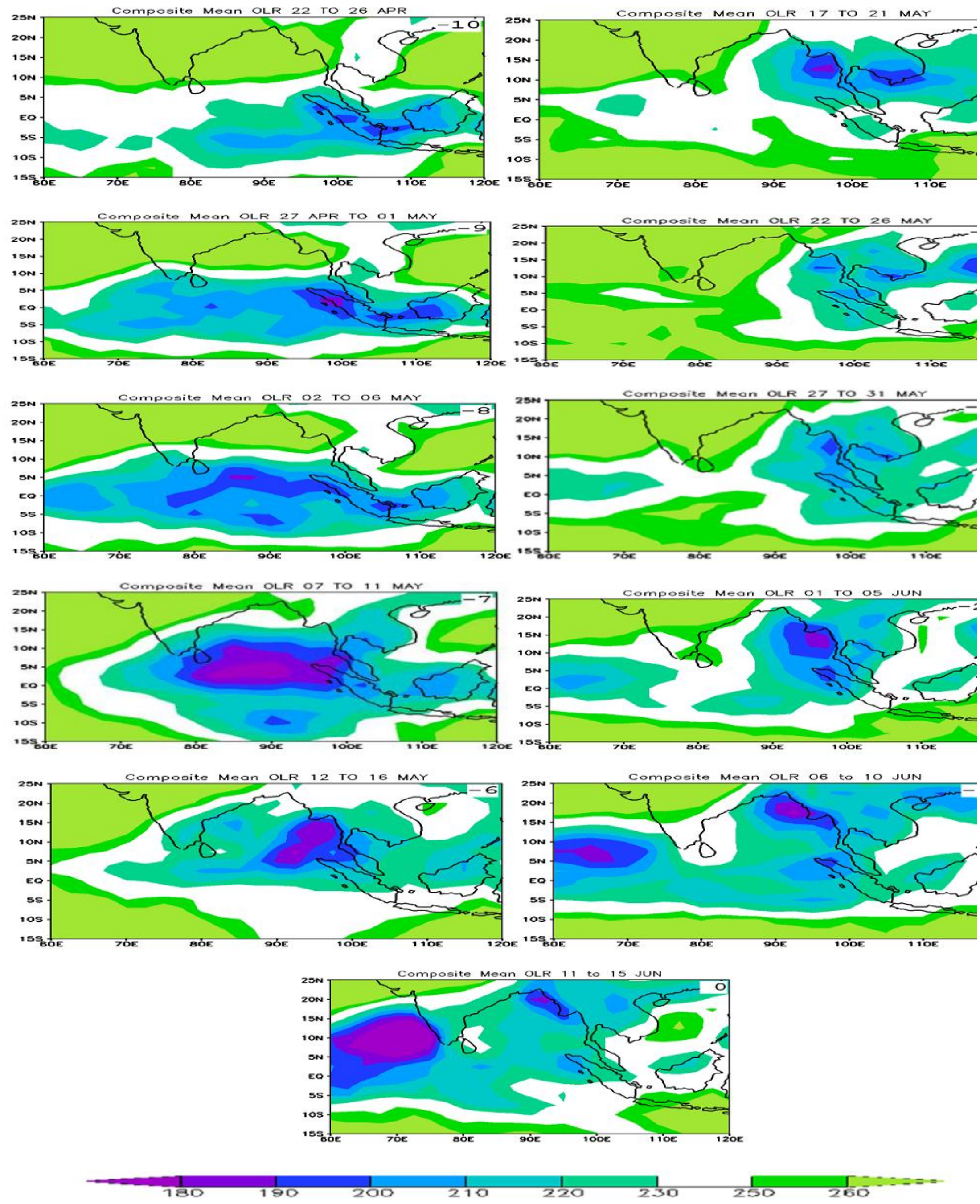


Figure 4.6: (a) The composite mean OLR (W m-2) for delayed MOK years (1979, 1983, 1986, 1995, 1996, 1997, 2002 and 2003) (pentads are marked on top right corner, MOK is taken as zero pentad)

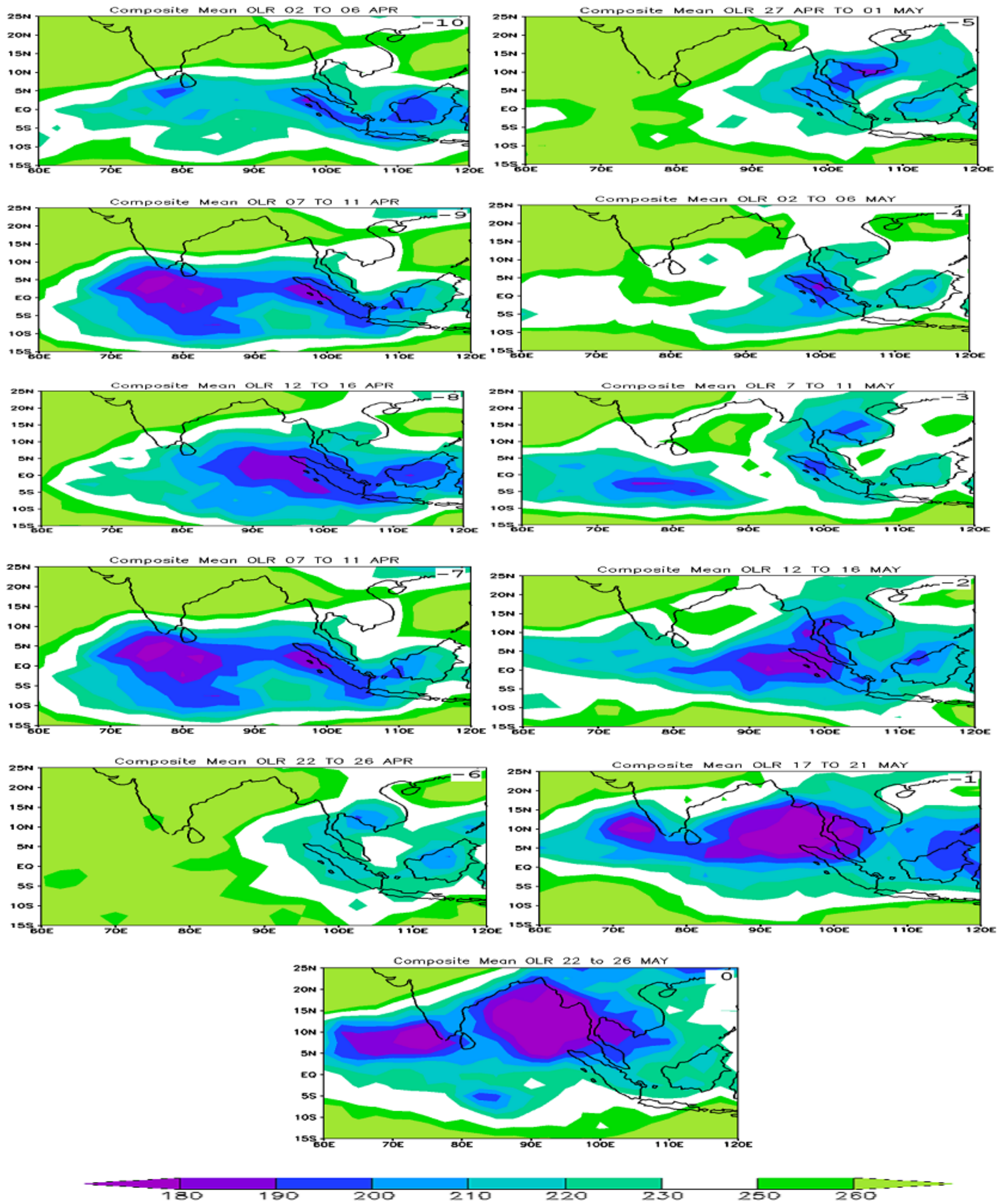


Figure 4.6: (b) The composite mean OLR (W m^{-2}) for early MOK years (1974, 1985, 1990, 1999 and 2009) (pentads are marked on top right corner, MOK is taken as zero pentad)

The composite mean OLR for early onset years for selected pentads is given in Figure 4.6(b).The development and migration of the convection bands for the early MOK years is similar pentad wise to delayed MOK years, but the PMRP and MOK are found to occur four pentads earlier date wise. Thus, PMRP for early MOK composite is on pentad 7-11 April and MOK on pentad 22-26 May. The composite OLR chart for normal onset years (figure not shown) was also constructed. The pentad wise evolution of OLR with zero pentad taken as DMOK is the same as for late and early onset composites. These three composites show that there is a minimum of convection in north Indian Ocean during pentads -5, -4 and -3.

4.4 Why convection occurs in the BoB earlier than in the Arabian Sea?

A question arises as to why SST reaches maximum in the annual cycle January-May, first in the BoB followed by in the Arabian Sea nearly a month later. Figure 4.7 (a and b) gives the monthly mean SST averaged over the period 1974-2014 for the months of March and April.

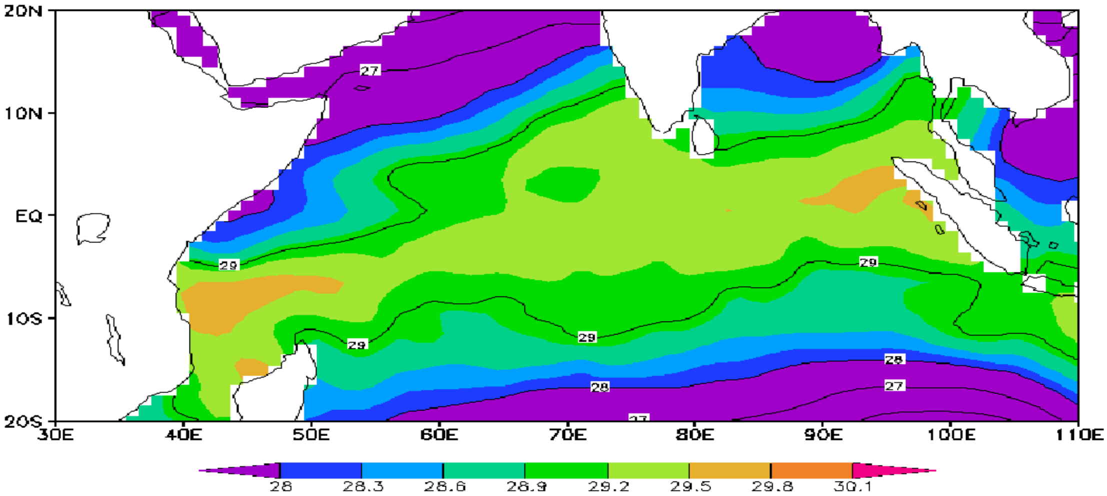


Figure 4.7: (a) Mean SST (°C) (1974-2014) for March

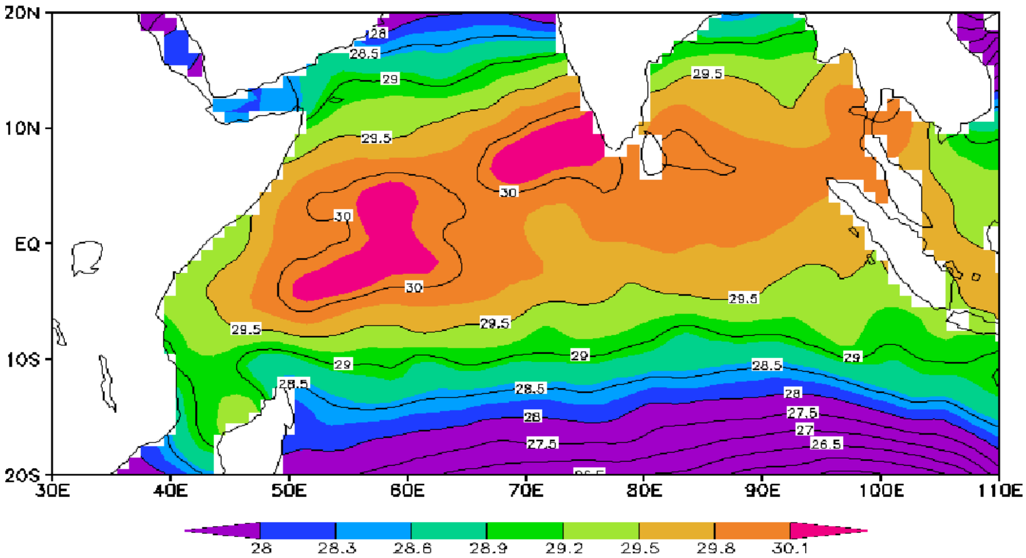


Figure 4.7: (b) Mean SST (°C) (1974-2014) for April

The axis of maximum SST is oriented from south-west to north east (in west Indian Ocean the SST maximum is close to the equator and in the east Indian Ocean it is at about latitude 10^0N). It is speculated that this south-west to north-east orientation is taken by the axis of SST maximum due to the strong cooling of SST of the Arabian Sea during the previous monsoon season. The strong low level winds of the monsoon are known to cool the Arabian Sea by causing coastal and open ocean upwelling and by evaporative cooling of the ocean surface. Figure 4.8 (a) shows the day to day change of the mean SST of the BoB Box (red curve) bounded by latitudes 10^0N and 20^0N and longitudes 85^0E and 95^0E and the Arabian sea box (blue curve) bounded by latitudes 10^0N and 20^0N and longitudes 60^0E and 70^0E. The BoB box reaches the SST maximum first around mid-April whereas the Arabian Sea box reaches the SST maximum about a month later.

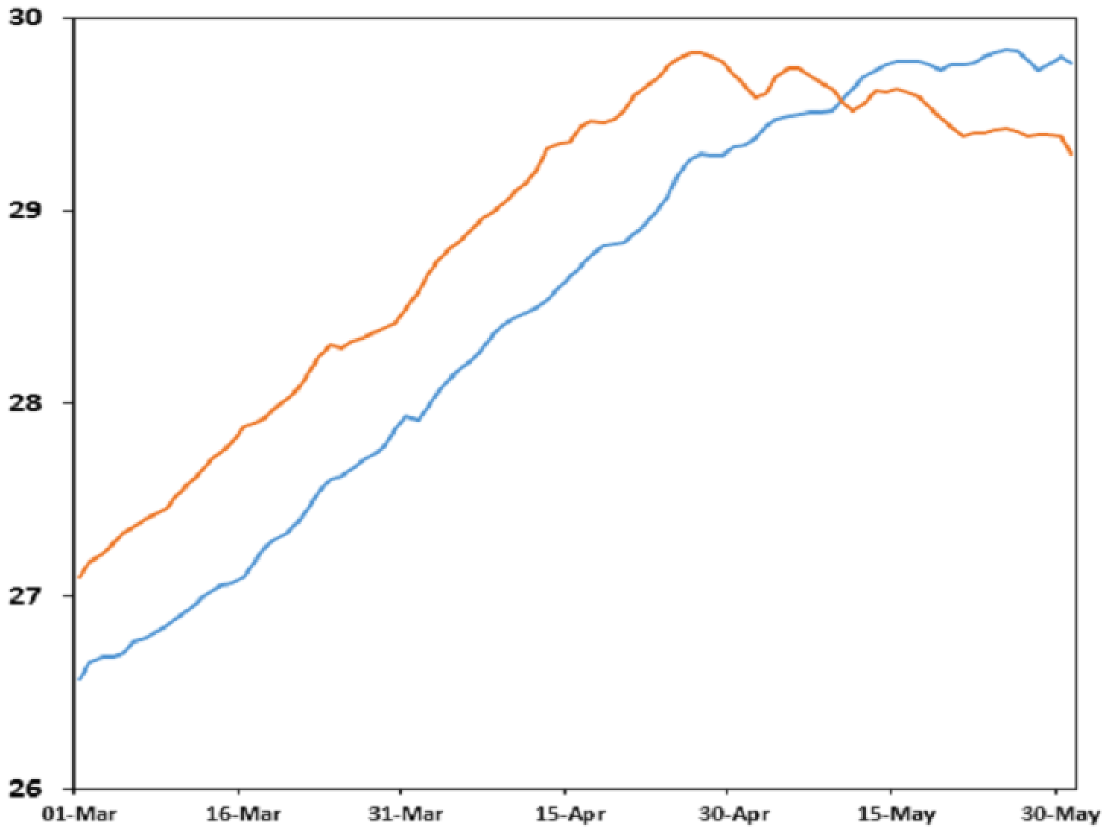
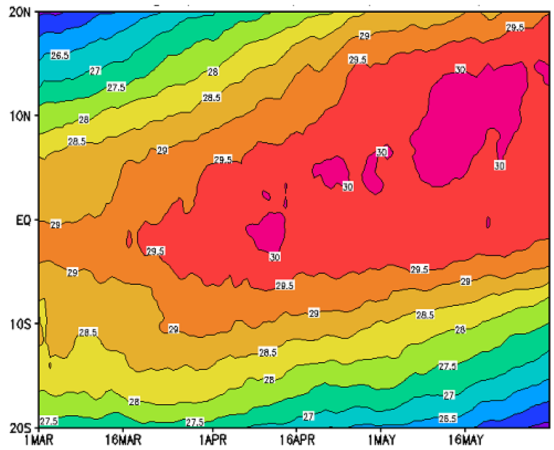


Figure 4.8: (a) Mean SST (^{0}C) before MOK in 60^0E- 70^0E, 10^0-20^0N (blue curve) and 85^0E- 95^0E, 10^0-20^0N (red curve) for the period 1991-2000

A Hovmoller diagram showing the time latitude change of the mean SST averaged over the longitudes 85^{0}E-95^{0}E over the BoB (Figure 4.8 b) and 60^{0}E-70^{0}E over the Arabian sea (Figure 4.8 c) is also shows the same.



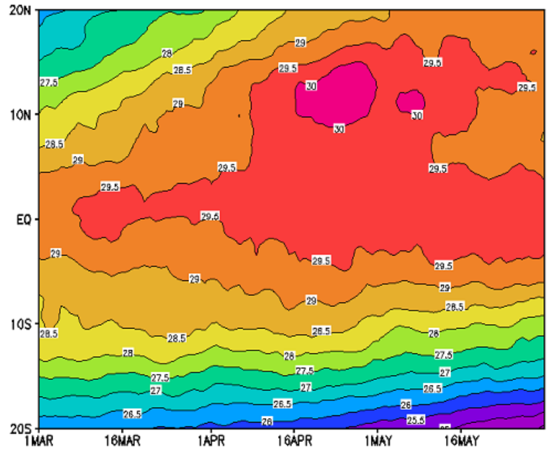


Figure 4.8: (b) Hovmoller of mean daily SST of 1991-2010 between longitudes 85^0 E- 95^0 E and, latitudes 20^0-20^0 N (BoB)

Figure 4.8: (c) Hovmoller of mean daily SST Of 1991-2010 between longitudes 60^0 E- 70^{0}E, and latitudes 20^0-20^0 N (Arabian Sea)

4.5 Decadal Variability in the DMOK

Figure 4.1(b) gives date of MOK from 1870 - 2014. It can be seen that in the beginning of last century, in many years of the decades 1900s and 1910s, MOK was delayed by one to two weeks, whereas in the middle of the last century, in many years of the decades 1950s and 1960s, MOK occurred early by one or two weeks. The mean date of MOK for the period 1900-1920 is 05 June with a standard deviation of 7.2 days. The respective values for the period 1950-1970 are 26 May and 8.5 days respectively. The difference in the mean dates of MOK for these two 21-year periods is statistically significant at the 99% level as per t-test. Thus on a decadal scale, there has been a statistically significant change in the date of MOK during last hundred and fifty years. There is, however, no linear trend for this long time series.

In order to study the influence of SST in the decadal time scale variability of DMOK five boxes (A to E) were selected over the oceans surrounding the equator as given in Table-4.5.

Table 4.5: Latitude and Longitude of box boundaries

Box	Longitude	Latitude
A	30^0-120^0E	5^0-20^0N
B	30^0-120^0E	0^0-20^0S
C	30^0-270^0E	5^0-20^0N
D	30^0-270^0E	0^0-20^0S
E	120^0-180^0E	0^0-15^0S

` The LCC between 11 year moving averages of SST gradient across the equator and MOK of pairs of these boxes is shown in Table-4.6. The boxes A (30-120^0E, 5-20^0N) and B (30-120^0E, 0-20^0S) represent areas north and south of the equator in the Indian Ocean. The LCC between 11 year moving average of SST gradient and MOK in these boxes during January and February (0.23) increased to 0.58 during March and April. The respective values for the boxes C (30-270^0E, 5-20^0N) and D (30-270^0E, 0-20^0S), representing north and south Tropical Indian Ocean and Pacific Ocean are 0.63 and 0.67 which are the maximum LCC. Figure 4.1(c) gives SST difference (gradient) for March and April every year during 1870-2014 between the boxes C (30-270^0E, 5-20^0N) and D (30-270^0E, 0-20^0S) with the eleven year moving average marked. The mean of SST gradient between these boxes for the period 1900-1920 is 1.04 with a SD of 0.13 in the season March and April. Respective values for the period 1950-1970 are 0.88 and 0.12. The difference in the mean SST for these two periods is statistically significant at the 99% level as per t-test. The LCC of 0.67 between the 11-year moving averages of March-April SST gradient and the date of MOK for the period 1870-2014 is large and statistically significant. Similarly, the LCC for the January-February SST gradient is 0.61 which is also large. Both the correlations are statistically significant at 99% level.

Table 4.6: LCC between the 11 year moving average of MOK and 11 year moving average of SST gradient for the season January-February and March-April. Significant values at 99.9 % are shown in red colour and that at 95 % significance is shown in green

SERIES	CORRELATIONS					
	B-A		D-C		E-A	
	JF	MA	JF	MA	JF	MA
NK - 1870 to 1980	0.04	0.52	0.51	0.53	-0.09	0.67
IMD (S) - 1901 to 2005	0.14	0.52	0.50	0.59	-0.07	0.63
IMD (S) - 1901 to 1970+ IMD (O) - 1971 to 2014	0.08	0.50	0.57	0.67	-0.11	0.56
NK+1 - 1870 to 1900 + IMD(S) -1901 to 1970 + IMD (O) - 1971 to 2014	0.23	0.58	0.63	0.67	0.02	0.63

In the introduction it was mentioned that the center of the Warm Pool of the tropical oceans gets shifted in the annual cycle from southwest Pacific to the north Indian Ocean during January to May (Joseph, 1990a, b, 2014) and the Warm Pool of May leads to the build-up of moisture in the atmosphere over north Indian Ocean and the adjoining Pacific Ocean making conditions favorable for MOK. The LCC between the 11-year moving averages of the SST gradient between the boxes E (south west Pacific Ocean) and A (north Indian Ocean) and DMOK is also studied, and the LCC for the months January-February is very low (0.02), but that for March-April is high (0.63) and are statistically significant at 99% level.

4.6 ITCZ transitions across the equator and DMOK

Joseph et al., (1994) had hypothesized that the time of transition of the ITCZ from south to north across the equator is related to DMOK, and the delayed crossing of the equator is associated with delay in DMOK. The ITCZ in monthly climatology has large north south movement over the Indian Ocean (the longitude zone 30^0E-120^0E), due to the monsoonal character of the area. In a large part of the longitude zone 120^0E- 270^0E the ITCZ lies only in the northern hemisphere both in January and July at low latitudes with only a few degrees latitude difference. The composite Hovmöller diagrams of the daily OLR averaged between longitudes 30^0E and 120^0E (Figure 4.9) and 120^0E and 270^0E (Figure 4.10) for the period 01 April -30 June of years of delayed and early monsoons, shows that the longitude belt 30^0E-120^0E (Indian Ocean), the minimum OLR zone has large south to north movement in association with MOK and smaller south to north motion in association with PMRP, both for delayed and early MOK. These features have considerable time difference, the crossing of the equator occurring much later in delayed monsoons as compared to early monsoons. For the longitude zone 120^{0}E-270^{0}E (Pacific Ocean) there is no equator crossing of the low OLR zone, but only in-situ deepening of convection in the ITCZ occurring earlier or later.

