AN INVESTIGATION ON AIR-SEA INTERACTION IN THE INDIAN OCEAN DURING CONTRASTING MONSOONS

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BY

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CERTIFICATE

We certify that this thesis, entitled " ***An Investigation on Air-Sea Interaction in the Indian Ocean during Contrasting Monsoons***" is an authentic record of research work carried by Charlotte B.V. under our supervision and guidance at the Department of Physical Oceanography, 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 or diploma in any University/ Institute.

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PREFACE

The customs, lives and economy of the South Asia are primarily on the distinct variability of annual cycle of wet and dry phases of monsoon. The rhythmic cyclicity of the warm and moist and cool and dry phases of the monsoon would seem to be the most conducive for agricultural societies; their very regularity makes agriculture susceptible to small changes in the annual cycle. Small variation in the timing and the quantity of rainfall has potential significance on societal consequences. A weak monsoon year generally corresponds to low crop yields and a strong monsoon usually produces abundant crops, although too much rainfall invite devastating floods. In addition to the importance of the strength of the overall monsoon in a particular year, forecasting the onset of the sub seasonal variability is of particular importance. A late or early onset of the monsoon or an ill-timed lull in the monsoon rains may have devastating effects on agriculture even if the mean annual rainfall is normal. As a result, forecasting monsoon variability on time scales ranging from weeks to years is an issue of considerable urgency.

A better understanding of air-sea interaction processes over the tropical Indian Seas on the synoptic and climatic time scales is required to study about the variability of Indian summer monsoon. The tropical oceans act as reservoir of heat and supply the necessary energy for the large-scale monsoon circulation. This region is characterised by intense air-sea exchange especially during the summer monsoon season. Studies carried out by a number of authors (Pisharoty, 1965; Mohanty et al, 1983) have established the linkage between the monsoon activity over the Indian subcontinent and the air-sea fluxes of heat and moisture over Indian Ocean.

This study is undertaken with an objective to investigate the linkage between air-sea fluxes in the Indian Ocean and monsoon forcing. Since the monsoon activity is linked to fluxes, the variability of surface marine meteorological fields under the variable monsoon conditions is also studied. To study the variability, individual years are categorised into two groups viz. deficient rainfall years (drought years) and excess rainfall years (flood years) based on the rainfall activity over the Indian subcontinent in these years. The percentage departure of all India - summer monsoon rainfall for the period 1945 to 1993 suggests that the rainfall activity has been highly variable. Excess rainfall from the normal was observed in 1956,1961,1964,1970,1973,1975,1983 and 1988. Severe drought conditions occurred in 1951,1965,1972,1979,1982 and 1987.

The data used in the study is Surface Marine Data 1994, objectively analysed fields of surface marine anomalies of sea surface temperature, sea level pressure, surface wind, fractional cloud cover, latent heat flux, short wave radiation, net heat flux and evaporation over most of the global ocean. These revised monthly mean fields are derived from individual observations in the Comprehensive Ocean-Atmosphere Data Set (COADS) from January 1945 to December 1993 and are analysed on a 1-degree by 1-degree global grid. The spatial domain of the study extends from 30° E to 120° E in the zonal direction and 30 $^{\circ}$ S to 25 $^{\circ}$ N in the meridional direction. This domain covers the south Indian Ocean, tropical Indian Ocean, Bay of Bengal and Arabian Sea.

The very objective of the present study is to document various sea surface parameters of the Indian Ocean and to examine the anomalies found in them. Hence it is attempted to relate the anomaly to the variability of monsoon over India, highlighting the occasion of contrasting monsoon periods.

The thesis comprises of Six chapters

Chapter I presents an introduction to the oceans in general and to the Indian Ocean in particular. It also briefly describes the climatology of Sea Surface Temperature (SST), Sea Level Pressure (SLP), Wind and Cloud cover over the Indian Ocean.

Chapter II explains the objective of the study, data source, methodology followed and the limitations of the data.

Chapter III deals with the anomalies of the sea surface temperature, sea level pressure, wind speed and direction and cloud anomalies for tropical deficit and surplus monsoon years and covers the following main questions.

(a) What is the nature of variability of sea surface temperature, sea level pressure, surface wind and cloud anomalies of the Indian Ocean area?

(b) Do these variability coincide with the occurrence of EQUINOO and dipole events of Indian Ocean and ENSO activities

(c) How do they feed back on the variability of monsoon over India?

Chapter IV investigates the anomalies of the Latent heat, Net heat, Short wave radiation, Evaporation and Heat budget anomalies of the Indian Ocean for the years mentioned in chapter III. An attempt is also made to investigate whether the anomalies can explain the occurrence of good and bad monsoons of India.

Prior to the description of the variability of the anomaly field, the actual fields are presented to get an overall picture and to understand the monthly evolution of different parameters.