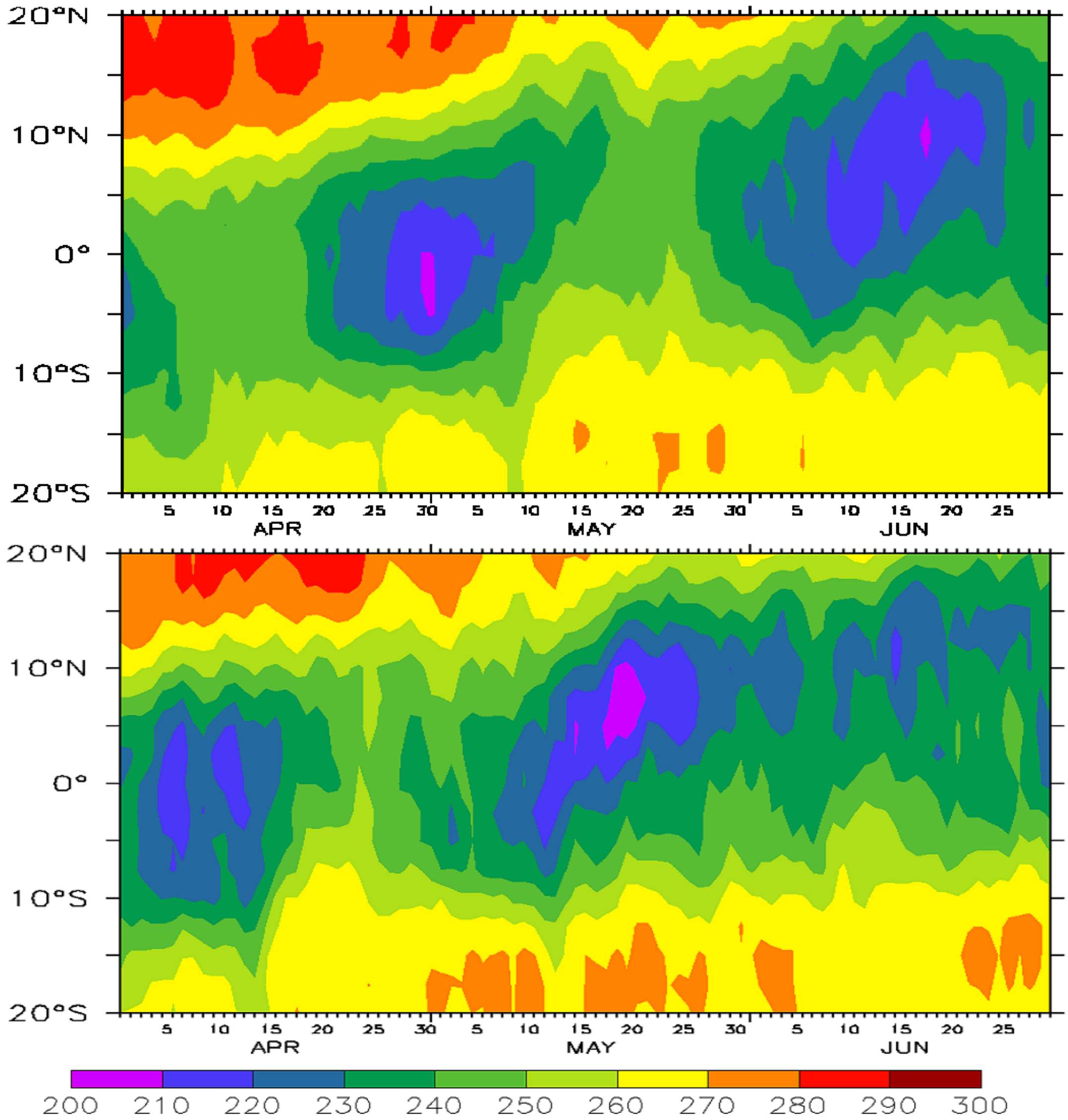


Figure 4.9: Hovmöller of the daily OLR (W m^{-2}) averaged between longitudes 30^{0}E and 120^{0}E (Indian Ocean) with delayed (upper panel) and early (lower panel) MOK

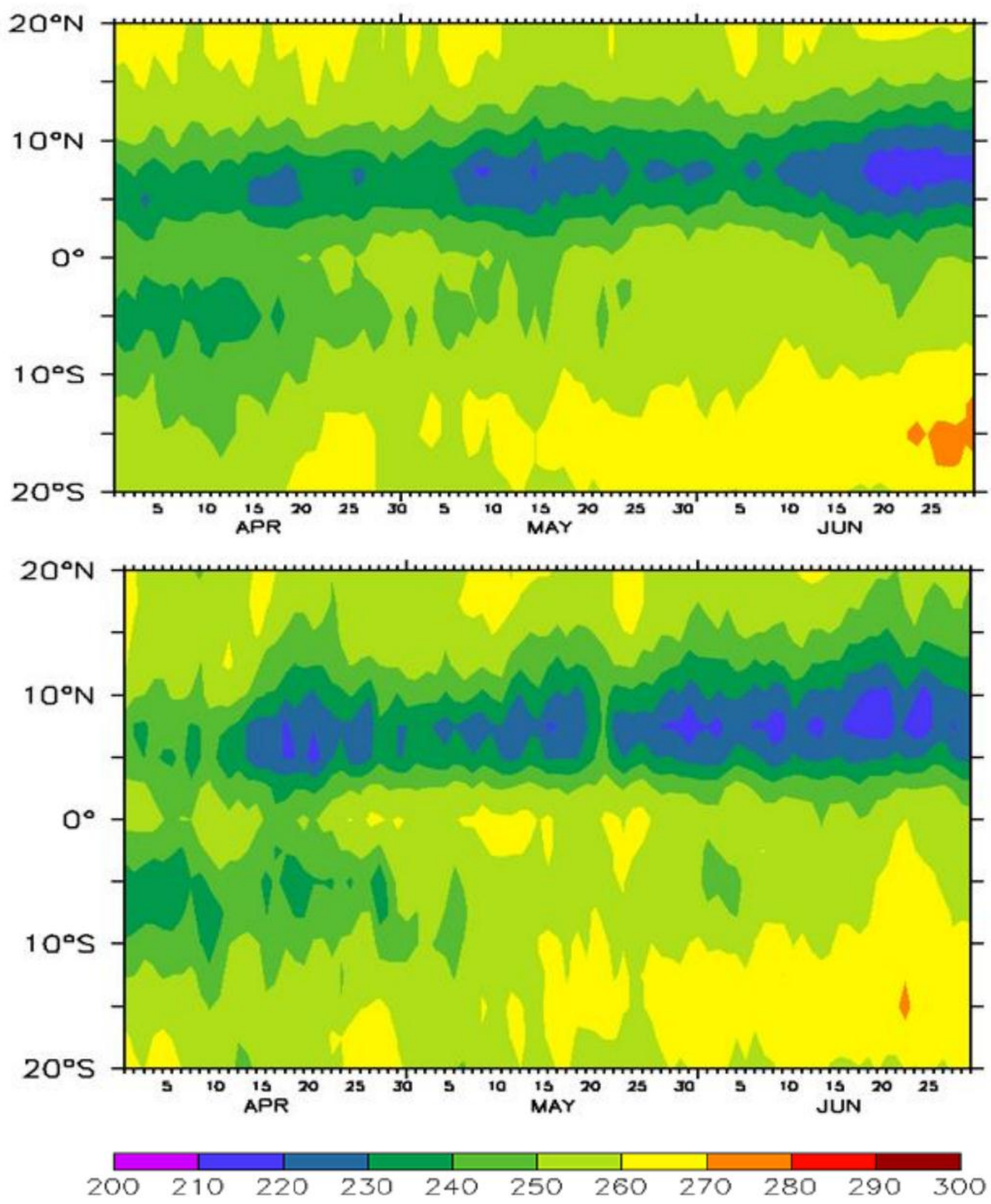


Figure 4.10: Hovmöllerof the daily OLR (W m^{-2}) averaged between longitudes 120^0E and 270^0E (Pacific Ocean) with delayed (upper panel) and early (lower panel) MOK

4.7 MOK , PMRP and Convection in the Indo-Pacific ocean basin

In section 4.4 pentad averages of OLR over a period of 50 days prior to MOK were studied for both delayed and early MOK cases. It was shown that during the pentads -5, -4 and -3 (in between MOK and PMRP), the Indian Ocean has very little convection (OLR high). It is during this period that convection shifts to the west Pacific Ocean (Joseph 1990a) and India comes under the downward limb of the Walker circulation. Flatau et al., (2001) have shown that this is the period when India gets increased occurrence of high summer temperature and high frequency of heat waves. Figure 4.11 gives the area averaged daily OLR for delayed MOK years and Figure 4.12 gives the same for early MOK years for the period 01 April to 30 June.

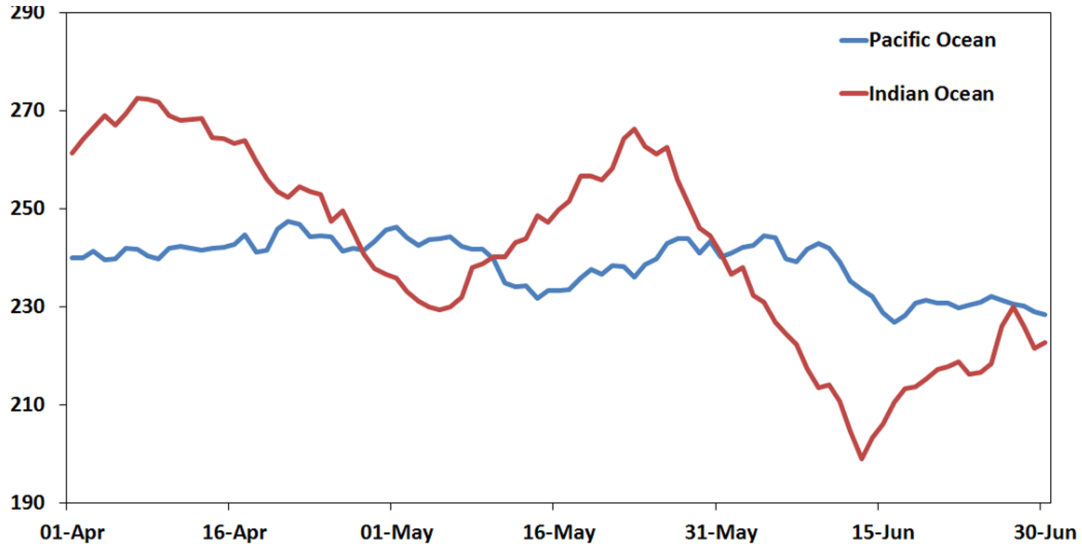


Figure 4.11: Area averaged daily OLR (W m^{-2}) of the Indian Ocean (blue curve, averaged over 5^0S-20^0N, 60^0E-90^0E) and the west Pacific Ocean (red curve, averaged over 5^0N-20^0N, 120^0E-180^0E) for composites of delayed monsoon years for the period 01 April to 30 June

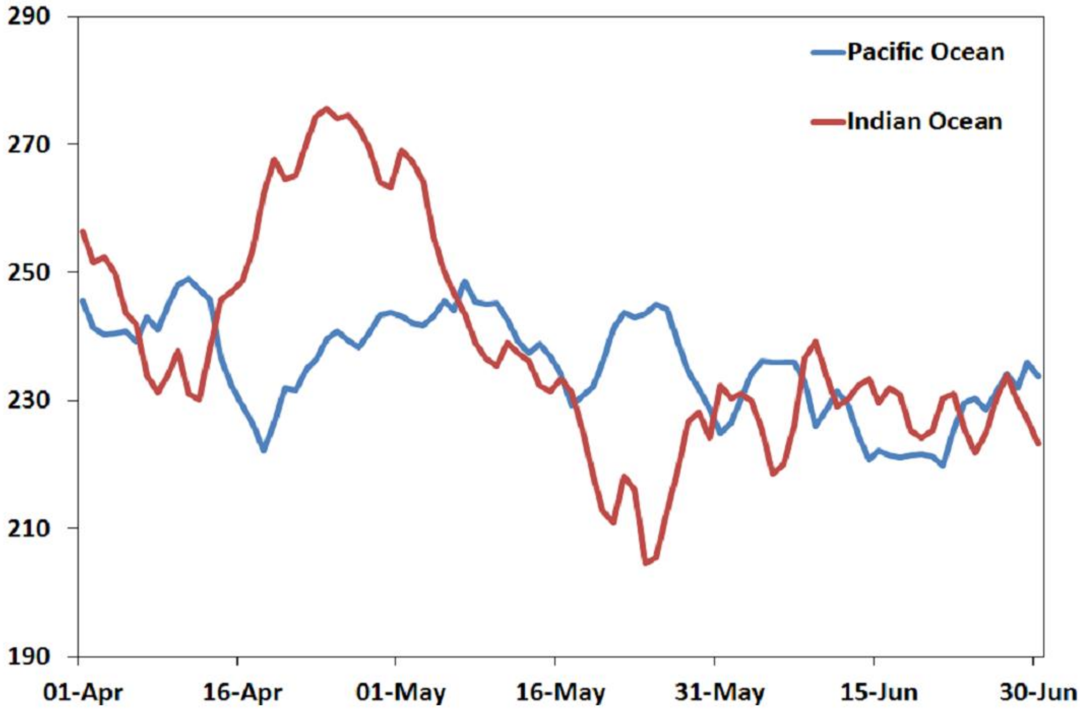


Figure 4.12: Area averaged daily OLR (W m^{-2}) of the Indian Ocean (blue curve, averaged over 5^0S-20^0N, 60^0E-90^0E) and the west Pacific Ocean (red curve, averaged over 5^0N-20^0N, 120^0E-180^0E) for composites of early monsoon years for the period 01 April to 30 June

PMRP and MOK are seen in the composites for the Indian Ocean (red curve). It also shows the high OLR period in between. The OLR for these boxes in the Indian and the Pacific oceans have a high linear correlation at a lag of about 10 days as can be seen from Figure 4.13.

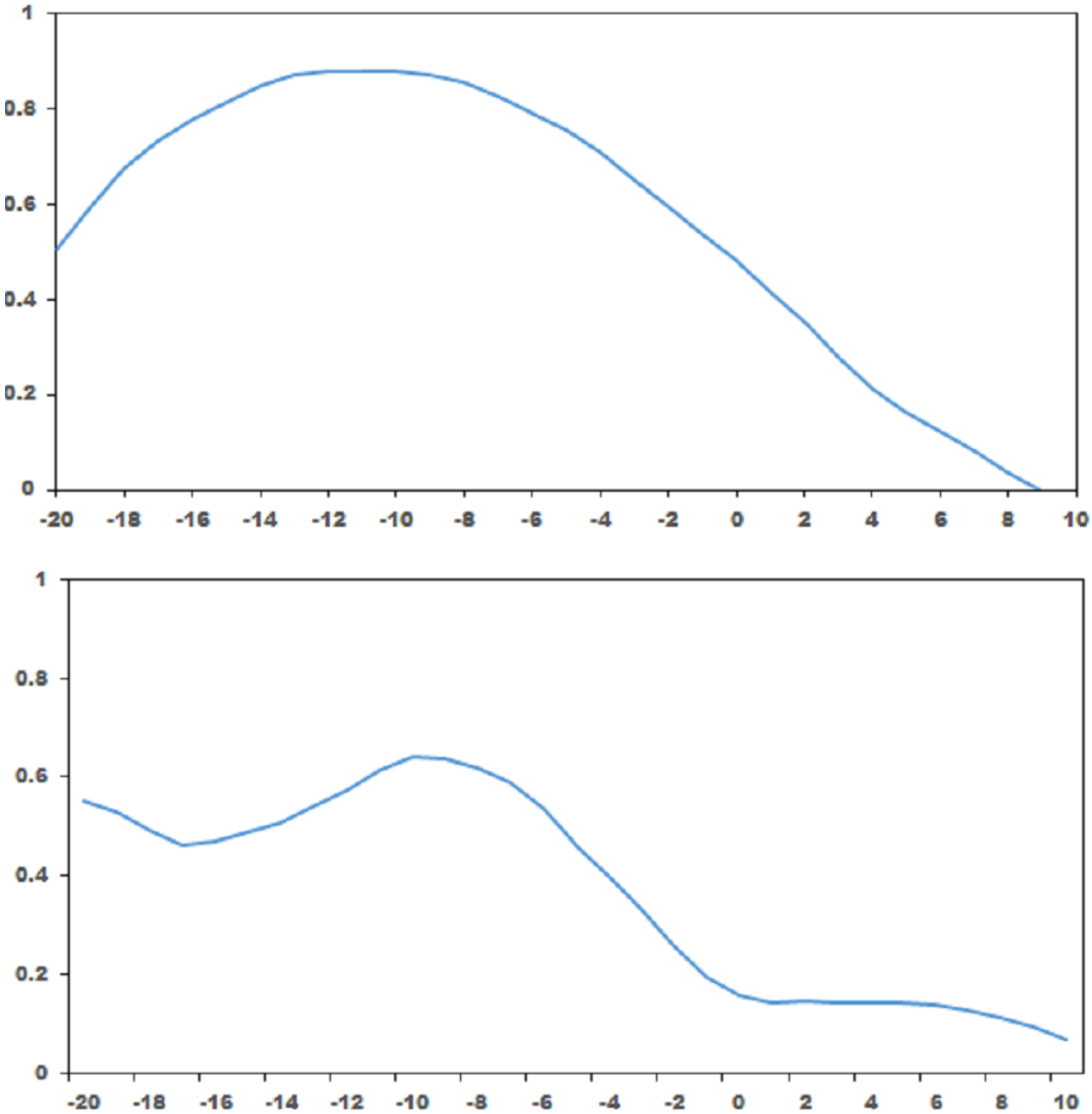


Figure 4.13: Lag correlation of area averaged daily OLR (W m^{-2}) of the Indian Ocean (upper panel) (averaged over 5^0S-20^0N, 60^0E-90^0E) and west Pacific Ocean (lower panel) (5^0N-20^0N, 120^0E-180^0E) for composites of delayed and early monsoon years for the period 01 April to 30 June

Chapter-5

Cyclogenesis over West Pacific Ocean in relation to PMRP and MOK

5.1 North West Pacific Cyclones (NWPC) and associated Tele-connections

Indo-Pacific Ocean is divided into several regions in which tropical cyclones have their genesis. All these regions are not equally productive, but the west Pacific Ocean is six times more prone to genesis of cyclones than the Arabian Sea (Ramage, 1959) and it generates 26 out of the 79 tropical cyclones produced by the global oceans in a year (Gray, 1979). On average about 60% of the west Pacific cyclones of a year occur during the four months (June - September), when India gets its summer monsoon rainfall. Many of these reach typhoon and super typhoon intensities. There are studies which examined the relationship between tropical cyclone activity over the west Pacific Ocean and the associated changes in the monsoon circulation over India (Iyer, 1931, 1935; Ramanna, 1969; Krishnamurthi et al., 1977; Saha et al., 1981; Chan, 1985, 2000; Chen et al., 1998; Joseph et al., 2002; Wang and Chan, 2002). Iyer (1935) studied the typhoons of the west Pacific Ocean and their associations with the Indian weather. Westward moving typhoons and their residual lows from China Sea cross the hilly regions of Indo-China and enter the Indian Ocean (BoB). Many of the cyclones that form in the BoB are the remnants of Pacific Typhoons. Pacific typhoons can provide nucleus for cyclones in the BoB (Ramanna, 1969). Raman (1955) associated 'break monsoon' with Pacific typhoons and he suggested that when typhoons in west Pacific move northwards, the axis of monsoon trough over India moves to foot hills of Himalayas, resulting in monsoon breaks. The relationship between the summer monsoon rainfall over India and the number of typhoon days over the northwest Pacific shows a negative correlation. The frequency of North West Pacific Cyclones (NWPC) tends to be enhanced during weak/break phases of the monsoon intra-seasonal variability; conversely the cyclone genesis is suppressed during periods of strong Indian summer monsoon (Rajeevan, 1993). Krishnamurti et al., (1977) examined the dynamics of westward moving disturbances from the western Pacific into the BoB. Saha et al., (1981) examined isallobaric maps of July and August and showed that majority of lows and depressions that formed over the BoB were associated with predecessor disturbances coming from the west Pacific Ocean. The variability of the Indian monsoon rainfall in relation to convective activity of the equatorial trough over the Indian and west Pacific oceans was examined by Joseph, (1990a). He noted that the cyclogenesis over the western Pacific was related to an out-of-phase variability in convection over the equatorial Indian Ocean and the west Pacific Ocean on the 30–50 day timescale. These convection episodes in turn are related to MOK.

In this chapter the tropical cyclone activity in the north-west Pacific Ocean is examined in relation to monsoon activity over the north Indian Ocean. It is known that both these areas oceans have a prominent 30-50 day signal in many atmospheric and oceanic parameters. PMRP and MOK are manifestations of the 30-60 day mode in rainfall associated with the ET (Equatorial Trough) over north Indian Ocean.

The cyclogenesis over the west and west-central Pacific Ocean north of the equator is found to be about 1.33 times higher during dry monsoon years compared to wet monsoon years (Kumar and Krishnan 2005). The frequency of formation of cyclones in the west/central Pacific Ocean increases during El Niño years (Chan, 1985) and the frequency of monsoon depression in the BoB decreases (increases) during El Niño (La Niña) years. Joseph et al., (1994) showed that, of the 22 years when the MOK was delayed for more than 8 days, 16 cases were associated with a moderate or strong El Niño. Further, it was found that, of the 13 strong El Niños during the same period, nine were associated with moderate to large delays in MOK. The association between convective systems over the northwest Pacific (NWP) and monsoon activity over the Indian subcontinent has been studied for the period 1951–2003 by Ramesh et al., (2009) and they found that no systems formed over the NWP region about 25 days prior to and after the MOK.

On the lines of previously found associations by Joseph (1990a) and Ramesh et al., (2009) this study seeks to re-examine the relationship between NWP tropical cyclones and the Indian summer monsoon using data of longer period now available.

5.2 Intra Seasonal Variability in NWPC

To study the variability of tropical cyclones in NWPC, three groups of years were chosen from period 1959-2013; the first group has years with Deficient Monsoon Rainfall (DMR), the second with Deficient Monsoon Rainfall associated with El Niño (DMRE) and the third with Excess Monsoon Rainfall (EMR).

The average monsoon rainfall (1 June to 30 September) of India for each year was used for this purpose. The standard deviation of ISMR is about 10% of the long term mean. Summer monsoon seasons with ISMR smaller than 90% of the mean are considered to be DMR/DMRE and those with ISMR larger than 110% of the mean are considered as EMR. (Tables 5.1, 5.2 and 5.3)

Data for tropical cyclones in the NWP were acquired from the UNISYS website (http://weather.unisys.com) and originate from the Joint Typhoon Warning Centre (JTWC) best track data set. The data set contains tropical cyclone characteristics such as name, life span, geographical location and track information. Since 1945, the tropical cyclones have been monitored by the United States Navy and from 1947 by the United States Air Force using aircraft data. Data from satellites have been used in the last few decades.

In this chapter, only systems within the 0°-50°N and 100-180 °E region were examined. In order to assess the role of synoptic systems associated with monsoon activity, the geographic location of the convective systems in NWP for the monsoon months (June–September) as well as their track and lifespan have been examined.

Table 5.1: Data in respect of DMR years

Years of dry monsoon rainfall (DMR)	Monsoon rainfall of India (in mm)	Date of monsoon onset over India (MOK)
1966	779.8	31-May
1968	791.6	8-Jun
1974	798.8	23-May
1979	724.8	12-Jun
1986	769.9	12-Jun

Table 5.2: Data in respect of DMRE years

Years of dry monsoon rainfall and El Niño (DMRE)	Monsoon rainfall of India (in mm)	Date of monsoon onset over India (MOK)
1965	738.3	26-May
1972	697.4	19-Jun
1982	767.4	30-May
1987	774.6	1-Jun
2002	737.3	9-Jun
2004	774.2	3-Jun
2009	698.2	23-May

Table 5.3: Data in respect of EMR years

Years of excess monsoon rainfall (EMR)	Monsoon rainfall of India (in mm)	Date of monsoon onset over India (MOK)
1959	1036.7	31-May
1961	1078.2	18-Jun
1970	998.7	26-May
1975	1011.4	2-Jun
1983	1001.5	12-Jun
1988	1094.1	2-Jun
1994	1001.2	28-May

Figure 5.1(a, b, c) gives the NWPC duration plotted against number of days counted from MOK which is taken as zero day for each year. Dashed line represents tropical storm stage and solid line represents the typhoon stage. The following broad inferences have been drawn from Figure 5.1.

(a) In EMR years, cyclogenesis does not occur during the first 25 days after MOK. In contrast in DMR and DMRE years only first few days are free of cyclogenesis.

(b) In all the three cases cyclogenisis once started continues till the end of 6 weeks.

(c) Typhoon days are many times more in DMRE and DMR compared to EMR.

Figure 5.2 gives similar data for a period of 90 days before MOK, which covers PMRP also. The following inference could be drawn from it.

(a) Cyclogenesis does not occur close to the dates of MOK.

(b) Cyclogenesis does not also occur around the days of PMRP.

(c) The 90-day period studied covers two 30-50 day cycles. There is hardly any cyclogenises in EMR years.

These results are in good agreement with those of Joseph (1990) and Ramesh et al., (2009).

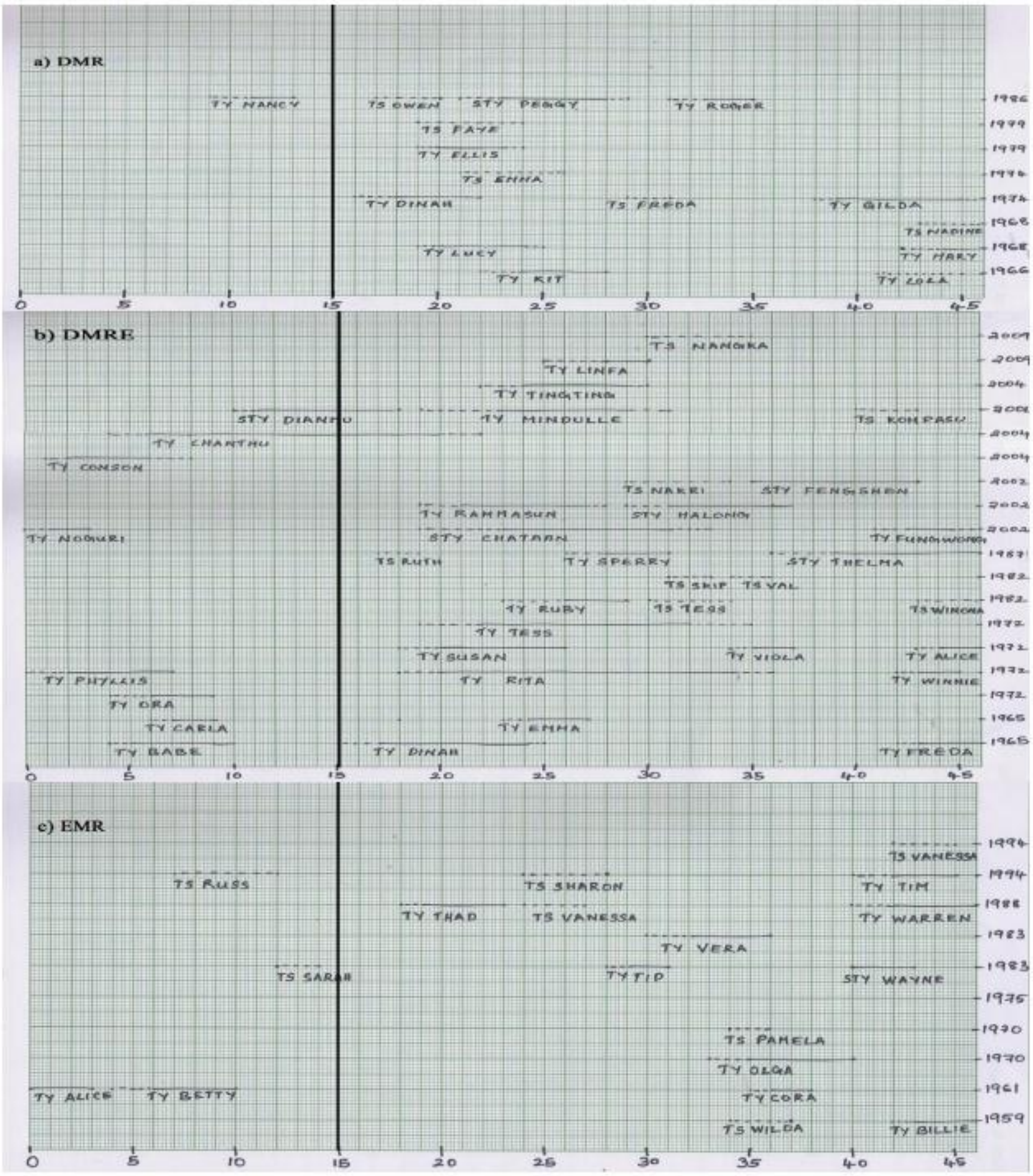


Figure 5.1: Time duration of NWPC with respect to date of MOK for (a) DMR, (b) DMRE and (c) EMR years during the period of 45 days after DMOK

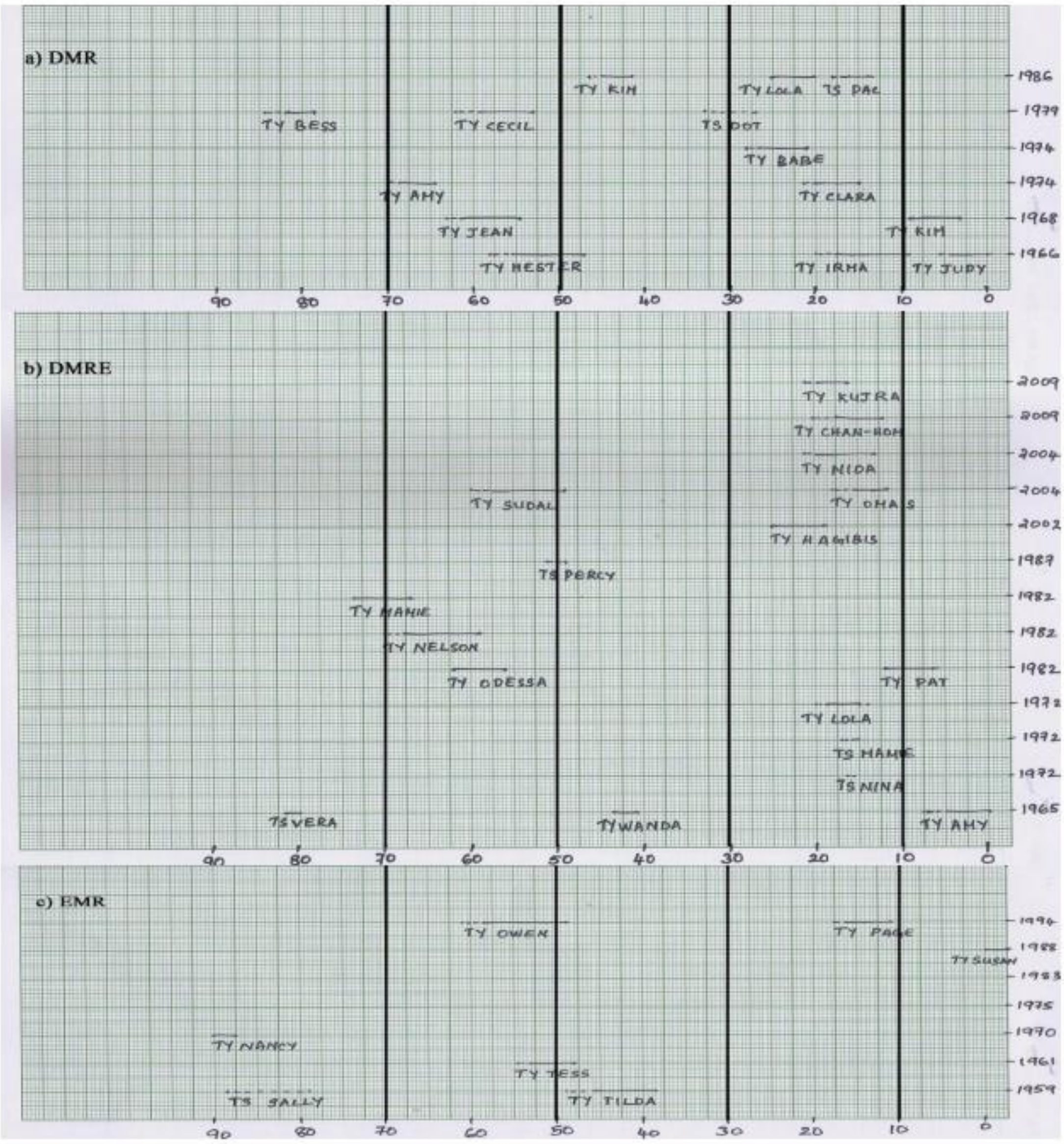


Figure 5.2: Time duration of NWPC with respect to date of MOK taken as zero day for 90 days before MOK for (a) DMR, (b) DMRE and (c) EMR years

Chapter-6

El Niño, IOD and ISMR Inter-relationships

6.1 El Niño, IOD and ISMR

There are three climatically important phenomena in the ocean-atmosphere system of the global tropics. One is the Asian summer monsoon of which the Indian summer monsoon is a major component occurring during June to September every year. The second is the El Niño (and La Niña) phenomenon lasting two to three years that occurs with an irregular periodicity of 3 to 7 years. The third is the IOD which has positive and negative phases each lasting less than a year, occurring once in a few years. These three phenomena are found to interact with each other on inter-annual and decadal time scales. This chapter is devoted to the study of the linkages that exist among ISMR (June to September rainfall of India), N3.4index (October-December) representing El Niño/La Niña in their mature phase and DMI (September- November) representing the mature phase of IOD, using a linear correlation method.

Figure 6.1 shows the linear correlation coefficient (LCC) between pairs of these features using data of the period 1950 to 2010. While the LCC between ISMR and IOD (DMI of SON) is small and statistically not significant, the LCC between ISMR and El Niño/La Niña (N3.4 of OND) and that between N3.4 of OND and DMI of SON are high and statistically significant at the 99.9% level.

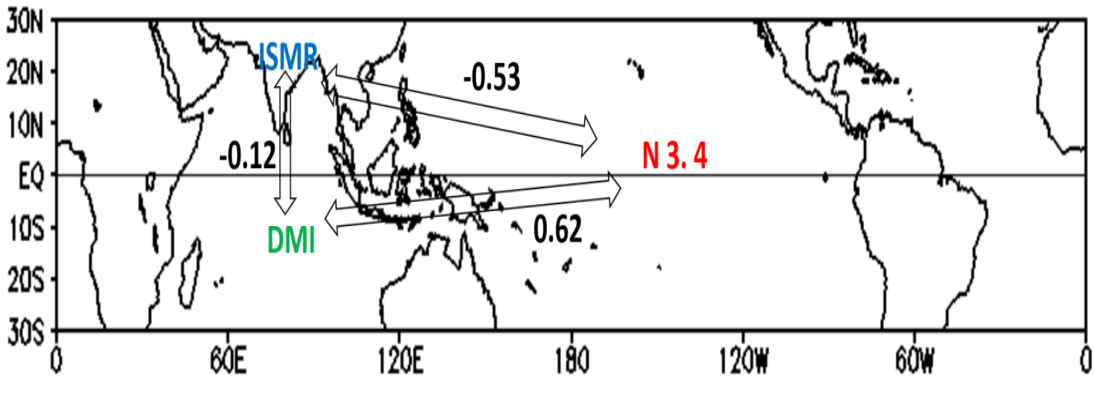


Figure 6.1: The locations of the three climatological phenomena ISMR, El Niño/La Niña (N3.4) and IOD (DMI) are shown along with the LCC among pairs of these using data of the period 1950-2010

6.2 El Niño/La Niña

Walker (1924) identified Southern Oscillation as a phenomenon encompassing the Indian and Pacific oceans. It later became known as the El Niño Southern Oscillation (ENSO). During the Tropical Ocean Global Atmosphere (TOGA) programme (1985-1994), the scientific community studied ENSO in great detail with particular focus on the Pacific Ocean. The basic physics of the evolution of El Niño/La Niña events is now reasonably well understood. The phenomenon called Bjerknes Feedback (positive feedback between warm SSTA in the central/eastern Pacific and westerly wind anomalies in the western Pacific) provides the necessary instability for an El Niño to develop. But favourable conditions are needed for the Bjerknes Feedback to result in an El Niño. The role of the tropical Pacific heat content (Wyrtki, 1975) is considered important in this connection. The Warm Water Volume (WWV) of the equatorial Pacific, defined as the volume of water above 20^0C within 5^0S to 5^0N by Meinen and McPhaden, (2000), is a critical parameter in the evolution of ENSO. An anomalously high WWV favours warming of the SST in the central Pacific (Wyrtki, 1985; Jin, 1997a, b), which is further enhanced by air-sea interactions to culminate into a fully developed El Niño by the end of the year. Inter-annual variations in WWV are largely driven by equatorial zonal wind stress (Jin , 1997a, b; Goddard and Philander, 2000; Clarke, 2008; Brown and Fedorov, 2010; Fedrov, 2010; Lengaigne et al., 2012). The westerly wind anomalies associated with an El Niño event then result in poleward heat transport that ‘‘discharges’’ the equatorial Pacific WWV, thus creating favourable grounds for a La Niña event to develop. A La Niña would in turn favour a WWV recharge and thus to the transition to an El Niño event.

ENSO is found to influence the global climate. The heat sources and sinks associated with the displacements of atmospheric deep convection in the tropical Pacific force atmospheric planetary waves, which are associated with global teleconnections (Trenberth et al., 1998). Zonal displacements and variations of the Walker circulation associated with El Niños also strongly affect the Indian Ocean. Increased subsidence and reduced wind speed over the Indian Ocean result in a basin scale warming (e.g. Klein et al., 1999; Ohba and Ueda, 2005, 2007, 2009a,b) which persists beyond the end of the El Niño (e.g. Xie et al., 2009). The effect of ENSO on the Indian summer monsoon has also been noted (e.g. Walker, 1924; Rasmusson and

Carpenter, 1983, Lau and Li, 1984; Gershunov et al., 2001; Ashok et al., 2004; Fasullo 2004; Xavier et al., 2007; Jing and Li 2011; Boschat et al., 2012; Cherchi and Navarra, 2012). The overall view that emerged after the TOGA programme was that ENSO was an intrinsic mode of Pacific variability to which the Indian Ocean responds rather passively.

6.3 Indian Ocean Dipole (IOD)

The view that Indian Ocean had only a passive role changed at the turn of last century, with the discovery of an intrinsic mode of climate variability in the Indian Ocean known as the IOD. As in the case of ENSO, a positive air-sea feedback was found to foster the development of IOD events. The positive phase of IOD is characterized by cold (warm) anomalies in the eastern (western) Indian Ocean and equatorial easterly wind anomalies in the central Indian Ocean that mutually enhance each other (Reverdin et al., 1986; Saji et al., 1999; Webster et al., 1999; Murtugudde and Busalacchi, 1999; Murtugudde et al., 2000). Like ENSO, the IOD was found phase locked to the seasonal cycle. It develops during boreal summer season, culminates in boreal fall, and decays by the end of the year. There is now ample evidence that IOD is an intrinsic mode of variability of the Indian Ocean (e.g. Annamalai et al., 2003; Fischer et al., 2005; Behera et al., 2006; Luo et al., 2008, 2010). There is, however, a tendency for the IOD events to co-occur with ENSO events (Reverdin et al., 1986; Murtugudde et al., 2000; Annamalai et al., 2003; Yamagata et al., 2004; Fischer et al., 2005; Xie et al., 2009). El Niño events indeed favour easterlies over the equatorial Indian Ocean which can trigger a positive IOD (Annamalai et al., 2003). While the Indian Ocean is no longer viewed as a climatically inactive ocean (e.g. Annamalai and Murtugudde, 2004), this still illustrates an influence of the Pacific on the Indian Ocean.

Many recent studies suggested that Indian Ocean SSTs may influence the tropical Pacific (Watanabe, 2008a, b; Jansen et al., 2009). Annamalai et al., (2005, 2010) showed that SSTA over the Indian Ocean influence the atmospheric circulation over the Pacific. Luo et al., (2010), Annamalai et al., (2010) further suggested that El Niños tend to be stronger when they co-occur with IOD events (also Behera and Yamagata, 2003).The possible role of the Indian Ocean basin-wide warming/cooling onto El Niño/La Niña asymmetries are discussed by Ohba and Ueda, 2009a; Ohba et al., 2010; Okumura and Deser, 2010; Okumura et al., 2011 and Ohba and Watanabe, 2012. Terray and Dominiak, (2005) also noted that cold SSTA in the southeastern Tropical Indian Ocean in February–March serve as an efficient predictor of the El Niño peak 10 months later. The influence of the Indian Ocean on the Pacific was also suggested to be through the monsoon, and was built into TBO (Meehl, 1987; Meehl, 1997, Meehl et al., 2003), a framework to explain the biennial tendency in the system formed by the Indian and Australian monsoons, Indian Ocean SST, and ENSO.

Izumo et al., (2010) have suggested that IOD events could be precursors of ENSO. Based on satellite data, they showed that positive (negative) IODs tended to be followed by La Niña (El Niño) events approximately 14 months later. They demonstrated significant skill in predicting ENSO events 14 months in advance by jointly using IOD and WWV as predictors in a simple linear regression model. They suggested the following mechanism to explain this delayed relationship. The warm SST anomaly in the eastern Indian Ocean associated with a negative IOD induces an easterly wind anomaly in the equatorial Pacific. This negative IOD anomaly and the associated wind signal over the Pacific abruptly disappear in November–December at the termination of the IOD event. The equatorial wave response to this wind anomaly and its sudden relaxation drives an eastward current anomaly in the western-central equatorial Pacific. The latter favors the onset of an El Niño by pushing the Warm Pool eastward. This IOD influence interacts with the intrinsic Pacific WWV preconditioning. For example, an IOD can interact constructively with positive WWV conditions to favor the onset of an El Niño. The scenario presented above by Izumo et al., (2010) describes the influence of an external forcing (the tropical Indian Ocean) on the Pacific Ocean. The analysis of Izumo et al., (2010) was based on only 30 years of data. A later study (Izumo et al., 2014) explored the robustness of the delayed IOD–ENSO relationship over the period 1872-2008.

As described in section 6.3 the DMI index is used to represent the strength and sign of the IOD each month. Figure 6.2 gives the mean September to November DMI for each year of the period 1871-2010. IOD reaches peak strength during the September-November season.

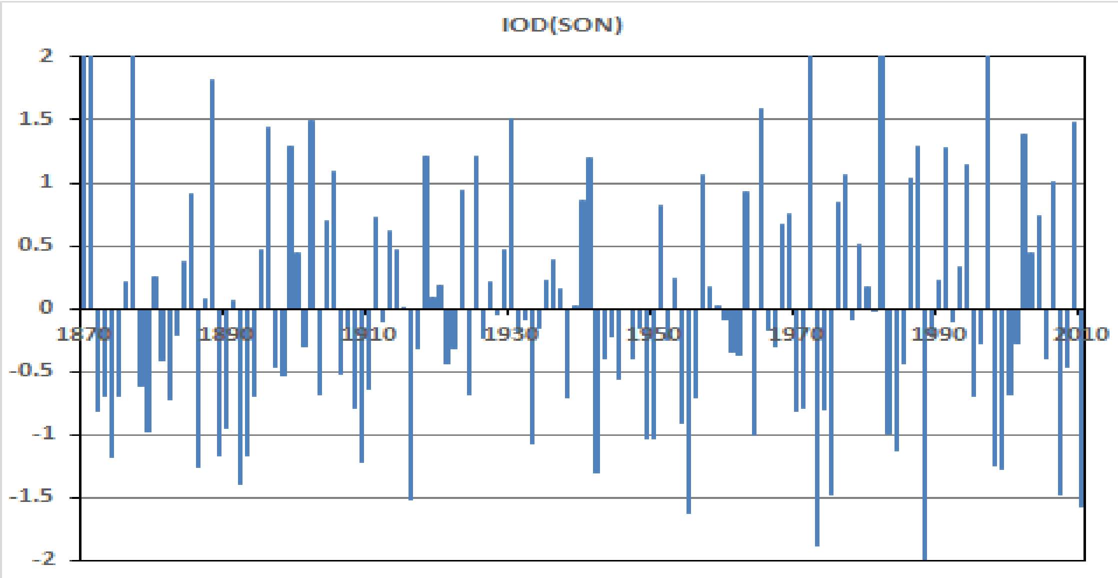


Figure 6.2: Mean September to November DMI which represents the strength and sign of IOD for each year of the period 1871-2010

6.4 Indian Summer monsoon Rainfall (ISMR)

There are many studies investigating the tele-connection between the two phenomena, El Niño and monsoon. The empirical relationship between ISMR and ENSO on the inter-annual timescale has been studied extensively (Sikka, 1980; Pant and Parthasarathy, 1981; Rasmusson and Carpenter, 1983; Shukla and Paolino, 1983; Parthasarathy and Pant, 1985; Shukla, 1987; Webster, 1987; Wang and LinHo, 2002; Wu et al., 2013). The physical mechanism through which ENSO is related to the monsoon (e.g., the SST variations in the eastern Pacific being negatively correlated with the ISMR) has been addressed in several studies (Webster, 1987a, b; Webster and Yang, 1992; Nigam, 1994; Goswami, 1998).

The inter-annual variation of the Indian monsoon is characterized by fluctuations of a regional Hadley circulation. A strong monsoon is associated with anomalous ascent around 25^0N and a weak monsoon is associated with anomalous ascent near the equator (Goswami et al., 1999). ENSO influences the Indian monsoon not by direct subsidence over the Indian continental region but through an interaction between the equatorial Walker circulation and the regional monsoon Hadley circulation (Goswami and Xavier, 2005). The warm episodes of ENSO are associated with a shift in the climatological Walker circulation to the eastern Pacific. This shift results in enhanced low level convergence over the equatorial Indian Ocean and in driving an anomalous Hadley circulation with descent over the Indian continent and decreased monsoon rainfall (Goswami, 1998).

Available rainfall measurements have shown a prominent decadal scale oscillation in ISMR as described in section 3.6 and shown in Figure-3.6. During the three-decade long dry epochs of 1901–30 and 1961– 90, India experienced drought monsoons on average once every three years. In contrast, during the three decade long wet epochs of 1871–1900 and 1931–60, the average frequency of droughts was approximately one in every 10–15 years.

Thus, during the 120 years of 1871 to 1990, alternating 30 year long dry and wet epochs have occurred regularly. An almost parallel decadal oscillation is seen in El Niño/La Niña (Gu and Philander, 1995) who found that the amplitude of ENSO is large from 1885 to 1915, to be small during the period 1915 to 1950 and to increase rapidly after about 1960. Figure 6.3 gives a plot of mean N3.4 of October to December season for each year of 1871-2010 which shows frequent El Niños in dry epochs of ISMR and frequent La Niñas in wet epochs.

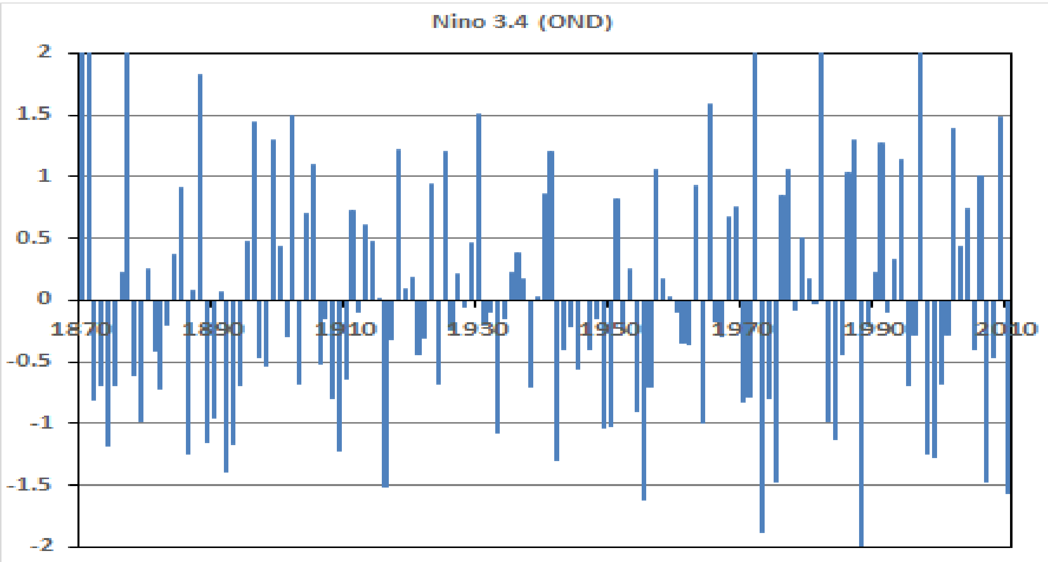


Figure 6.3: Average N3.4 anomalies of the October-December season of each year of the period 1871-2010 are shown. Positive anomaly of 0.5 and more may be taken as an El Niño year and negative anomaly -0.5 and less as a La Niña year

6.5 Relation between ISMR and N3.4

Rasmusson and Carpenter, (1982) in their classical study on the SST anomaly associated with composite of several El Niños have shown that the main growth of the warm SSTA of El Niño in the equatorial east Pacific Ocean is during the monsoon season. Their study shows that in March to May there is only a small area with positive SSTA close to the South American coast in the El Niño composite. By August to October, the positive SSTA region has grown to cover the equatorial area from longitude 160E to the South American coast and the magnitude of the SSTA there is very large. After October further increase of the area and magnitude of the SSTA is small. Thus the main growth in area and intensity of the SSTA in an El Niño has been during the monsoon season June to September. It is well known that dry and wet monsoons are associated with El Niño and La Niña, respectively. A question therefore arises whether ISMR has a role in the generation of El Niño/La Niña?

Table 3.1 gives the years of the 7 very dry monsoons and the 9 wet monsoons of the period 1950-2010 and also of the 7 strong El Niños and 10 strong La Niñas of the same period. Composites of the monthly N3.4 of January to December are given in Figure 6.4 for the very dry and wet monsoons. It is seen that the main increase in N3.4 in the composite very dry monsoon is from May to October and the main decrease in N3.4 in the composite wet monsoon is also from May to October. Thus the main period of growth of N3.4 is during the four monsoon months June to September culminating in the mature phase of E l Niño/La Niña by the year end.

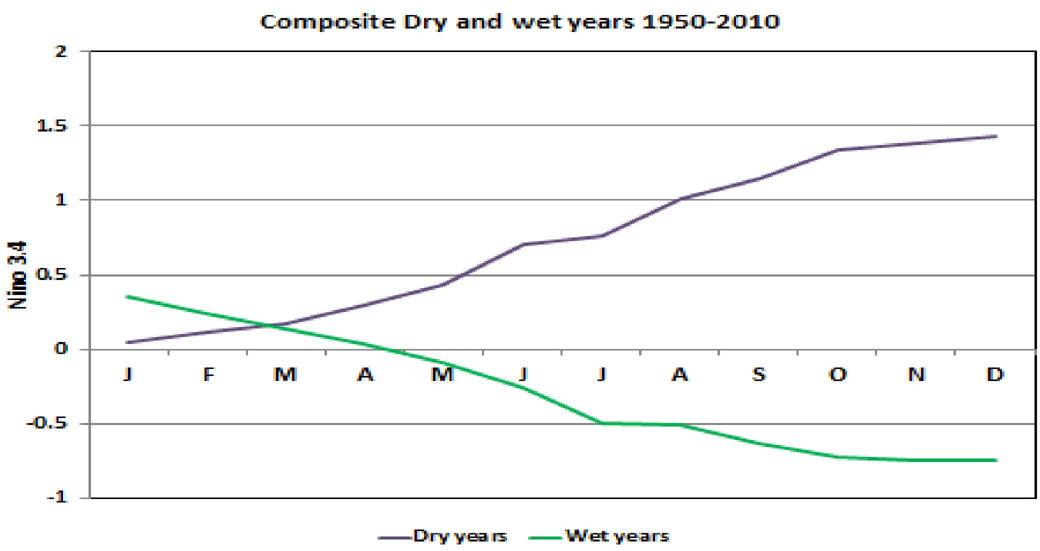


Figure 6.4: Growth of the positive and negative N3.4 index in the composites of the very dry and wet monsoons of Table 3.1

Similar composites for the strong El Niños and strong La Niñas of the period 1950-2010 (Table 3.1) are given in Figure 6.5 which also shows that the main growth of SSTA in the N3.4 region associated with El Niño and La Niña is during the monsoon months June to September.

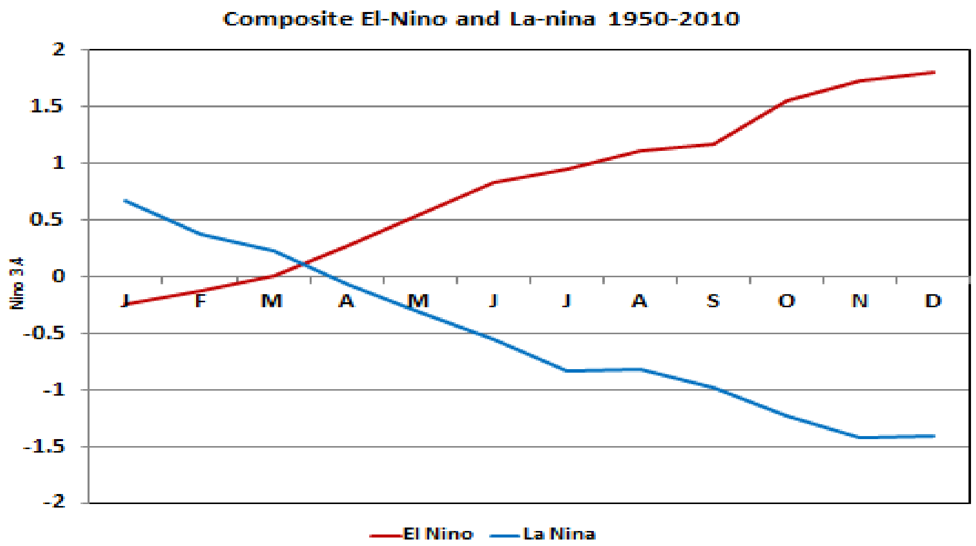


Figure 6.5: Growth of the positive and negative N3.4 index in the composites of the strong El Niños and strong La Niñas of Table-3.1

Using the data of the 51 year period 1960 to 2010 a scatter diagram is drawn (Figure 6.6) between ISMR and N3.4 (October minus May) which is the growth of the monthly N3.4 index across the monsoon season June to September. The linear regression line relating the two is marked in the Figure 6.6. The LCC between ISMR and N3.4 (October minus May) is -0.49 which is negative and statistically significant at 95% level. The two ovals marked (Figure 6.6) shows the outliers whose years are marked in two digits against the dots (example 97 for 1997 and 07 for 2007). Almost all the years in the top oval are strong El Niño years based on the N3.4 index of Oct-Dec (last column of Table 6.1). The prominent exception is the year 2003. Almost all the years in the bottom oval are strong La Niña years. The prominent exceptions are the two years 1992 and 1995 as may be seen from the last column of Table 6.2.

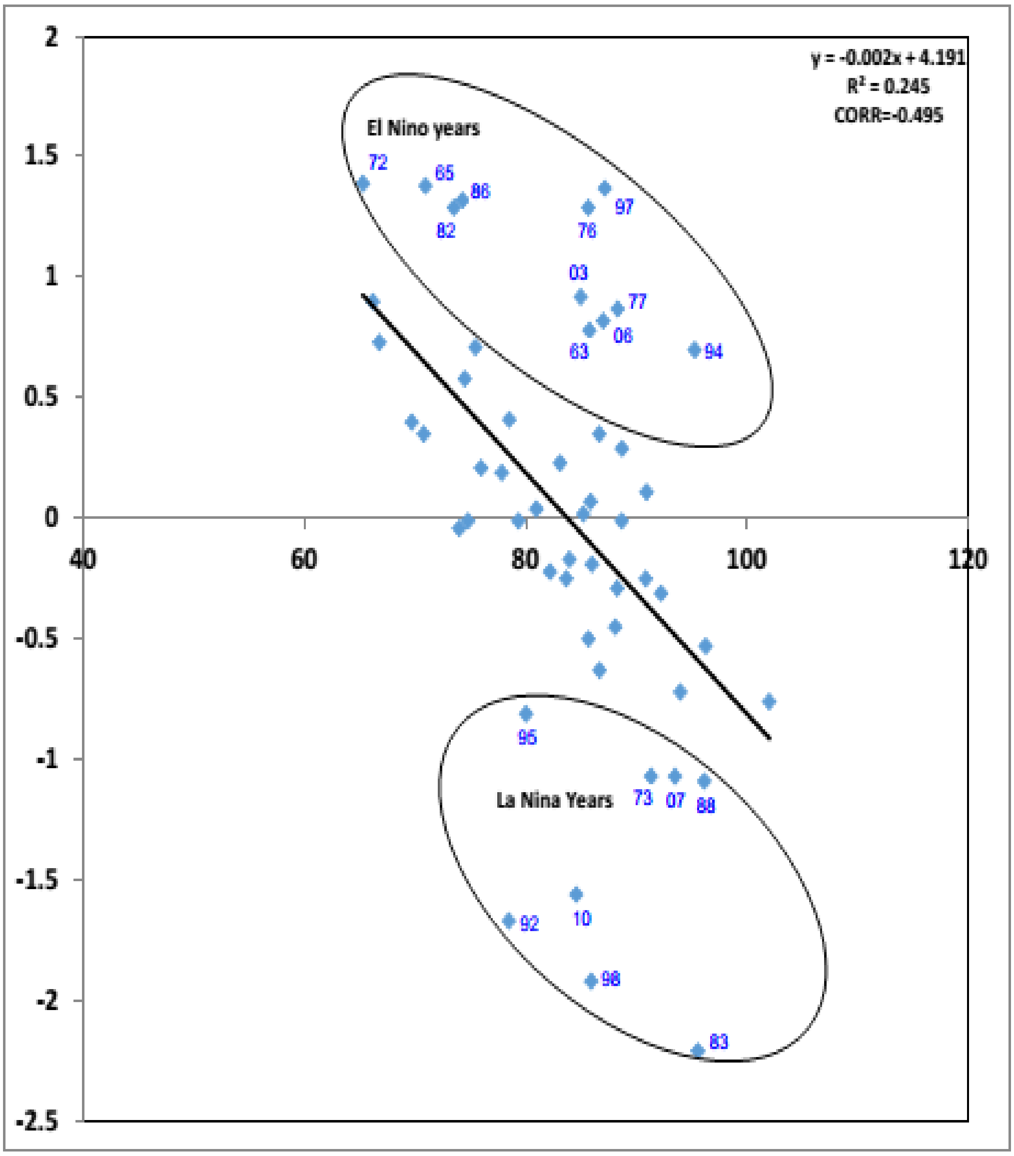


Figure 6.6: Scatter diagram relating ISMR and October minus May N3.4 (growth of N3.4 index across the monsoon season June to September), The outliers are in the two ovals with the years in 2 digits marked against the dots

Table 6.1: ISMR, N3.4 (Oct minus May) and N3.4 (Oct-Dec) of the years in the top oval of the Figure 6.6

El Niño years	ISMR	N3.4 ΔB (OCT-MAY)	N 3.4 (OND)
1963	85.77	0.78	0.930
1965	70.92	1.38	1.593
1972	65.28	1.39	2.003
1976	85.66	1.29	0.850
1977	88.3	0.87	1.060
1982	73.51	1.29	2.187
1986	74.29	1.32	1.030
1994	95.27	0.7	1.137
1997	87.14	1.37	2.397
2003	84.95	0.92	0.440
2006	86.99	0.82	1.003

Table 6.2: ISMR, N3.4 (Oct minus May) and N3.4 (Oct-Dec) of the years in the bottom oval of the Figure 6.6

La Niña years	ISMR	N3.4 ΔB (OCT-MAY)	N 3.4 (OND)
1973	91.32	-1.07	-1.887
1983	95.56	-2.21	-0.983
1988	96.14	-1.09	-2.027
1992	78.48	-1.67	-0.093
1995	80.02	-0.81	-0.690
1998	85.93	-1.92	-1.250
2007	93.5	-1.07	-1.477
2010	84.56	-1.56	-1.567

Table 6.3: ISMR, N3.4 (Oct minus May) and N3.4 (Oct-Dec) of the 32 residual years of Figure 6.6

Residual years	ISMR	N3.4 ΔB (OCT-MAY)	N 3.4 (OND)
1960	83.95	-0.17	-0.090
1961	102.01	-0.76	-0.343
1962	80.96	0.04	-0.363
1964	92.24	-0.31	-1.000
1966	73.99	-0.04	-0.173
1967	85.99	-0.19	-0.300
1968	75.45	0.71	0.673
1969	83.1	0.23	0.753
1970	93.97	-0.72	-0.820
1971	88.67	-0.01	-0.787
1974	74.79	-0.01	-0.800
1975	96.25	-0.53	-1.470
1978	90.92	0.11	-0.083
1979	70.77	0.35	0.507
1980	88.26	-0.29	0.170
1981	85.21	0.02	-0.023
1984	83.65	-0.25	-1.130
1985	75.97	0.21	-0.433
1987	69.7	0.4	1.290
1989	86.65	0.35	-0.220
1990	90.84	-0.25	0.227
1991	78.52	0.41	1.273
1993	86.66	-0.63	0.330
1996	85.86	0.07	-0.280
1999	82.19	-0.22	-1.277
2000	77.82	0.19	-0.683
2001	79.34	-0.01	-0.277
2002	66.19	0.9	1.383
2004	74.47	0.58	0.740
2005	85.68	-0.5	-0.393
2008	88.7	0.29	-0.470
2009	66.76	0.73	1.477

The data regarding the 32 residual years (outside the ovals) are given in Table 6.3. It is seen that in 24 out of the 32 years N3.4 of the Oct-Dec season is within -1 to +1. Only in 4 cases N3.4 (Oct-Dec) is more than 1.0 (strong El Niños). They are the years 1987, 1991, 2002 and 2009. Only in 4 cases N3.4 is less than -1.0 (strong La Niñas). They are the years 1964, 1975, 1984 and 1999.For the 32 residual years the LCC between ISMR and N3.4 (October minus May) is -0.78 which is very high and statistically significant at the 99.9% level. Figure-6.7 gives a scatter diagram between ISMR and N3.4 (Oct minus May) of the 32 residual years. The high LCC between ISMR and N3.4 (Oct minus May) shows the important role of ISMR in relation to the development of the SSTA in El Niño/La Niña. Further studies are needed to know the mechanisms involved. In some of the years of negative anomalies in ISMR the growth of N3.4 is large during the monsoon season and that happens to be a strong El Niños (the years in top oval of Fig-6.6). Further research is needed to understand what other atmospheric/oceanic processes are involved for the same. Will it be Madden Julian Oscillations of large amplitude or long active-break cycles of the monsoon in addition to the negative anomalies in ISMR? The case of strong La Niñas in the bottom oval of Figure 6.6 also needs research to find out the additional factors involved.

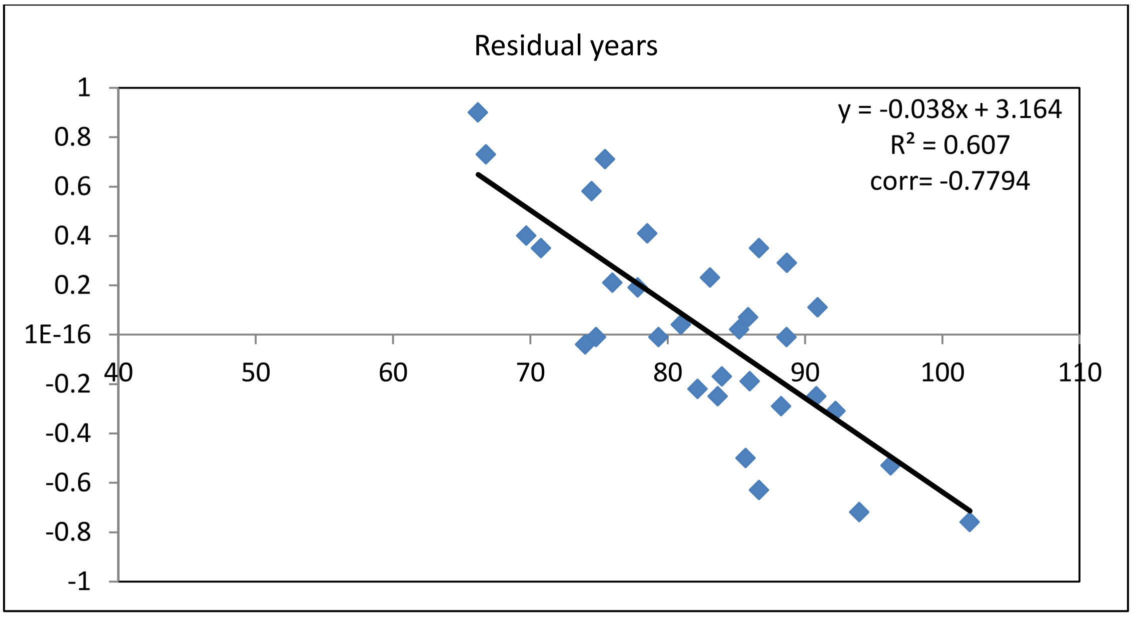


Figure 6.7: Scatter diagram between ISMR along x-axis and N3.4 (Oct minus May) along y-axis residual years of the 32. The LCC between these two factors is -0.78 which is very high and statistically significant. The linear regression line between these two is marked

6.6 Relation between IOD (DMI) and N3.4/ISMR

DMI (Sep-Nov) and N3.4 (Oct-Dec) were shown to have a high and statistically significant positive LCC of 0.62 for the period 1950 to 2010. For the same period the LCC between ISMR and N3.4 (Oct-Dec) was also high and statistically significant but negative at -0.53 epochs of ISMR of the long period 1871 to 1990 as described in Section 6.4. Table 6.4 and 6.5 give these LCCs. (see Section 6.1 and Figure 6.1). Do they have high LCCs also in the dry and wet It is found that the LCCs are strong both in the dry and wet epochs of the monsoon.

Table 6.4: LCC of DMI (SON) with N3.4 (OND) of the same year in dry and wet epochs of monsoon 1871-1990

LCC of DMI (OND) with N3.4 (OND) of the same year	
Epoch	**LCC**
1871-1900	0.681
1901-1930	0.732
1931-1960	0.474
1961-1990	0.636

Table 6.5: LCC of N3.4 (OND) with ISMR (JJAS) of the same year in dry and wet epochs of monsoon 1871-1990

LCC of N3.4 (OND) with ISMR (JJAS) of the same year	
Epoch	**LCC**
1871-1900	-0.730
1901-1930	-0.612
1931-1960	-0.659
1961-1990	-0.659

Table 6.6: LCC of DMI (SON) with N3.4 (OND) of the following year in dry and wet epochs of monsoon 1871-1990

LCC of DMI (OND) with N3.4 (OND) of the next year	
Epoch	**LCC**
1871-1900	0.009
1901-1930	-0.312
1931-1960	-0.116
1961-1990	-0.443

Izumo et al (2010) had shown that IOD could be precursors for ENSO as described in section 6.3 of this thesis. Negative (positive) IOD tends to be followed by El Niño (La Niña) events about 14 months later. For the period 1960-2010 it is found that the LCC between DMI (Sep-Nov) and N3.4 (Oct-Dec of the following year) is negative (-0.44) which is statistically significant at the 95% level. Table 6.6 gives the LCCs for the four 30-year dry and wet epochs. It is found that the LCCs are statistically significant only in the dry monsoon epochs. In the wet epochs of monsoon the LCCs are very small.

Using the LCCs for the period 1960-2010 a schematic diagram connecting the three factors ISMR, DMI and N3.4 has been given in Figure 6.8. From left to right, a dry (wet) ISMR is followed by an El Niño (La Niña) and positive (negative) IOD, all in the same year. The LCC between DMI of a year and N3.4 of the following year is negative meaning that negative IOD is followed by El Niño in the following year as shown by Izumo et al., (2010). The *sign is to show that this is valid only in the dry epochs of the monsoon like 1901-1930 and 1961-1990.

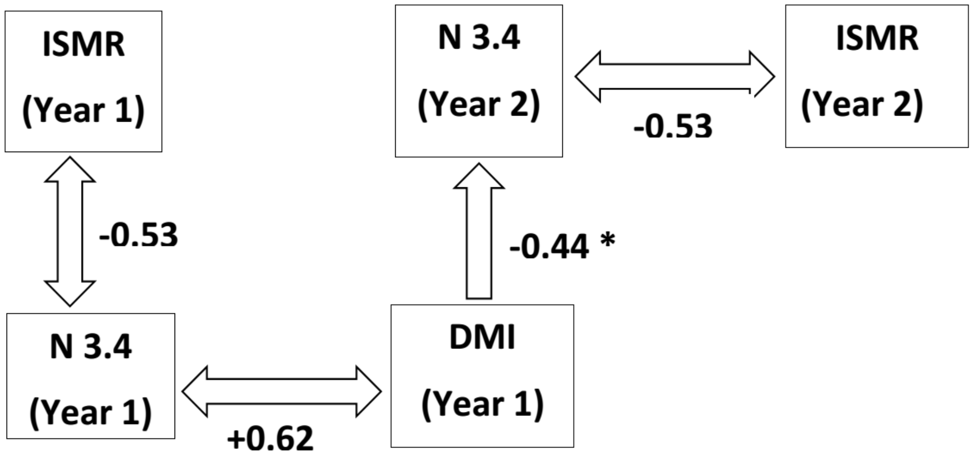


Figure 6.8: Schematic diagram connecting the three factors ISMR, DMI and N3.4 in year -1 and the following year -2. The * mark shows that there is a relation (large LCC) only in the dry epochs of the monsoon. The figure also shows biennial oscillation in ISMR and El Niño

To further understand this aspect of epochal dependence of the LCCs were computed moving 30-year LCCs using data of the period 1871-2010. The variation with time of the LCC between DMI (SON) and N3.4 (OND) of the following year is given in Figure 6.9. The broken line marked is for LCC at the statistical level of significance of 95%. A small portion of the dry epoch 1901-1930 has statistically significant LCC. The whole of the dry epoch 1961-1990 has statistically significant LCC. From the 1940s the magnitude of the LCC between the IOD (DMI of SON) and N3.4 (OND) of the following year is having an increasing trend showing that the El Niño predictability using the IOD index as shown by Izumo et al., (2010) is now increasing.

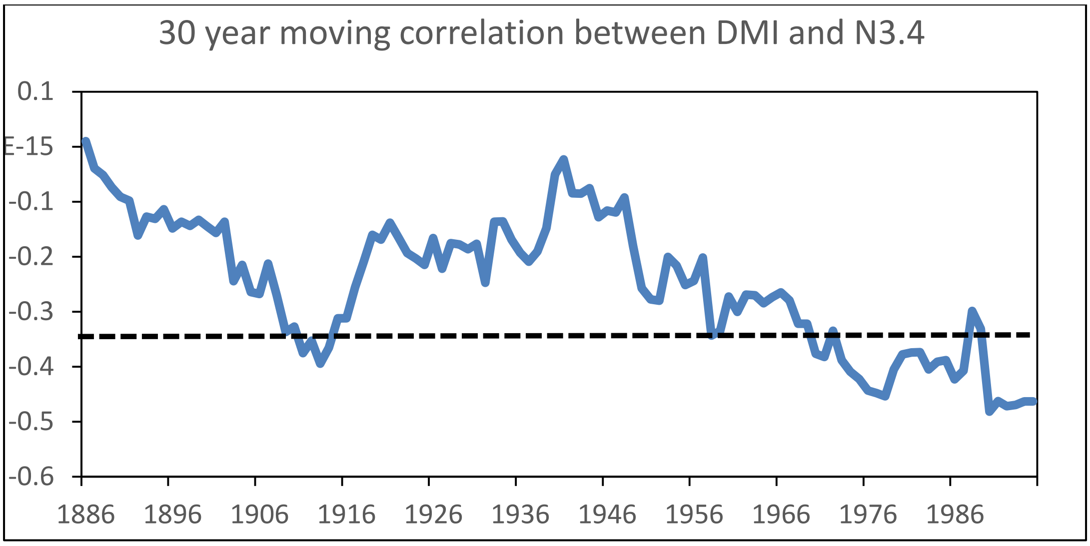


Figure 6.9: Moving 30 year LCCs between DMI (SON) and N3.4 (OND) of the following year using data of the period 1871-2010 is shown by the blue line. LCC at statistical significance of 95% is marked by the broken line

Chapter-7

Summary and Conclusions

7.1 About this thesis

In this thesis a study is made on three important climatic features of the global tropics namely the ISMR, ENSO and the IOD and their inter-relationships and also their relation to the tropical SSTA. The aim of this thesis is outlined in the Preface to this thesis. It is as follows:

I. To investigate the role of El Niño/La Niña and ISMR in global warming, from 1870 to the present through their role in changing the SST of the global oceans this causes changes in the surface air temperature of the global atmosphere.

II. To study the inter-annual variability of DMOK, 1870-2015 and its relation to SSTA, and to generate a long time series 1870-2015 of the DMOK merging all available data on MOK for use in climate change studies.

III. To study the genesis of tropical cyclones over the west Pacific Ocean and its relation to the date of PMRP and DMOK of the period 1959-2014, updating the earlier study done by Joseph (1990a).

IV. To study the inter-relation among the three phenomenon IOD, El Niño/La Niña and ISMR at the inter-annual and decadal time scales.

7.2 Global Warming and its Hiatuses

Greenhouse gas concentration in the atmosphere has been increasing, which is considered as the cause for the warming trend of globally averaged surface air temperature (global warming). However, the long term observations have shown a decadal time scale rapid warming and slow warming (even slight cooling) phases called hiatus occurring alternately. The rapid global warming periods are found to have more frequent El Niños and dry monsoons. During the hiatuses there were more La Niña and wet monsoons. Analysing SST in relation to an El Niño/La Niña Index it was found that the mean SST of the area between latitudes 30^0S and 30^0N and that between 80^0S and 80^0N showed warm SSTA lasting several seasons from the middle of the El Niño year and cooling lasting several seasons from the middle of the La Niña year. The observed global warming was found to be the net result of warming due to the increase of greenhouse gases and warming (cooling) due to El Niño (La Niña) events and the associated dry (wet) monsoons. Regressing tropical belt (30^0S-30^0N) and global (80^0S-80^0N) SSTA to an El Niño/La Niña Index (SSTA of the N3.4 region of October- December) the cumulative SST anomaly year by year were worked out for the two latitude belts which has very realistically reproduced the rapid warming phases of global warming and also the hiatuses during the long period from 1871 to the present.

7.3 DMOK and its variability

Variability of the date of monsoon onset over Kerala has been studied for the long period from 1870 to 2015 and several results have been obtained, some confirming earlier findings using shorter data sets and some new results. Spatially large SSTA mainly in the Pacific and Indian oceans similar to those associated with El Niño (La Niña) were found to be associated with the delayed (early) occurrence of MOK in individual years (confirming old result). There is a statistically significant variation in the date of MOK on the decadal scale during the last 150 years. The SST gradient across the equator over the global oceans (Indian and Pacific) was found to have a high and statistically significant linear correlation coefficient with the DMOK on the decadal scale (new result). This SST gradient persisted during the months prior to MOK (from January). Increased convection and rainfall occurs over the Indian Ocean around DMOK and about 40 days earlier around the PMRP also called “bogus monsoon onset” and in between for about 3 pentads there is a period of suppressed convection and rainfall over the Indian Ocean when increased convection occurs over the west Pacific Ocean (old result). Using the objective DMOK data (free from bogus monsoon onset dates) it has been shown that there is systematic change in the convection and rainfall over tropical Indian ocean with respect to the DMOK during the prior two month period for composites of years of delayed, early and normal monsoon onset dates (new result). It has also shown that DMOK is closely related to the date when the ITCZ crosses the equator to the northern hemisphere in the Indian Ocean (new result).

A long time series for DMOK spanning the period 1870 to 2015 is derived. The details of how these data were merged is given in chapter 2. The period prior to 1971 of this data has not been cleaned from bogus monsoon onsets that occur once in several years, this data set which is given in Table 7.1 will be useful for climate change studies until an objective method is developed for deriving DMOK of the period prior to 1971 free from bogus onset dates. It has been shown in chapter 4 that the DMOK series derived by Ordonez et al., (2016) using available Indian ocean ship wind observations of a long period prior to 1971 cannot be used to extend the objectively derived DMOK series (1971 to date) back to the 1870s.

Table 7.1: Combined time series of the Date of Monsoon Onset over Kerala 1870 to 2015 for use in Climate Change studies. Note that the data prior to 1971 has not been corrected for bogus monsoon onsets that occur once in several years

Year	MOK	Year	MOK	Year	MOK	Year	MOK	Year	MOK
1870	04-Jun	1900	11-Jun	1930	08-Jun	1960	14-May	1990	18-May
1871	01-Jun	1901	07-Jun	1931	04-Jun	1961	18-May	1991	02-Jun
1872	02-Jun	1902	06-Jun	1932	02-Jun	1962	17-May	1992	05-Jun
1873	24-May	1903	12-Jun	1933	22-May	1963	31-May	1993	03-Jun
1874	17-May	1904	07-Jun	1934	08-Jun	1964	06-Jun	1994	28-May
1875	04-Jun	1905	10-Jun	1935	12-Jun	1965	26-May	1995	10-Jun
1876	09-Jun	1906	14-Jun	1936	19-May	1966	31-May	1996	09-Jun
1877	08-Jun	1907	08-Jun	1937	04-Jun	1967	09-Jun	1997	12-Jun
1878	10-Jun	1908	11-Jun	1938	26-May	1968	08-Jun	1998	03-Jun
1879	18-May	1909	02-Jun	1939	05-Jun	1969	17-May	1999	22-May
1880	27-May	1910	02-Jun	1940	14-Jun	1970	26-May	2000	01-Jun
1881	10-Jun	1911	06-Jun	1941	23-May	1971	27-May	2001	26-May
1882	01-Jun	1912	08-Jun	1942	10-Jun	1972	19-Jun	2002	09-Jun
1883	04-Jun	1913	02-Jun	1943	29-May	1973	05-Jun	2003	13-Jun
1884	10-Jun	1914	04-Jun	1944	03-Jun	1974	23-May	2004	03-Jun
1885	05-Jun	1915	15-Jun	1945	05-Jun	1975	02-Jun	2005	07-Jun
1886	01-Jun	1916	02-Jun	1946	29-May	1976	30-May	2006	26-May
1887	02-Jun	1917	31-May	1947	03-Jun	1977	29-May	2007	28-May
1888	29-May	1918	11-May	1948	11-Jun	1978	29-May	2008	31-May
1889	31-May	1919	03-Jun	1949	23-May	1979	12-Jun	2009	23-May
1890	29-May	1920	03-Jun	1950	27-May	1980	03-Jun	2010	31-May
1891	03-Jun	1921	02-Jun	1951	31-May	1981	30-May	2011	29-May
1892	27-May	1922	31-May	1952	20-May	1982	30-May	2012	05-Jun
1893	27-May	1923	11-Jun	1953	07-Jun	1983	12-Jun	2013	01-Jun
1894	05-Jun	1924	02-Jun	1954	31-May	1984	01-Jun	2014	06-Jun
1895	13-Jun	1925	27-May	1955	29-May	1985	24-May	2015	05-Jun
1896	04-Jun	1926	06-Jun	1956	21-May	1986	12-Jun		
1897	07-Jun	1927	27-May	1957	01-Jun	1987	01-Jun		
1898	05-Jun	1928	03-Jun	1958	14-Jun	1988	02-Jun		
1899	28-May	1929	29-May	1959	31-May	1989	04-Jun		

7.4 West Pacific Ocean Tropical cyclone Genesis in relation to PMRP/DMOK

Genesis of tropical cyclones over the west Pacific Ocean and its relation to the date of PMRP and DMOK of the period 1959-2014 has been studied, updating the earlier study done by Joseph (1990,a) using a shorter period of data. The following are the main findings.

(1) There is increased convection in the Indian Ocean around DMOK and PMRP when cyclogenesis does not occur in the west Pacific Ocean.

(2) In EMR (Excess Monsoon Rainfall) years, cyclogenesis does not occur during the first 25 days after DMOK. In contrast in DMR (Deficient Monsoon Rainfall) years, only the first few days after DMOK are free of cyclogenesis.

(3) Typhoon days are many times more in DMR years compared to EMR years.

(4) During the period between PMRP and DMOK cyclogenesis occurs in west Pacific Ocean in DMR years and hardly any in EMR years.

7.5 ISMR, El Niño/La Niña and IOD – inter-relationships

It is found that the main growth of SSTA over the equatorial eastern Pacific ocean, particularly in the Niño3.4 region (positive anomalies in El Niño and negative in La Niña) occur during the four monsoon months June to September and the N3.4 Index (October minus May) has statistically significant LCC with ISMR using data of the 51 year period, 1960- 2010. There are a few out-liners in the scatter diagram relating the two parameters, one group representing strong El Niño years and the other strong La Niña years, totaling about 40% of the data pairs. The LCC between the remaining 60% of the data pairs is very large (LCC=-0.76), showing that the growth of the SSTA in El Niño/La Niña is closely related to the intensity of ISMR. The outliers show that other factors like Madden Julian Oscillation, Active-Break cycle of the monsoon and IOD may be important.

Relationships among pairs of the three phenomena ISMR, El Niño/La Niña and IOD were studied using data of the period 1960-2010 and also for the 30 year long dry and wet epochs during 1871-1990. The LCC between ISMR and N3.4 (Oct-Dec) of the same year is large and negative and the LCC between N3.4 (Oct-Dec) and DMI (Sep-Nov) representing IOD of the same year is large and positive for the period 1960-2010 and also for the thirty year long dry and wet epochs between 1871 and 1990. All these LCCs are also statistically significant. But the LCC between DMI (Sep-Nov) and N3.4 (Oct-Dec of the following year) is statistically significant only for the dry 30 year epochs of the period 1871-1990. Doing a 30-year moving LCC study it was found that since 1940s the magnitude of the LCC between these two factors has been having an increasing linear trend. For the latest 30-year period the LCC is around minus 0.5, statistically significant at the 95% level. This has reference to Izumo et al., (2010) study showing that IOD of the year end can be a predictor for the El Niño during the end of the following year (14 month lead in predicting El Niño).

7.6 Suggestions for future work

IMD is now using an objective method for DMOK. This method uses rainfall, wind data of lower tropospheric levels and OLR for fixing the date of the monsoon onset over Kerala. Using this objective method it has been derived DMOK for the period 1971 to date and not for earlier years in the absence of the required data. The use of wind and OLR data is to free DMOK from bogus monsoon onsets, where increased convection and wind occurs only east of longitude 70^0E. In real monsoon onset over Kerala there is increased lower tropospheric wind and rainfall between equator and latitude 10^0N west of longitude 70^0E. Wind and rainfall data of this area will help us to derive DMOK data free of bogus onsets. There was shipping lane over this area that has high density synoptic observations made by merchant ships from 1870. These observations include data on occurrence of rain (in present and past weather reported) and wind direction and wind strength in Beaufort scale. This data can use to remove bogus monsoon onsets in the DMOK data and derive objective DMOK for the period from 1870-1971.

In Section 6.5 it was shown that the main growth of SSTA over the equatorial eastern Pacific Ocean, particularly in the N3.4 region (positive anomalies in El Niño and negative in La Niña) occur during the four monsoon months June to September suggesting a role for ISMR anomalies in the excitation of El Niño/La Niña. N3.4 (Oct minus May) has high correlation with ISMR. Magnitude of N3.4 (Oct minus May) is found to be very large in strong El Niños and strong La Niñas much more than what this correlation demands. It was suggested in Section 6.5 that other factors like Madden Julian Oscillation, Active-Break cycle of the monsoon and IOD may be responsible for this increase in the magnitude of N3.4 (Oct minus May). It is suggested that research work may be done on this aspect.

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Appendix A

Papers Published in Journals

Preenu P N, P V Joseph and P K Dinesh Kumar (2017) Variability of the Date of Monsoon Onset over Kerala (India) of the period 1870-2014 and its relation to Sea Surface Temperature.

DineshKumar PK, Paul YS, Muraleedharan KR, Murty VS and **Preenu P N** (2016) Comparison of long-term variability of sea surface temperature in the Arabian Sea and Bay of Bengal. Regional Studies in Marine Sciences, 3: 67-75.

Papers Presented in International Conferences

Preenu P N, P V Joseph and P K Dinesh Kumar (2013) Role of El-Nino La-Nina and Asian summer monsoon on Global warming. *India-EU workshop II Monsoon and ocean variability, climate change and sea level variations 11-13 Nov, 2013 Bolgatty Palace & Island Resort, Kochi, India.*

Preenu P N, P V Joseph and P K Dinesh Kumar (2015) Climate change in monsoon onset over Kerala. *World Ocean Science Congress (WOSC) 5-8 Feb 2015, Kochi, India.*

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Variability of the date of monsoon onset over Kerala (India) of the period 1870–2014 and its relation to sea surface temperature

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Monsoon onset over Kerala (India) which occurs every year is a major climatic phenomenon that involves large scale changes in wind, rainfall and sea surface temperature (SST). Over the last 150 years, the date of monsoon onset over Kerala (DMOK) has varied widely, the earliest being 11 May, 1918 and the most delayed being 18 June, 1972. DMOK has a long term (1870–2014) mean of 01 June and standard deviation of 7–8 days. We have studied the inter-annual and decadal time scale variability of DMOK and their relation with SST. We found that SST anomalies of large spatial scale similar to those in El Nino/La Nina are associated with the inter-annual variability in DMOK. Indian Ocean between latitudes 5°S and 20°N has two episodes of active convection associated with monsoon onset over Kerala (MOK), one around DMOK and the other about six weeks earlier (called pre-monsoon rain peak or bogus monsoon onset) and in between a two week period of suppressed convection occurs over north Indian Ocean. A prominent decadal time scale variability was found in DMOK having large and statistically significant linear correlation with the SST gradient across the equator over Indian and Pacific oceans, the large correlation persisting for several months prior to the MOK. However, no linear trend was seen in DMOK during the long period from 1870 to 2014.

Keywords. Sea surface temperature (SST); outgoing long-wave radiation (OLR); date of monsoon onset over Kerala (DMOK); delayed and early monsoon onset; inter-annual and decadal variability in DMOK.

1. Introduction

Monsoon onset over Kerala (MOK) heralds the main rainy season in the Indian subcontinent. Monsoon onset is preceded by large scale changes in the atmosphere and ocean in the Indo-Pacific region (Ananthakrishnan *et al.* 1983; Pearce and Mohanty 1984; Ananthakrishnan and Soman 1988; Soman and Krishna 1993; Joseph *et al.* 1994, 2006). Climatologically monsoon sets in over the extreme south of India (Kerala) by the end of May. Accompanying the monsoon onset there is rapid increase in the daily rainfall of Kerala, vertically integrated moisture in the atmosphere and strength (kinetic energy) of the low level monsoon flow (Krishnamurti 1985).

One of the major parameters associated with the date of MOK (DMOK) is the sea surface temperature (SST) over the Indian and Pacific oceans. The centre of the warm pool of the tropical oceans gets shifted in the annual cycle from south-west Pacific to the north Indian Ocean from January to May (Joseph 1990a, b, 2014). The warm pool of May leads to a build-up of moisture in the atmosphere over north Indian Ocean and the adjoining Pacific Ocean over a month long period and makes conditions favourable for MOK (Pearce and Mohanty 1984; Joseph *et al.* 2006).

The long term mean date of MOK is 1 June. This date has varied widely over the years; the earliest was on 11 May in 1918 and the most delayed monsoon was on 18 June in 1972. Joseph *et al.* (1994) observed that prior to and during MOK dramatic changes occur in the atmosphere and oceans as described earlier (Pearce and Mohanty 1984; Krishnamurti 1985). Joseph *et al.* (1994) showed evidence that delays in MOK are associated with El Nino. Analysis of the SST field also showed that delayed MOK is associated with warm SST anomalies at and south of the equator in Indian and Pacific oceans and cold SST anomalies in the tropical and subtropical oceans to the north during the season prior to the monsoon onset (i.e., March–May). They hypothesized that such SST anomalies caused the inter-annual variability of MOK through their action in affecting the timing of the northward movement across the equator of the cloud band associated with the ITCZ. DMOK has been defined using different criteria that include rainfall, outgoing long wave radiation (OLR), lower tropospheric winds, etc. (Ananthakrishnan and Soman 1988; Joseph *et al.* 2006; Wang *et al.* 2009). India Meteorological Department (IMD) determined the date of MOK using a subjective method for more than 100 years as reviewed by Ananthakrishnan *et al.* (1967). Changes in rainfall and parameters like low level wind and moisture in the atmosphere were seen in a qualitative way to arrive at a subjective estimate of the date of MOK. At MOK rainfall has to be widespread spatially over Kerala and persistent for a few days. In 2006 IMD adopted an objective criteria for declaring MOK using criteria derived by Joseph *et al.* (2006) (as adopted by Pai and Rajeevan 2009) based on daily rainfall of Kerala, the depth of westerlies in a box (equator to latitude 10°N and longitude 55−80°E), the zonal wind speed over the area bounded by latitude 5°−10°N and longitude 70°−80°E and mean OLR in the box latitude 5°−10°N and longitude 70°−75°E. In both the subjective and objective methods, a sharp increase in rainfall of Kerala during MOK is important.

Li and Yanai (1996) showed that the onset of Asian summer monsoon is concurrent with the reversal of the meridional temperature gradient in the upper troposphere south of the Tibetan plateau. This reversal is the result of the large temperature increases in May–June over Eurasia centered on the plateau. The Tibetan heat source according to them is mainly contributed by the sensible heat flux from the ground surface. They however did not study the inter-annual variability of the date of monsoon onset. On the lines of their study Xavier *et al.* (2007) found that the meridional gradient of the tropospheric temperature (averaged between 600 and 200 hPa) is proportional to the meridional gradient of deep tropospheric heating and could lead to acceleration of the deep tropospheric circulation. Their objective definition of the large scale monsoon onset (over India and not Kerala) is based on the reversal of GrTT (Gradient in Tropospheric Temperature as average of 600 to 200 hPa) between a northern box (40°−100°E, 5−35°N) and a southern box (40°−100°E, 15°S−5°N) denoted by GrTT. The onset date (GrTT onset) is defined as the date when GrTT changes sign from negative to positive.

Fasullo and Webster (2003) used the vertically integrated moisture transport through India to define the date of monsoon onset over India (not Kerala). The authors claimed that their index is indicative of the transition in the large scale monsoon circulation over India. Joseph (2012) has compared the objective dates of monsoon onset over Kerala (Pai and Rajeevan 2009) with the monsoon onset dates over India as given by Xavier *et al.* (2007) and Fasullo and Webster (2003) using data of the period 1971–2000 and found the linear correlation coefficients between them are not large (0.59 and 0.56 respectively).

A very recent publication (Ordoñez *et al.* 2016) has given a method for and also derived objective dates for monsoon onset over Kerala using surface wind direction data collected by merchant ships plying over the Indian Ocean during the period 1877–2013, excluding several years where wind data was scarce, particularly, due to world wars. We have, later in this paper, compared this MOK dataset with IMD objective onset dates of the period from 1971.

In this paper we have constructed a long time series of the dates of MOK spanning the period from 1870 to 2014 using all available sources of data on MOK. Such a dataset is required to study the climate change during monsoon and its relation to the changes in the global atmosphere and oceans. We have, in particular, studied the inter-annual and decadal scale variability of the date of MOK and their relationship with SST over the Indian and Pacific oceans.

2. Data used

The onset dates for north Kerala from 1870 to 1900 were taken from Ananthakrishnan and Soman (1989) and for the period 1901–1970 from Ananthakrishnan and Soman (1988). These are objectively determined dates, using only the daily average rainfall of a network of rain-gauge stations in north Kerala. The objective dates of MOK from 1971 to 2005 were taken from Pai and Rajeevan (2009) and for the later years till 2014 from the annual monsoon reports published by IMD in their journal Mausam. We have thus a long dataset of the date of monsoon onset for north Kerala and Kerala for the period 1870–1900 and 1901–2014, respectively. Such a long dataset will be very useful for climate change studies. These dates are given in table 1. Their 11-year moving averages are given in figure 1(a). During the period 1901–1980, we have onset dates for north Kerala by Ananthakrishnan and Soman (1988) and for whole Kerala the subjective onset dates as derived by IMD. For this period, the mean difference in onset dates between these two series is one day, north Kerala onset, date being the earlier one. For declaring monsoon onset, IMD's subjective criteria stipulates two consecutive days of rainfall over Kerala and onset is declared on the second day. Thus there is really no difference in the onset dates of these two series. To get a long series for the onset dates over Kerala (DMOK) for the period 1870–2014, we have adopted the following method. For the period 1870–1900, we have added 1 to the onset dates for north Kerala as derived by Ananthakrishnan and Soman (1989). We have taken the subjective onset dates for Kerala as derived by IMD for the period 1901–1970. To this we have added the objective dates of MOK as derived by IMD of the period 1971 to 2014. This long time series of DMOK is represented by the red bars in figure 1(b) and their 11-year moving average by the black line. The statistical properties of above-mentioned series is given in table 3. The SST gradient in Indo-Pacific region during the season March–April also reflects a similar pattern of inter-annual variability of DMOK is presented in figure 1(c). The mean DMOK for 1870–2014 is June 1st and the standard deviation (SD) is around 7 days. Those who use this long dataset for climate change studies should know that the onset dates of the period 1870–1970 have not been corrected for bogus onset occurrences. Onset dates of this period have been derived without using low level wind or satellite data which are needed for the whole Indian Ocean basin to identify pre-monsoon rain peak or bogus monsoon onset (which occurs once in about 10 years) as described in Joseph *et al.* (2006). In this connection, a study by Sabeerali *et al.* (2012) on the withdrawal dates is relevant and they observed a clear shift prior (posterior) to the 1976/77 climate shift; most of the withdrawal dates are associated with a late (early). However, we did not notice any shift in onset dates.

The objective dates of MOK derived by IMD is from 1971. Ordoñez *et al.* (2016) have derived a method to eliminate bogus onset dates of monsoon using surface wind direction as recoded by merchant ships plying in Indian Ocean. We compared the MOK dates derived by them (called ship wind data) with the objective MOK dates derived by IMD. The difference between these dates (IMD objective onset dates – ship wind onset dates) is positive in most of the years and large positive in several years as may be seen in table 2. We are therefore unable to use this dataset to derive a long dataset of objective MOK dates.

Monthly mean SST data from 1870 to 2014 was obtained from HadISST datasets (Rayner *et al.* 2003). The OLR data has been taken from National Oceanic and Atmospheric Administration (NOAA) and is available from June 1974 to present, except for 1978 when the data is missing due to satellite problems (Liebmann and Smith 1996). The data is interpolated in space to remove the effect of missing values. The OLR data is mapped onto a $2.5^\circ \times 2.5^\circ$ grid size, which represented an average of OLR values twice daily (one daytime and one night time) as the satellite passes. National Oceanic and Atmospheric Administration (NOAA) Optimum Interpolation SST (OISST) Version 2 dataset (Reynolds *et al.* 2007) data (September 1981 to present) have been made available by the NOAA Earth System Research Laboratory Physical Science Division (ESRL/PSD) through their website at http://www.esrl.noaa.gov/psd/. These are daily SST records (one daily value for each pixel), with spatial resolution of 0.250×0.250, based on the Advanced Very High Resolution Radiometer (AVHRR) infrared satellite

Table 1. *Date of monsoon onset over Kerala (DMOK); NK: North Kerala; IMD(S): IMD subjective; IMD(O): IMD objective. Dates are counted in days from 01 May as follows: 01 May = 01; 01 June = 32; etc.*

Year	NK	IMD(S)	IMD(O)	Year	NK	IMD(S)	IMD(O)	Year	NK	IMD(S)	IMD(O)	Year	NK	IMD(S)	IMD(O)	Year	NK	IMD(S)	IMD(O)
1870	34	–	–	1901	36	38	–	1932	15	33	–	1963	35	31	–	1994	–	28	28
1871	31	–	–	1902	37	37	–	1933	22	22	–	1964	35	37	–	1995	–	39	41
1872	32	–	–	1903	43	43	–	1934	39	39	–	1965	37	26	–	1996	–	34	40
1873	23	–	–	1904	32	38	–	1935	45	43	–	1966	31	31	–	1997	–	40	43
1874	16	–	–	1905	39	41	–	1936	21	19	–	1967	40	40	–	1998	–	33	34
1875	34	–	–	1906	44	45	–	1937	34	35	–	1968	40	39	–	1999	–	25	22
1876	39	–	–	1907	37	39	–	1938	26	26	–	1969	32	17	–	2000	–	32	32
1877	38	–	–	1908	41	42	–	1939	37	36	–	1970	26	26	–	2001	–	23	26
1878	40	–	–	1909	33	33	–	1940	45	45	–	1971	25	27	27	2002	–	29	40
1879	17	–	–	1910	37	33	–	1941	22	23	–	1972	53	49	48	2003	–	39	44
1880	26	–	–	1911	35	37	–	1942	41	41	–	1973	37	35	36	2004	–	18	34
1881	40	–	–	1912	36	39	–	1943	13	29	–	1974	23	26	23	2005	–	38	38
1882	31	–	–	1913	35	33	–	1944	34	34	–	1975	31	31	33	2006	–	–	26
1883	34	–	–	1914	35	35	–	1945	36	36	–	1976	31	31	30	2007	–	–	28
1884	40	–	–	1915	42	46	–	1946	34	29	–	1977	38	30	29	2008	–	–	31
1885	35	–	–	1916	28	33	–	1947	33	34	–	1978	29	28	29	2009	–	–	23
1886	31	–	–	1917	30	31	–	1948	40	42	–	1979	43	44	43	2010	–	–	31
1887	32	–	–	1918	8	11	–	1949	13	23	–	1980	31	32	34	2011	–	–	29
1888	28	–	–	1919	35	34	–	1950	27	27	–	1981	–	30	30	2012	–	–	36
1889	30	–	–	1920	34	34	–	1951	32	31	–	1982	–	30	30	2013	–	–	32
1890	28	–	–	1921	37	33	–	1952	33	20	–	1983	–	44	43	2014	–	–	37
1891	33	–	–	1922	31	31	–	1953	48	38	–	1984	–	31	32				
1892	26	–	–	1923	41	42	–	1954	32	31	–	1985	–	28	24				
1893	26	–	–	1924	32	33	–	1955	16	29	–	1986	–	35	43				
1894	35	–	–	1925	27	27	–	1956	20	21	–	1987	–	33	32				
1895	43	–	–	1926	38	37	–	1957	18	32	–	1988	–	26	33				
1896	34	–	–	1927	27	27	–	1958	44	45	–	1989	–	34	35				
1897	37	–	–	1928	34	34	–	1959	15	31	–	1990	–	19	18				
1898	35	–	–	1929	32	29	–	1960	15	14	–	1991	–	33	33				
1899	27	–	–	1930	38	39	–	1961	20	18	–	1992	–	36	36				
1900	41	–	–	1931	30	35	–	1962	10	17	–	1993	–	27	34				

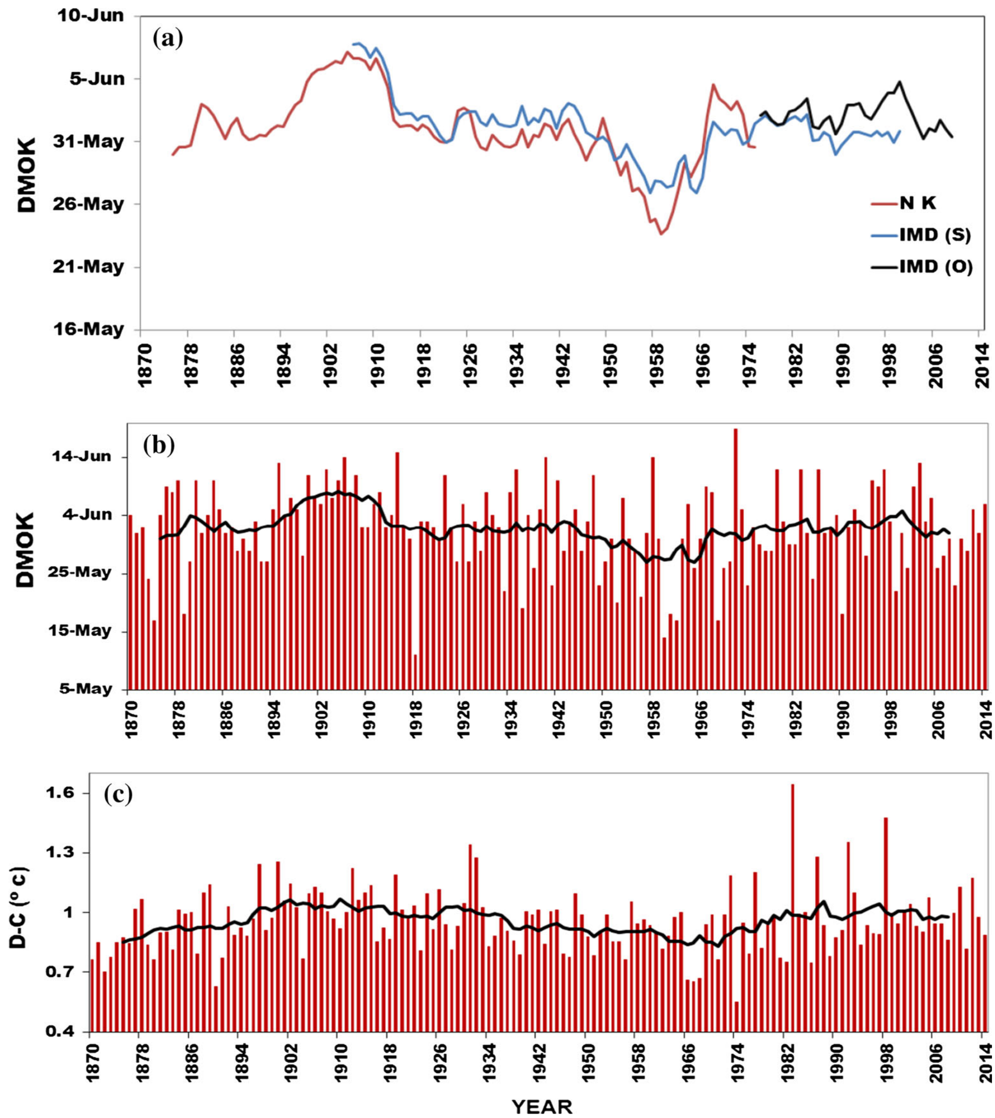

Figure 1. (**a**) 11-year moving averages of dates of monsoon onset over north Kerala, IMD monsoon onset over Kerala (subjective) and IMD monsoon onset over Kerala (objective). (**b**) Date of monsoon onset over Kerala for the period 1870–2014 combining the three series given at (**a**) is shown in red bars and their 11-year moving average is marked by black line. (**c**) SST difference for each year of 1870–2014 between the latitude–longitude boxes 30°−270E, 0−20S (D) and 30°−270°E, 5°−20°N (C) for the season March–April and its 11-year moving averages.

measurements. The ENSO index is based on the SST anomalies averaged in the Niño3.4 region (170°−120°W, 5°S−5°N). The 'Dipole Mode Index' (DMI) is based on the SST anomalies in 50°−70°E, 10°S−10°N minus those in 90°−110°E, 10°S−0°N (Saji *et al.* 1999).

3. Inter-annual variability of MOK and its relation with SST anomalies

SST plays a vital role in the timing of monsoon onset over Kerala (Joseph *et al.* 1994, 2006; Joseph 2014). SST anomalies on either side of equator in Indian and Pacific oceans are found to be related to the date of MOK. In our study, MOK is considered as delayed if it has occurred eight days or more (one standard deviation) after 01 June. Similarly, if MOK is eight days or more before 01 June, it is defined as early. Joseph (2014) found the linear correlation coefficient (LCC) between the objective dates of MOK and the SST using 30 years of data (1971–2000). We have used longer period data and found the LCC between IMD's objective DMOK and SST of January–February and March–April for the period 1971–2014 which is given in figure 2. The LCCs are high and statistically significant over large spatial areas marked by the blue lines in the figure which gives correlation coefficient of 0.3 which is significant at 95% level using t-test.

Table 2. *Difference (in days) between the DMOKs (IMD objective minus ship wind) of each year 1972 to 2005.*

	IMD objective onset date minus ship wind onset date			
−5 and below	−4, −3, −2	−1, 0, +1	+2, +3, +4	+5 and above
	1982	1974, 1987	1975, 1977	1972 (06 days)
		1992, 2005	1979, 1983	1973 (10 days)
			1984, 1985	1976 (05 days)
			1988, 1990	1978 (10 days)
			1991, 1993	1980 (06 days)
			1994, 1997	1981 (05 days)
			1998	1986 (06 days)
				1989 (11 days)
				1995 (07 days)
				1996 (06 days)
				1999 (10 days)
				2000 (18 days)
				2001 (08 days)
				2002 (17 days)
				2003 (08 days)
				2004 (21 days)
	1 year	4 years	13 years	16 years

MOK is delayed when the SST anomaly is positive at south of the equator and negative at north of it. This correlation pattern was also persistent for several months prior to MOK. The LCC patterns over the Pacific and Indian oceans show that delayed MOK is associated with El Nino whereas an early MOK is associated with La Nina. The association of delayed MOK with El Nino has been shown by Joseph *et al.* (1994). The composite SST anomalies of January–February and March–April of delayed and early DMOK years of the period 1901–2014 are given in figure 3. According to the study by Joseph *et al.* (1994), monsoon onset delays occur in El Nino (0) year and more often in El Nino (+1) year. During the period 1870–2014, there were 21 cases of monsoon onset delays of one standard deviation or more and 13 out of these 21 cases have occurred in El Nino(0) or (+1) year (table 4). For compositing, the SST anomalies were derived by subtracting an 11-year mean SST centered over the year for each case. The SST anomalies are similar to those derived from the correlation study in figure 2. Further, we computed first EOF mode of SST over Indo-Pacific domain and this mode depict El Nino related SST anomaly in Pacific (figure 4). The time series of this anomaly has statistically significant correlation of 0.2 with the monsoon onset dates of 1870–2014, for the January–February season. For the March–April season, the correlation is smaller. However, the correlation between dates of monsoon onset and Nino 3.4 index of January–February and March–April seasons are statistically significant at 95% level (figure 5). The correlation between dates of monsoon onset and IOD index (figure not shown) is also tested but is not significant. The spatial pattern of EOF-1 shows that in delayed onset years there is warm SST anomaly south of the equator in Indian and Pacific oceans. EOF-1 is similar to the figure showing the correlation between date of monsoon onset and SST as shown in figure 2.

Table 3. *Mean and Standard deviation of different series.*

Series	Mean	SD
NK: 1870 to 1900	01 June	6.5 days
IMD(S): 1901 to 1970	01 June	7.9 days
IMD(O): 1971 to 2014	01 June	6.7 days
NK+1: 1870 to 1900+ IMD(S): 1901 to 1970+ IMD(O): 1971 to 2014	01 June	7.2 days

NK: North Kerala; IMD(S): IMD subjective; IMD(O): IMD objective.

4. Evolution of OLR in delayed and early MOK

As described in Introduction, the center of the warm pool (area of maximum SST) shifts in the annual cycle from southwest Pacific Ocean to north

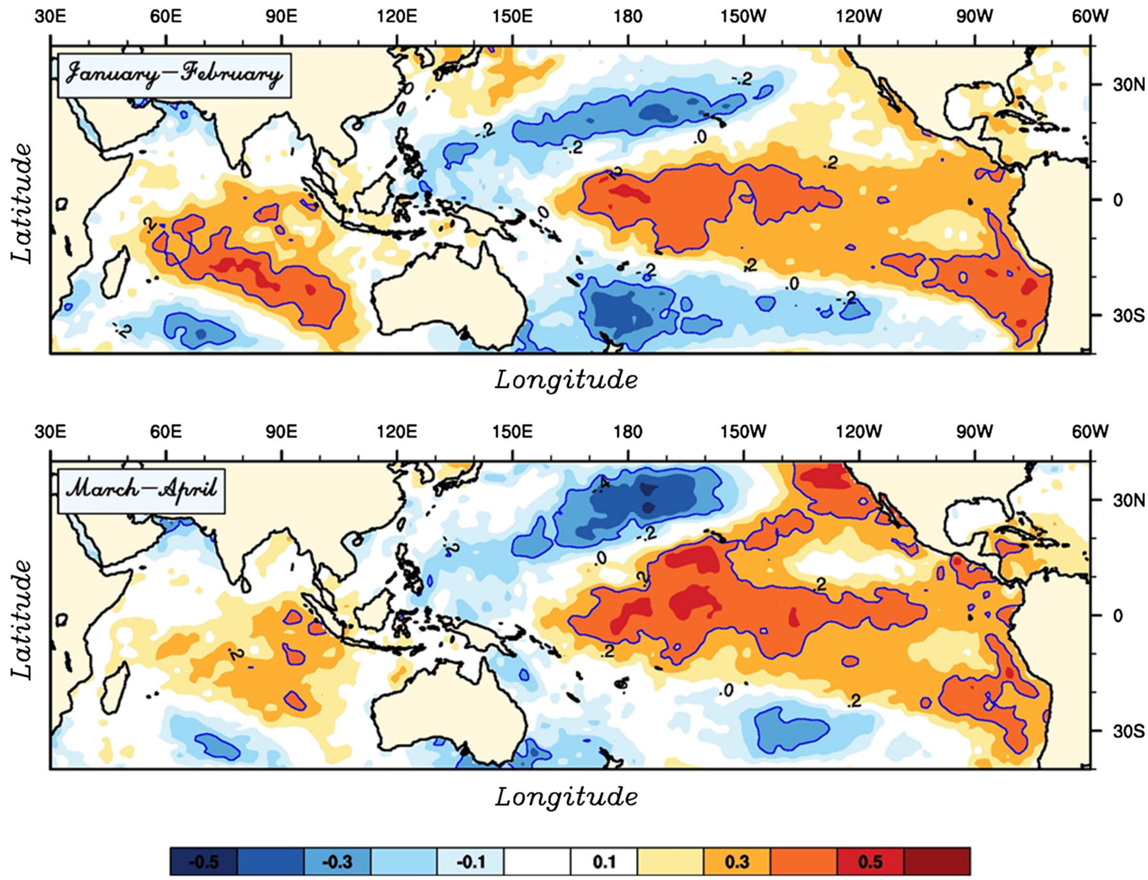

Figure 2. Linear correlation coefficient between HadISST and IMD's objective dates of MOK from 1971 to 2014 for January–February (above) and March–April (below). The significant values are marked by thick blue line. Linear correlation coefficient is 0.3 for a significance level of 95% by a t-test and $r = 0.3$ is marked by a blue line.

Indian Ocean from January to May. But there is a fine structure to the changes in SST and the associated convection (OLR) in Northern Indian Ocean prior to MOK. The fine structure of the changes in SST for the MOK of 2003 is described in Joseph (2014). SST of north Indian Ocean reaches maximum first over the Bay of Bengal, about 7–8 pentads before the date of MOK, when SST reaches high values of 31°C or even 32°C in central Bay of Bengal. To the south of the region of this SST maximum, in the area of large SST gradient, deep convection develops and grows in area and intensity. The latent heat released in this convective area heats the atmosphere there and generates strong cross equatorial low level wind flow. The convective clouds and the low level winds cool the Bay of Bengal. When these changes are happening in the Bay of Bengal, Arabian Sea has much lower SST with very little convection and only feeble low level winds which make the SST of Arabian Sea warm rapidly. About three pentads before MOK, SST over central Arabian Sea reaches 31°C or 32°C and in the SST gradient area south of this region convection and strong low level winds develop the northward movement which brings about monsoon onset over Kerala.

We studied the evolution of convection (OLR) prior to MOK for the delayed and early monsoon onset years. Composite pentad OLR charts are made for the delayed onset years (1979, 1983, 1986, 1995, 1996, 1997, 2002 and 2003) and for the early onset years (1974, 1985, 1990, 1999 and 2009) for several pentads prior to MOK with MOK taken as occurring around zero pentad.

Figure 6(a) shows the composite mean pentad variations in the OLR of delayed onset years of the period 1974–2014 (see table 4). Zero pentad is that of monsoon onset over Kerala. Coinciding with pentad-10 (22–26 April) there is deep convection (low values of OLR) around equatorial Bay of Bengal extending to the west Pacific Ocean. This band of convection increases in size and moves north up to pentad-7 (07–11 May); when its left-most portion brings rain over Kerala. This has been called pre-monsoon rainfall peak (PMRP) by Joseph and Pillai (1988) and 'bogus

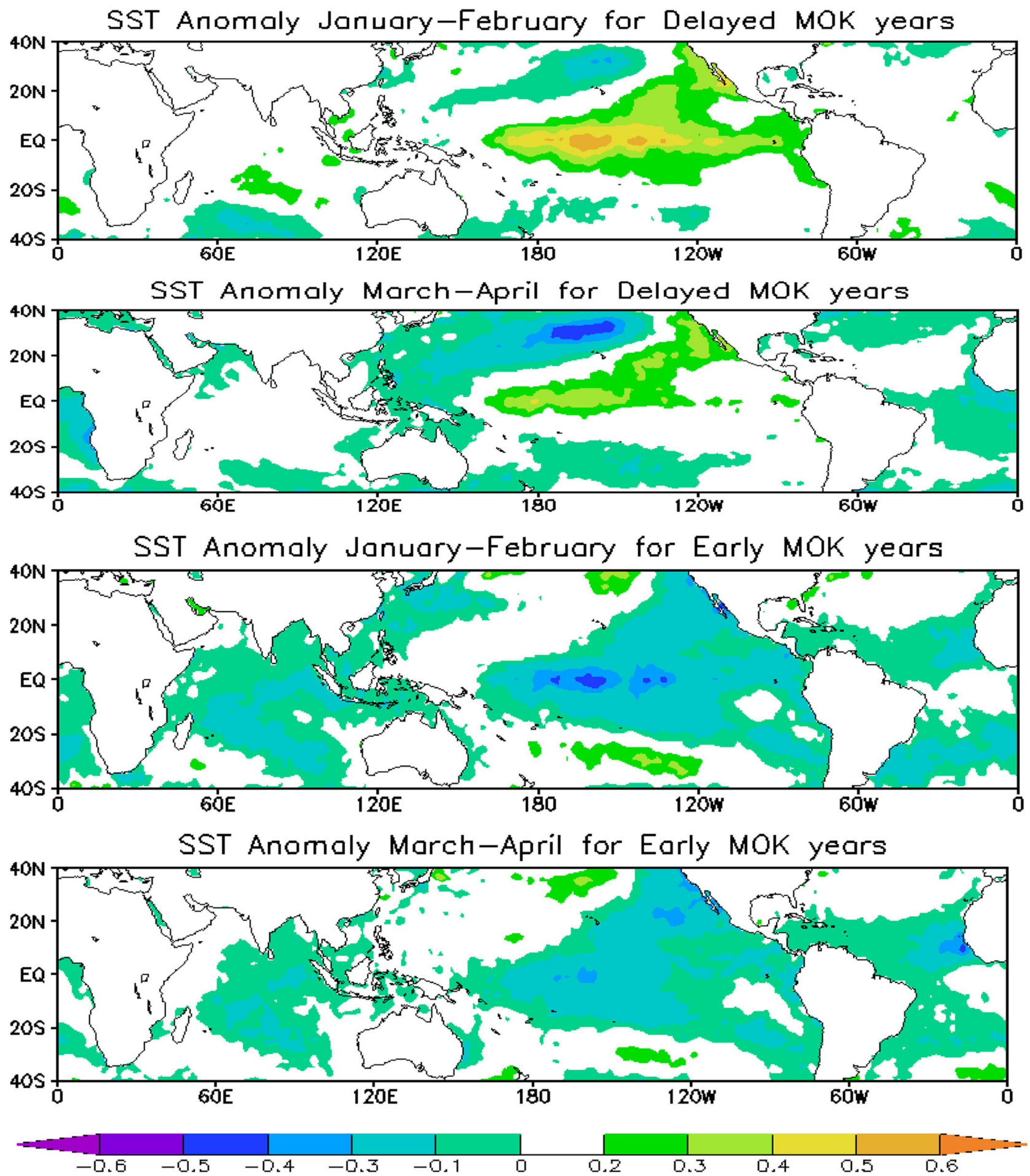

Figure 3. The composite SST anomaly of the January–February and March–April of delayed and early DMOK years of the period 1901–2014.

Table 4. *Years of delayed monsoon onset over Kerala (MOK) during 1901–2014. The El Nino year is marked by (0) and the following year by (+1).*

Years of delayed MOK		
1903(+1)	1940(0)	1983(+1)
1905(0)	1942(+1)	1986(0)
1906	1948	1995(+1)
1908	1958(+1)	1996
1915	1967	1997(0)
1923(0)	1972(0)	2002(0)
1935	1979	2003(+1)

monsoon onset' by Flatau *et al.* (2001) associated with strong low level winds. Soon after the PMRP, the band of convection moves northeastward to usher in monsoon onset in some years over the Indo-China peninsula at pentads-6 and -5 as shown by Joseph *et al.* (2006). At pentad-4 (22–26 May), the OLR field suggests that Indian Ocean is practically free of convection. At pentad-3 (27–31 May) a fresh elongated narrow band of convection formed close to the equator (in the Indian Ocean south of Arabian Sea). This band of convection grows rapidly in intensity and area as seen by the OLR of pentads-2 (01–05 June) and -1 (06–10 June), bringing about the monsoon onset over Kerala at zero pentad (11–15 June).

The composite mean OLR for early onset years for selected pentads is given in figure 6(b). The

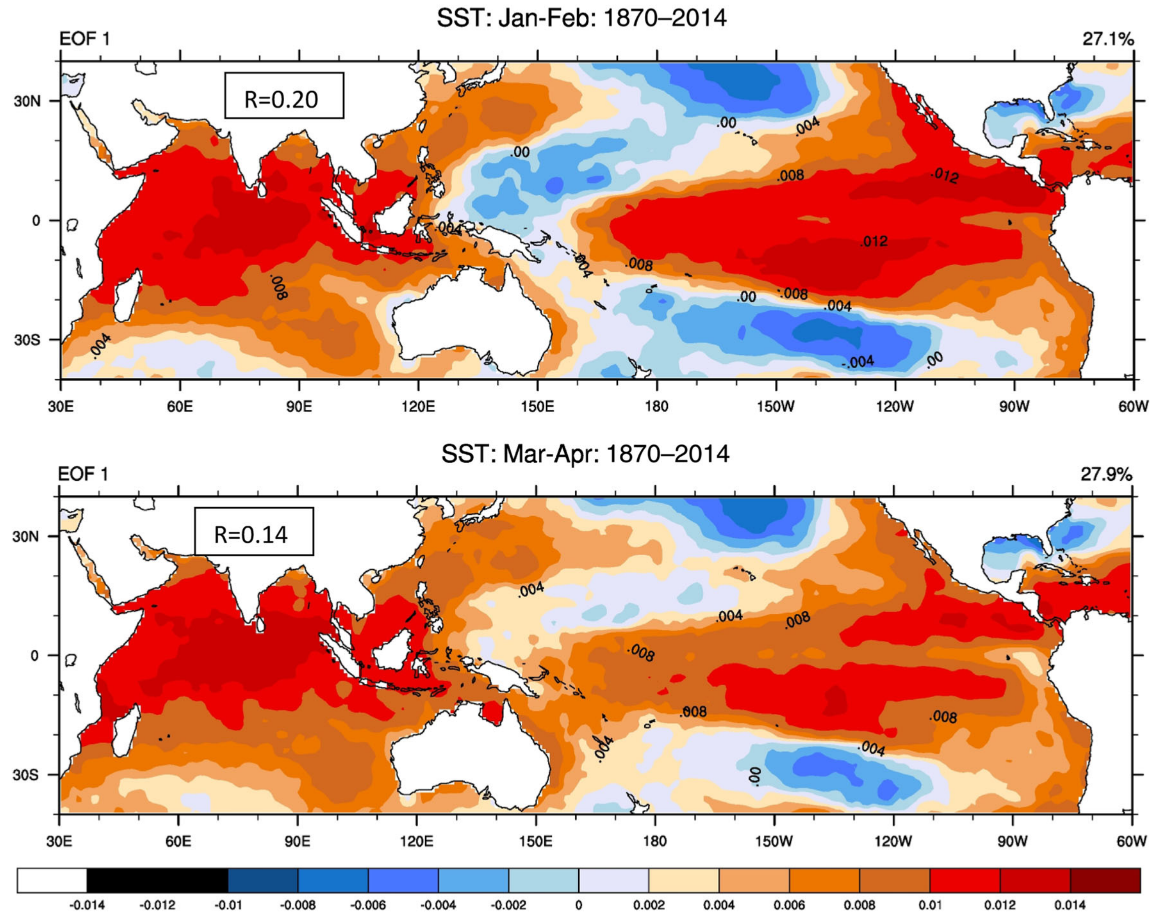

Figure 4. EOF1 of SST over Indo-Pacific domain for the seasons January–February (above) and March–April (below). The correlation of time series of first PC (principal component) with monsoon onset dates are marked by R.

development and migration of the convection bands for the early MOK years is similar pentad-wise to delayed MOK years but the PMRP and MOK are found to occur four pentads earlier date-wise. Thus PMRP for early MOK composite is on pentad 7–11 April and MOK on pentad 22–26 May. We also constructed composite OLR charts for normal onset years (figure not shown). The pentad-wise evolution of OLR with zero pentad taken as DMOK is the same as for late and early onset composites. These three composites show that there is a minimum of convection in north Indian Ocean during pentads-5, -4 and -3.

5. Why convection occurs in Bay of Bengal earlier than Arabian Sea?

A question arises as to why SST reaches maximum in the annual cycle January–May, first in Bay of Bengal followed by Arabian Sea nearly a month later. Figure 7(a, b) gives the monthly mean SST averaged over the period 1974–2014 for the months of March and April. The axis of maximum SST is oriented from south-west to north-east (in west Indian Ocean, the SST maximum is close to the equator and in the east Indian Ocean it is at about latitude 10°N). It is speculated that this south-west to north-east orientation is taken by the axis of SST maximum due to the strong cooling of SST of the Arabian Sea during the previous monsoon season. The strong low level winds of the monsoon are known to cool the Arabian Sea by causing coastal and open ocean upwelling and by evaporative cooling of the ocean surface. Figure 8(a) shows the day-to-day change of the mean SST of the Bay of Bengal Box (red curve) bounded by latitudes 10°−20°N and longitudes 85°−95°E and the Arabian Sea box (blue curve) bounded by latitudes 10°−20°N and longitudes 60°−70°E. Bay of Bengal box reaches SST maximum first around mid-April whereas the Arabian Sea box reaches SST maximum about a month later. We have also represented this by a Hovmöller diagram showing

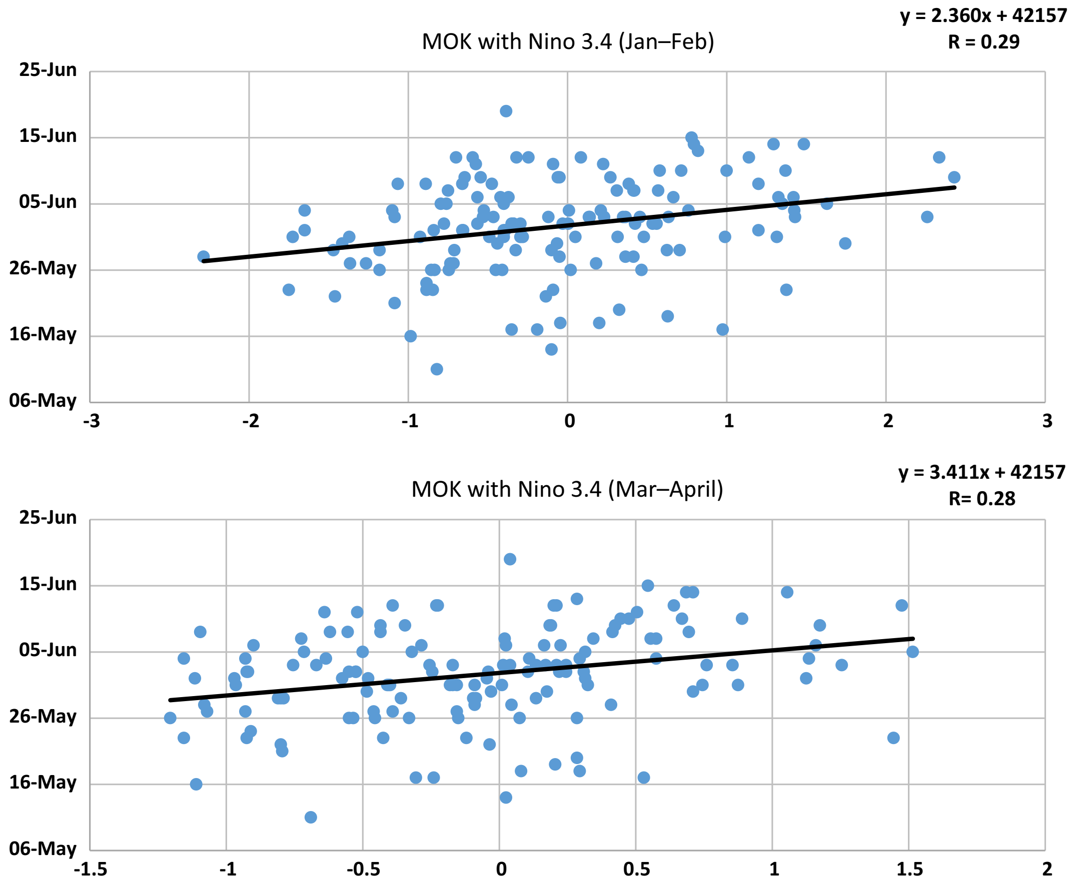

Figure 5. Correlation between dates of monsoon onset and Nino 3.4 index of January–February (above) and March–April (below).

the time latitude change of the mean SST averaged over the longitudes 85°−95°E over the Bay of Bengal (figure 8b) and 60°−70°E over the Arabian Sea (figure 8c).

6. Decadal variability in the date of MOK

Figure 1(b) gives date of MOK from 1870 to 2014. It can be seen that in the beginning of last century, in many years of the decades 1900s and 1910s, MOK was delayed by one to two weeks, whereas in the middle of the last century, in many years of the decades 1950s and 1960s, MOK occurred early by one or two weeks. The mean date of MOK for the period 1900–1920 is 05 June with a standard deviation of 7.2 days. Similar values for the period 1950–1970 are 26 May and 8.5 days, respectively. The difference in the mean dates of MOK for these two 21-year periods is statistically significant at the 99% level as per t-test. Thus on a decadal scale, there has been a statistically significant change in the date of MOK during the last 150 years. There is, however, no linear trend for this long time series.

In order to study the influence of SST in the decadal time scale variability of DMOK, we selected five boxes (A–E) over the oceans surrounding the equator as given in table 5. The LCC between 11-year moving averages of SST gradient across the equator and MOK of pairs of these boxes is shown in table 6. The boxes A (30°−120°E, 5°−20°N) and B (30°−120°E, 0−20°S) represent areas north and south of the equator in the Indian Ocean. The LCC between 11-year moving average of SST gradient and MOK in these boxes during January and February (0.23) which increased to 0.58 during March and April. Similar values for the boxes C (30°−270°E, 5°−20°N) and D (30°−270°E, 0−20°S) representing north and south tropical Indian and Pacific oceans have the maximum LCC compared to all other LCC. They are 0.63 and 0.67 respectively. Figure 1(c) gives SST difference (gradient) for March and April every year during 1870–2014 between the latitude–longitude boxes C (30°−270°E, 5°−20°N) and D

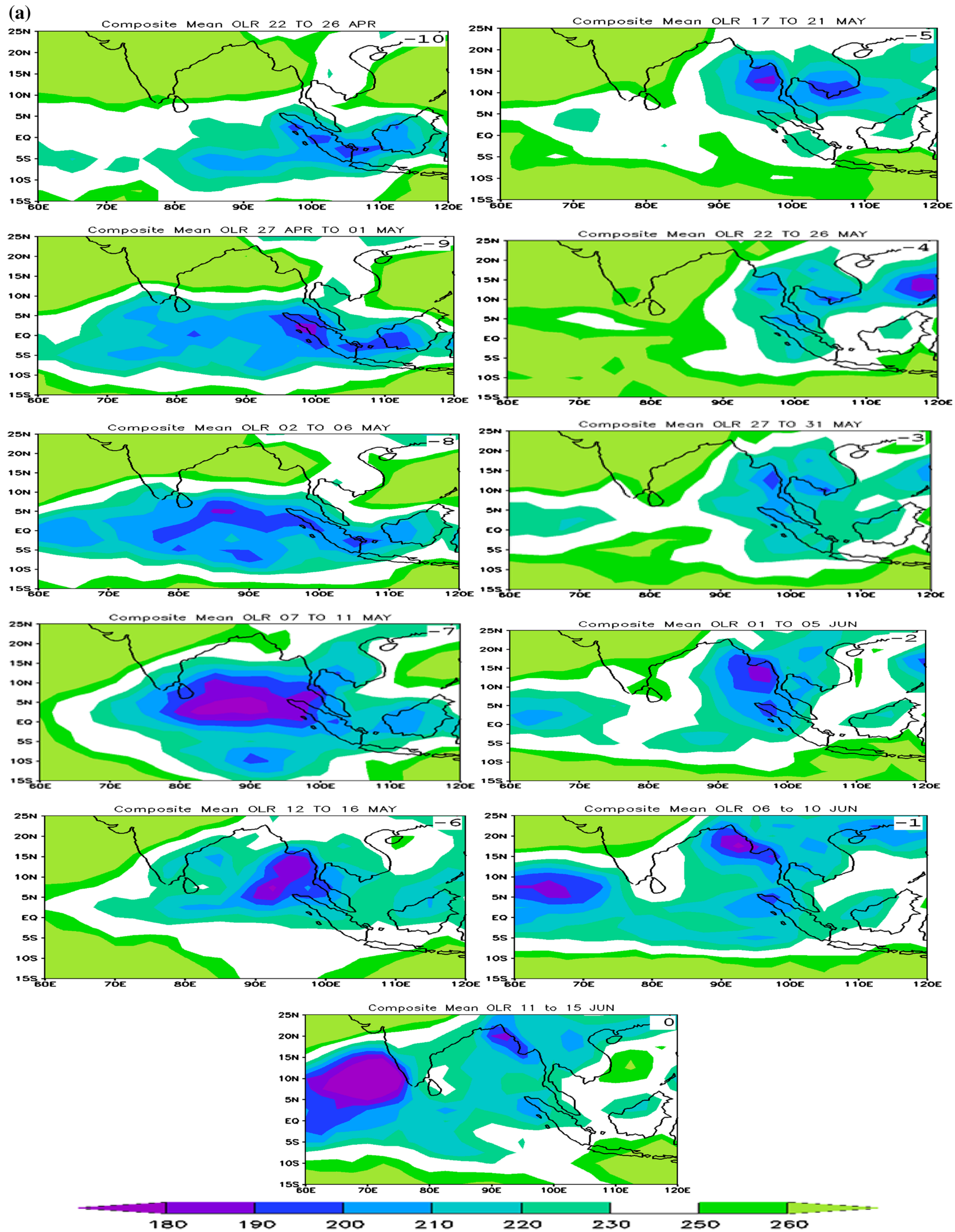

Figure 6. (**a**) The composite mean OLR (Wm^{-2}) for delayed MOK years (1979, 1983, 1986, 1995, 1996, 1997, 2002 and 2003) (pentads marked on top right corner, MOK is taken as zero pentad). (**b**) The composite mean OLR (Wm^{-2}) for early MOK years (1974, 1985, 1990, 1999 and 2009) (pentads marked on top right corner, MOK is taken as zero pentad).

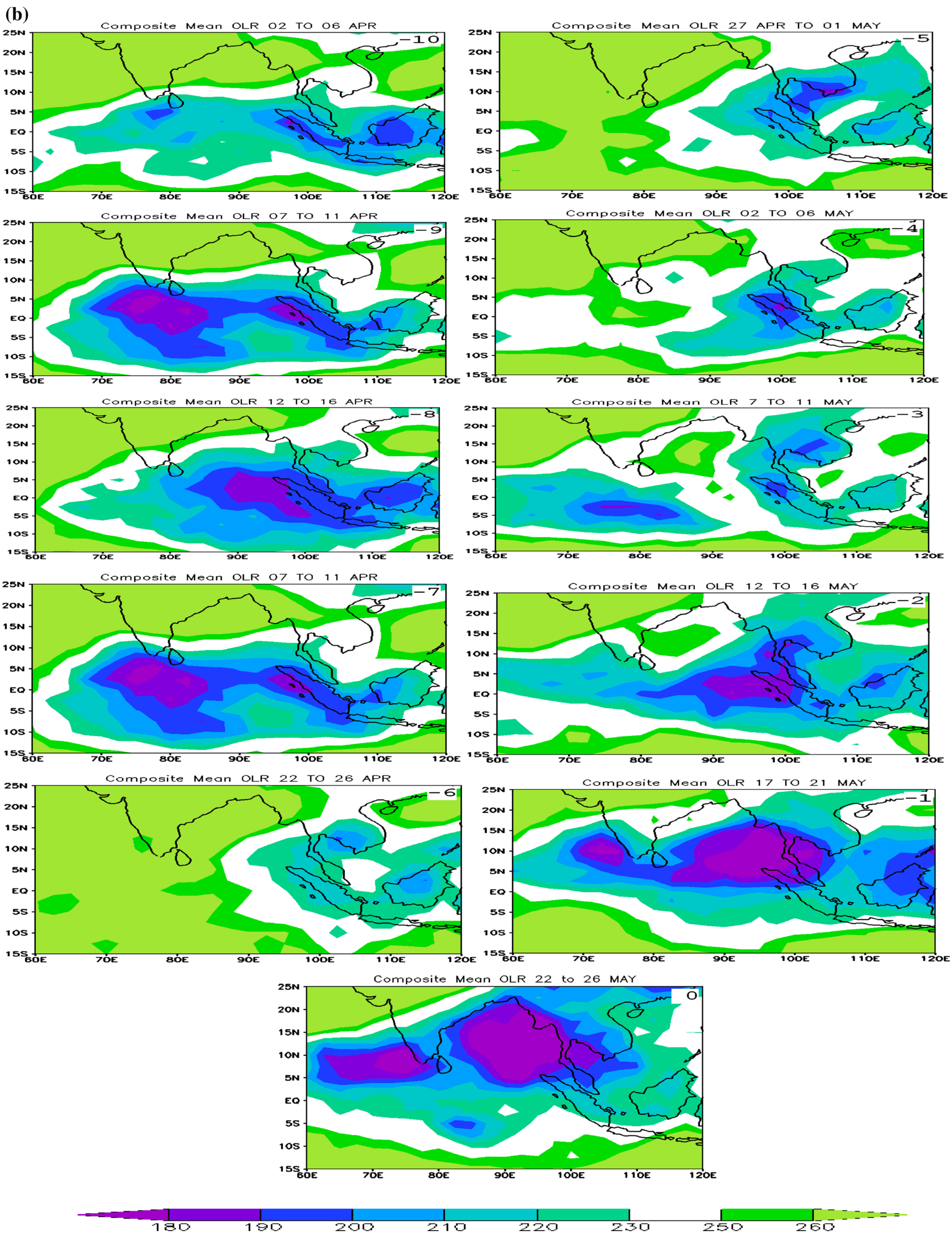

Figure 6. (*Continued.*)

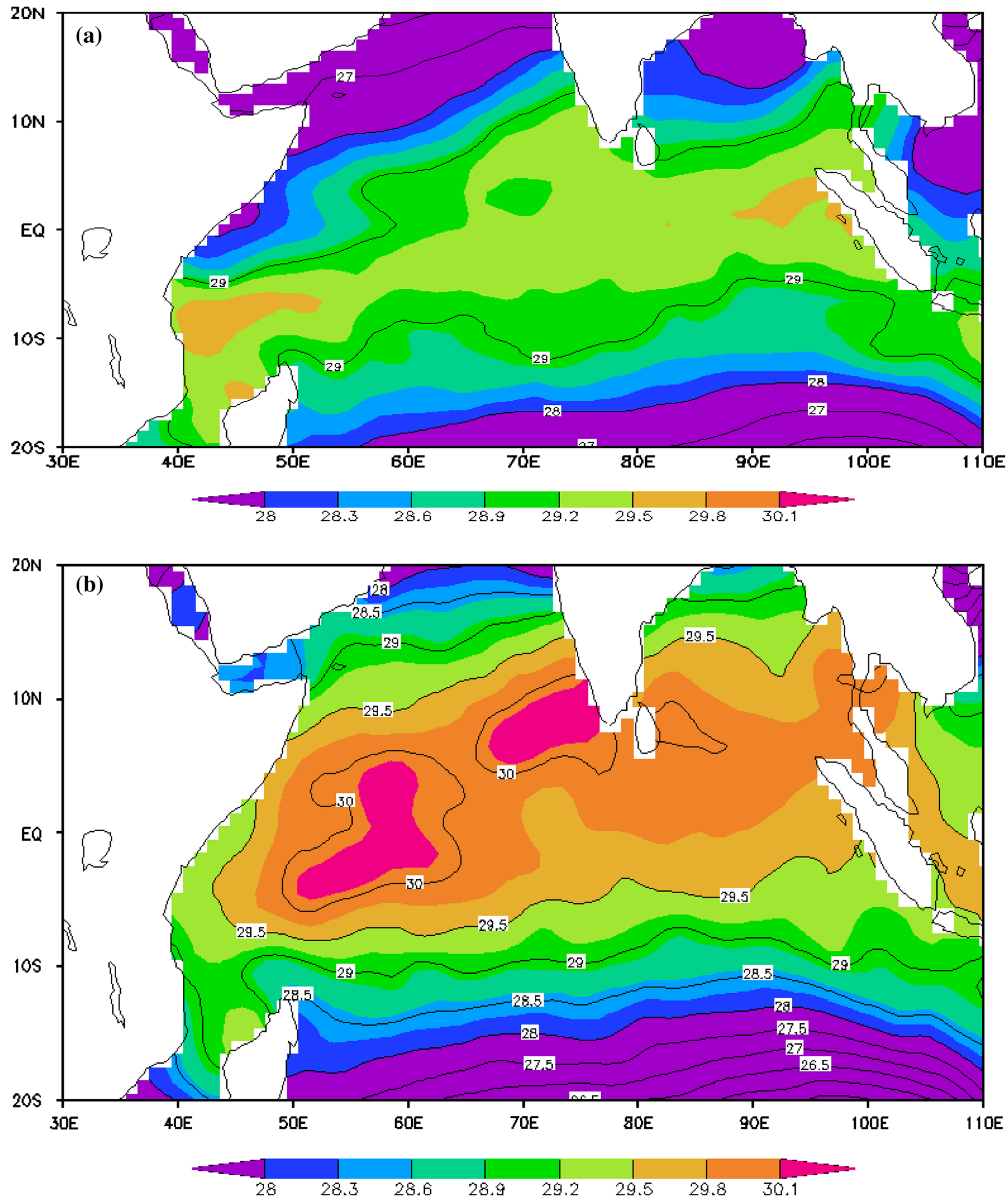

Figure 7. (**a**) Mean SST (°C) (1974–2014) for the month March. (**b**) Mean SST (°C) (1974–2014) for the month April.

(30°−270°E, 0−20°S) with the 11-year moving average marked. The mean of SST gradient between these boxes for period 1900–1920 is 1.04 with an SD of 0.13 in the season March and April. Similar values for the period 1950–1970 are 0.88 and 0.12 respectively. The difference in the mean SST for these two periods is statistically significant at 99% level as per t-test. The LCC of 0.67 between the 11-year moving averages during March–April SST gradient and the date of MOK for the period 1870–2014 is large and statistically significant. Similarly, the LCC for the January–February SST gradient is 0.61 which is also large. Both the correlations are statistically significant at 99% level. In the introduction it was mentioned that the center of the warm pool of the tropical oceans gets shifted in the annual cycle from southwest Pacific to the north Indian Ocean during January–May (Joseph 1990a, b, 2014) and the warm pool of May leads to a build-up of

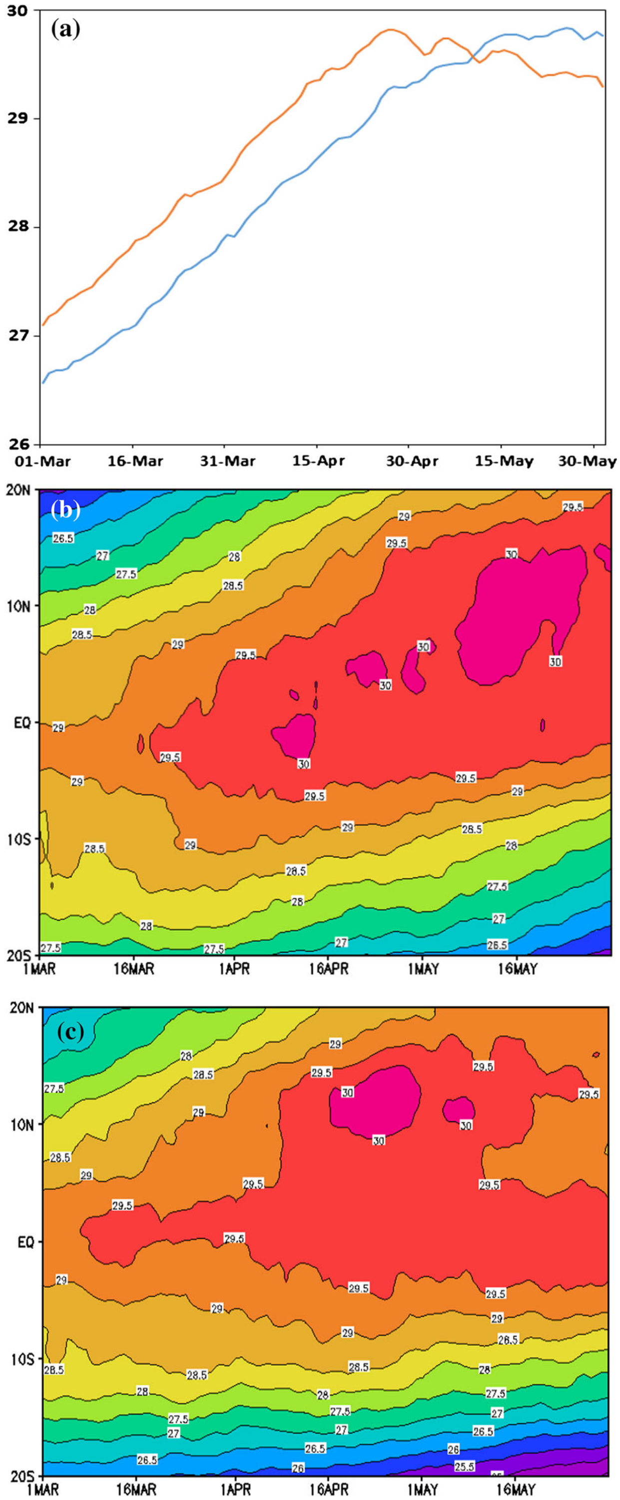

Figure 8. (**a**) Mean daily SST (°C) of 1991 to 2010 for 01 March to 31 May of the area 60°–70°E, 10°–20°N (blue curve) and 85°–95°E, 10°–20°N (red curve). (**b**) Hovmöller of mean daily SST (°C) of 1991–2010 between longitudes 60°E and 70°E for the period 01 March to 31 May and latitudes 20°S to 20°N. (**c**) Hovmöller of mean daily SST (°C) of 1991–2010 between longitudes 85°E and 95°E for the period 01 March to 31 May and latitudes 20°S to 20°N.

Table 5. *Latitude and longitude of box boundaries.*

Box	Longitude	Latitude
A	30°–120°E	5°–20°N
B	30°–120°E	0°–20°S
C	30°–270°E	5°–20°N
D	30°–270°E	0°–20°S
E	120°–180°E	0°–15°S

moisture in the atmosphere over north Indian Ocean and the adjoining Pacific Ocean making conditions favourable for MOK. We studied the LCC between the 11-year moving averages of the SST gradient between the boxes E (south-west Pacific) and A (north Indian Ocean) of DMOK. The LCC for the months January–February is very low of 0.02, but that for March–April is high at 0.63 and statistically significant at 99% level.

7. ITCZ transitions across the equator and DMOK

Joseph *et al.* (1994) had hypothesised that the time of transition of the ITCZ from south to north across the equator is related to DMOK, delayed crossing the equator of the ITCZ associated with delay in DMOK. The ITCZ in monthly climatology has large north–south movement over the Indian Ocean (the longitude zone 30°–120°E) studied in this paper due to the monsoonal character of the area. In a large part of the longitude zone 120°–270°E, the ITCZ lies only in the northern hemisphere both in January and July at low latitudes with only a few degrees latitude difference. We have made composite Hovmöller diagrams of the daily OLR averaged between longitudes 30° and 120°E (figure 9) and 120° and 270°E (figure 10) of the period 01 April–30 June of years of delayed and early monsoons. For the longitude belt 30°–120°E (Indian Ocean), the minimum OLR zone has large south to north movement in association with MOK and smaller south to north motion in association with PMRP both for delayed and early MOK. These features have considerable time difference, the crossing of the equator occurring much later in delayed monsoons as compared to early monsoons. For the longitude zone 120°–270°E (Pacific Ocean), there is no equator crossing of the low OLR zone but only *in-situ* deepening of convection in the ITCZ occurring earlier or later.

Table 6. *LCC between the 11-year moving average of MOK and 11-year moving average of SST gradient for the seasons January–February and March–April.*

Series	Correlations					
	B-A		D-C		E-A	
	JF	MA	JF	MA	JF	MA
NK: 1870 to 1980	0.04	**0.52**	**0.51**	**0.53**	−0.09	**0.67**
IMD(S): 1901 to 2005	0.14	**0.52**	**0.50**	**0.59**	−0.07	**0.63**
IMD(S): 1901 to 1970+ IMD(O): 1971 to 2014	0.08	**0.50**	**0.57**	**0.67**	−0.11	**0.56**
NK+1: 1870 to 1900 + IMD(S): 1901 to 1970+IMD(O): 1971 to 2014	*0.23*	**0.58**	**0.63**	**0.67**	0.02	**0.63**

The value which is 99.9% significant is shown in bold and 95% significant is shown in italics.

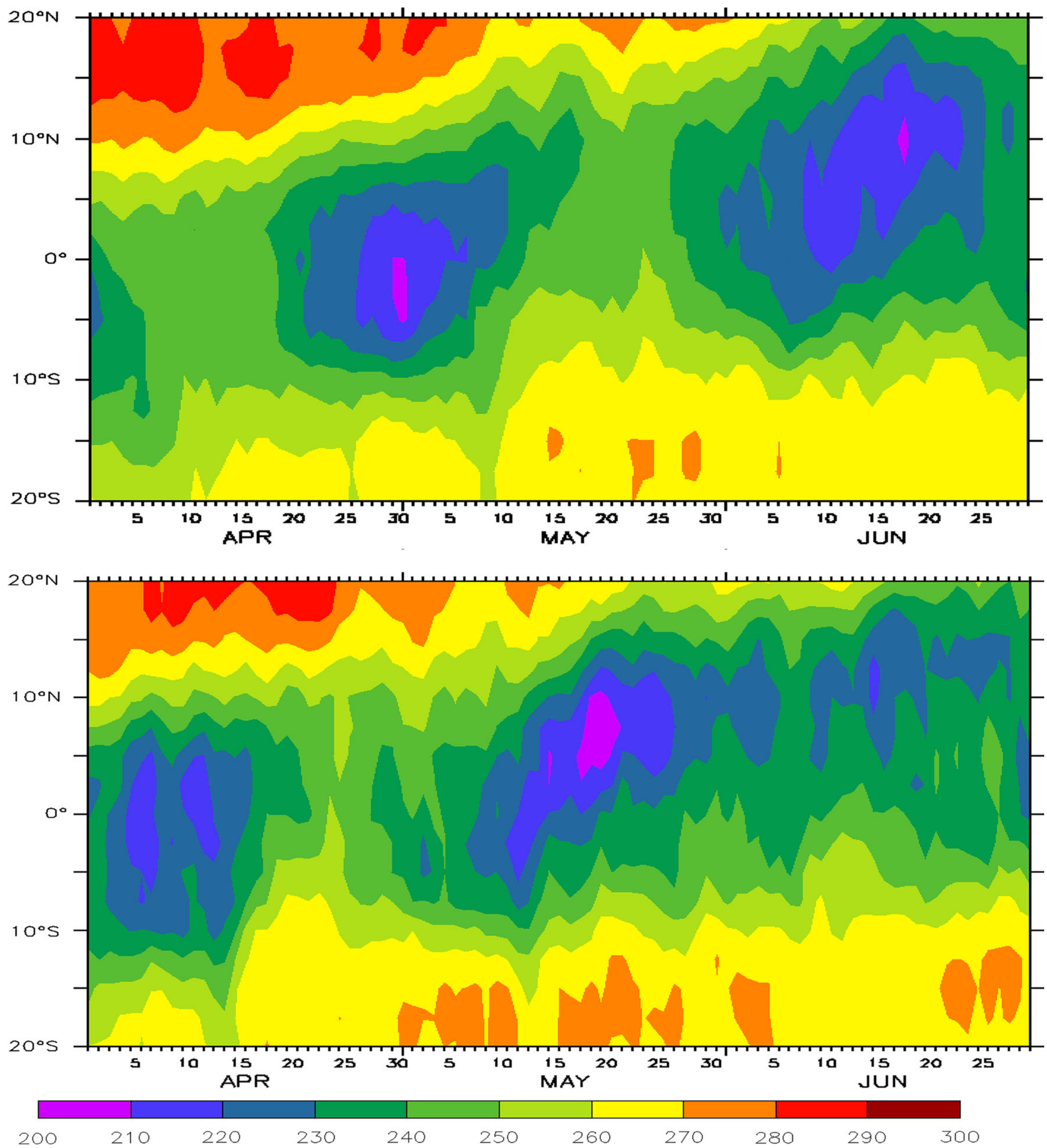

Figure 9. Hovmöller diagrams of the daily OLR (Wm^{-2}) averaged between longitudes 30° and 120°E (Indian Ocean) with delayed (above) and early (below) MOK.

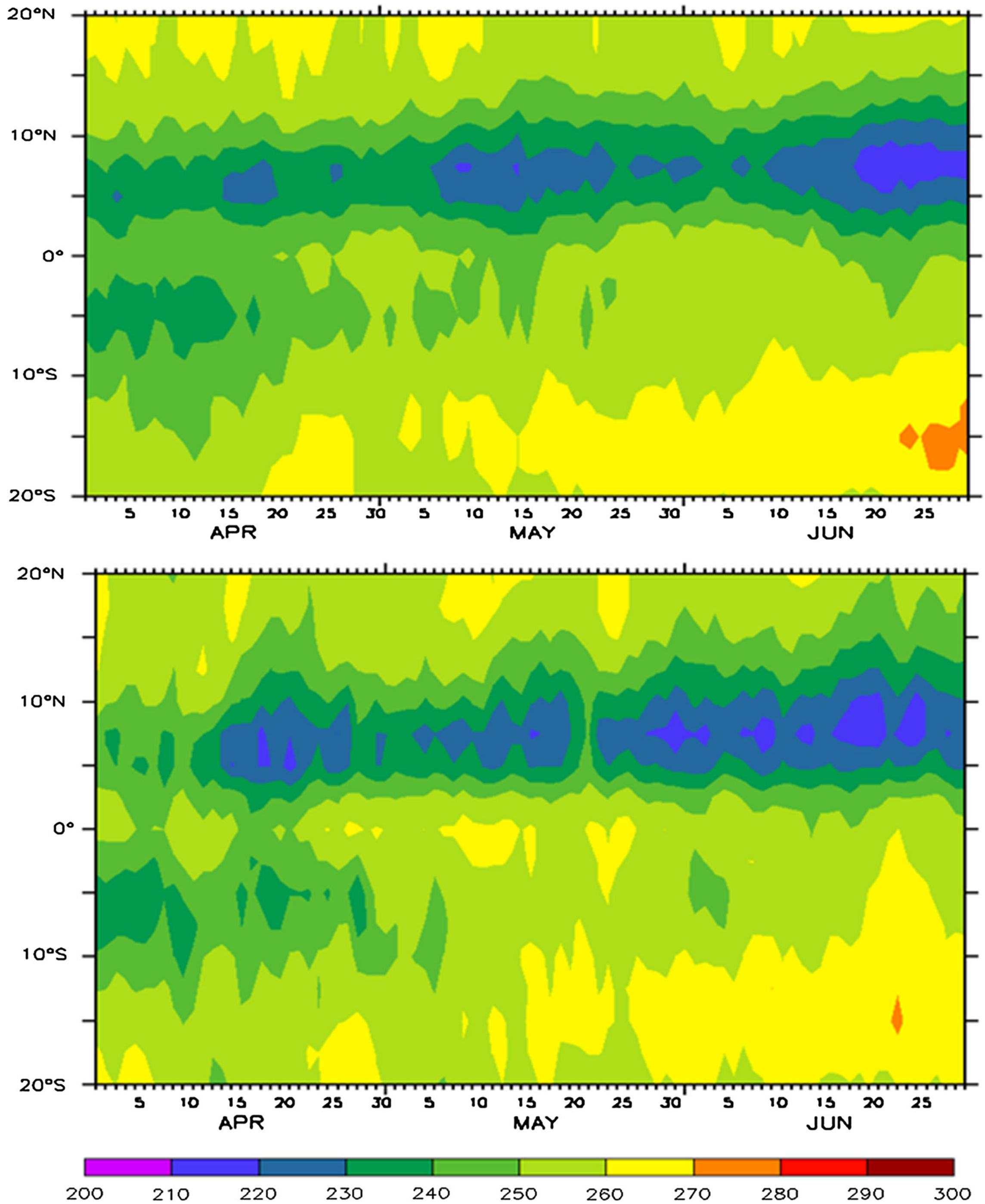

Figure 10. Hovmöller diagrams of the daily OLR (Wm^{-2}) averaged between longitudes 120° and 270°E (Pacific Ocean) with delayed (above) and early (below) MOK.

8. MOK, PMRP and convection in the Indo-Pacific ocean basin

In section 4, pentad averages of OLR over a period of 50 days prior to MOK were studied for both delayed and early MOK cases. It was shown that during the pentads-5, -4 and -3 (in between MOK and PMRP) Indian Ocean has very little convection (high OLR). It is during this period that convection shifts to the west Pacific Ocean (Joseph 1990a, b) and India comes under the downward limb of the Walker circulation. Flatau *et al.* (2001) have shown that this is the period when India gets increased occurrence of high summer temperature and high frequency of heat waves. Figure 11 gives the area averaged daily OLR

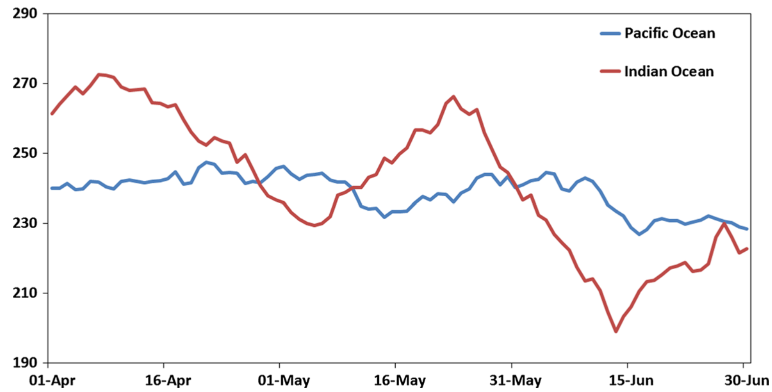

Figure 11. Area averaged daily OLR (Wm^{-2}) of Indian Ocean (blue curve, averaged over 5°S–20°N, 60°–90°E) and west Pacific Ocean (red curve, averaged over 5°–20°N, 120°–180^{0}E) for composites of delayed monsoon years for the period 01 April to 30 June.

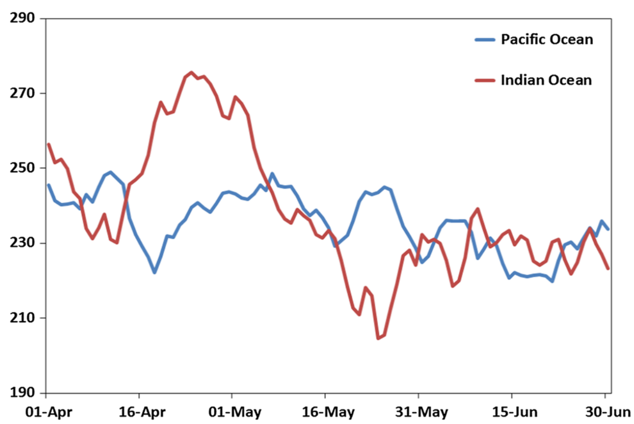

Figure 12. Area averaged daily OLR (Wm^{-2}) of Indian Ocean (blue curve, averaged over 5°S–20°N, 60°–90°E) and west Pacific Ocean (red curve, averaged over 5°–20°N, 120°–180°E) for composites of early monsoon years for the period 01 April to 30 June.

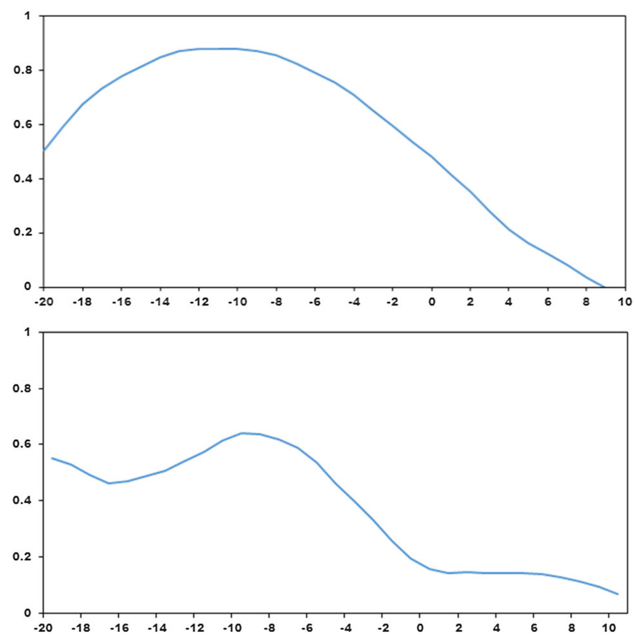

Figure 13. Lag correlation of area averaged daily OLR (Wm^{-2}) of Indian Ocean (above) (averaged over 5°S–20°N, 60°–90°E) and west Pacific Ocean (below) (5°–20°N, 120°–180°E) for composites of delayed and early monsoon years for the period 01 April to 30 June.

for delayed MOK years and figure 12 gives the same for early MOK years for the period from 01 April to 30 June. PMRP and MOK are seen in the composites for the Indian Ocean (red curve). It also shows the high OLR period in between. The OLR for these boxes in Indian and Pacific oceans have a high linear correlation at a lag of about 10 days as may be seen from figure 13. Figure 1 of Joseph *et al.* (2006) is relevant to this discussion.

9. Summary

Spatially large SST anomalies mainly in the Pacific and Indian oceans similar to those associated with El Nino (La Nina) are found to be associated with the delayed (early) occurrence of MOK in individual years. There is a statistically significant variation in the date of MOK on the decadal scale during the last 150 years. In the two decades 1900s and 1910s, MOK was in many years one or two weeks late and in the 1950s and 1960s one or two weeks early. The sea surface temperature (SST) gradient across the equator over the global oceans (Indian and Pacific) is found to have a high and statistically significant linear correlation coefficient with the date of MOK on the decadal scale which has persistence during the months prior to MOK (from January). Increased convection and rainfall occurs over the Indian Ocean around MOK and about 40 days earlier around PMRP or bogus monsoon onset and in between, for about three pentads there is a period of suppressed convection and rainfall over the Indian Ocean when increased convection occurs over the west Pacific Ocean. Using the objective DMOK data (free from bogus monsoon onset dates) we have shown that there is systematic change in the convection and rainfall over tropical Indian Ocean with respect to the date of DMOK during the prior 40-day period for composites of years of delayed, early and normal monsoon onset dates. We have also shown that DMOK is closely related to the date when the ITCZ crosses the equator to the northern hemisphere in the Indian Ocean. In years when there is warm SST anomaly south of the equator over the Indo-Pacific Ocean basin (generally associated with El Nino) persisting from January to April, the time of equator crossing of the ITCZ is delayed, leading to a delayed monsoon onset over Kerala.

By combining the available data on the date of monsoon onset over Kerala we have derived a long time series for DMOK spanning the period 1870–2014. Although for the period prior to 1971 this data has not been cleaned from bogus monsoon onsets that occur once in several years, we hope that this dataset will be useful for climate change studies till an objective method is derived for DMOK for the period prior to 1971. The DMOK series derived by Ordoñez *et al.* (2016) cannot be used to extend the DMOK series back to 1870s (see table 2).

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