Chapter V depicts the relation between Indian Summer Monsoons and the occurrences of Indian Ocean Dipole, Equatorial Indian Ocean Oscillation and ENSO related features.

Chapter VI summarises the results and portrays the conclusions of the work.

The references cited are furnished at the end of the thesis.

Contents

LIST OF FIGURES AND TABLES

CHAPTER I

GENERAL INTRODUCTION

1.1 Introduction

Ocean is considered as the flywheel of climate system. Because of the huge volume and very large heat capacity of oceans, they act as a source of supply of heat and moisture to the atmosphere. Almost three quarter of the total solar energy reaching, the oceans absorb the bottom of the atmosphere. Due to the large thermal inertia of the oceans, there exists a lag between the absorption and the subsequent release of energy back to atmosphere. This can influence the seasonal and temporal variations of the climate. The absorbed heat is redistributed spatially by different dynamical processes by the oceans, ultimately modifying the climate and causing its variability.

The ocean and the atmosphere act as a coupled system in a very complex manner. On a large scale, such as a continent surrounded by oceans, heat build up on land over time will result in low-density air masses, or areas of low pressure. Conversely denser air associated with high pressure dominates ocean surfaces. Wind and ocean currents that result from air flowing from high to low pressure mix areas of warmer and colder air and water contributing to the global energy balance. This exchange is evident at different levels of atmosphere. Air converging to a low - pressure center at the surface rises, leading to moisture condensation and the subsequent release of heat into the upper atmosphere.

Diverging air at the surface in a high-pressure center is associated with subsiding air from the upper atmosphere and evaporation, a mechanism for energy storage

Just as energy imbalances develop between land and water surfaces, the variations in space and time of solar heating due to earth's tilt create seasonal hemispherical imbalances. The hemisphere receiving the more direct radiation (during summer months) experiences a net radioactive heating and the winter hemisphere simultaneously experiences a net radioactive cooling. The south west monsoon is a result of such complex interaction between two fluids, the mobile atmosphere and the slowly changing ocean below with a linkage of winter and summer hemispheres by strong winds and ocean currents across the equator. The inter-hemispheric temperature contrast and the consequent wind are enhanced by the difference in heat storage between the dry land and the oceans. This leads to the warming of the landmass relative to ocean in summer and cooling of landmass relative to ocean in winter.

There are various scales of motion taking place in the atmosphere and ocean. They are micro-scale (turbulent motion) with periods of few minutes and length up to 100 m; convective scale (pronounced vertical motion) with periods ranging from few minutes to one hour and length ranging from 100 m to 10 km; meso-scale with periods of several hours and length up to 100 km; synoptic scale (cyclonic and anti cyclonic vortices) with periods of few days and lengths up to 1000 km; and large scale (planetary waves, quasi stationary circulation etc) with periods ranging from few weeks to decades and length ranging from 1000 m to the entire Ocean.

Since this thesis is mainly concerned with the study of air-sea interaction processes of the Indian Ocean a brief description of the study area is presented below.

1.1 Oceanography of Indian Ocean

Various Oceanographic expeditions have taken place in Indian Ocean area before 1960. *Challenger* (1872-76), German expedition on board *Valdivia* (1898-99) and *Gauss (*1901-1903), Dutch expedition on board Siboga (1899-1900), Weber *Snellius* (1929-1930), British on board HMS *Investigator1* (1881-1900) H.M.S *InvestigatorII* (1908-1947) *Johh Murray* expedition (1933-1934),Sweedish Deep sea expeditions *Albatross* are some among them. In spite of all these expedition Indian Ocean prior to 1960 was the least known oceanic area. Recognizing this UNESCO and SCOR embarked on a great expedition known as *International Indian Ocean Expedition* during 1960-1965. The *Indian Ocean Expedition (INDEX)* (1975-1976) *Monsoon Experiment (MONEX*) (1978-1979) made extensive measurement of oceanic and atmospheric parameters of the Indian Ocean .

The Indian Ocean is the third biggest ocean of the world (Fig:1.1). The annual heating cycle and the presence of vast Asian land -mass, with mighty Himalayan and Tibetan plateau in the north, bordered to the west by East African Highland make Indian Ocean a unique ocean. The Indian peninsula divides the northern part of Indian Ocean into two great gulfs; the Bay of Bengal and the Arabian Sea that are extended to the north by the Gulf of Oman and the Persian

Gulf and to the west by Gulf of Aden and the Red Sea. The Arabian Sea and the Bay of Bengal push the geographic domain to 30°N. Oceanographically the southern limit of Indian Ocean can be placed in the region of subtropical convergence situated at 40° S latitude (Svedrup et al 1942). The total area of the Ocean, including the adjacent seas is calculated to be about $49x10^2$square Kilometre.

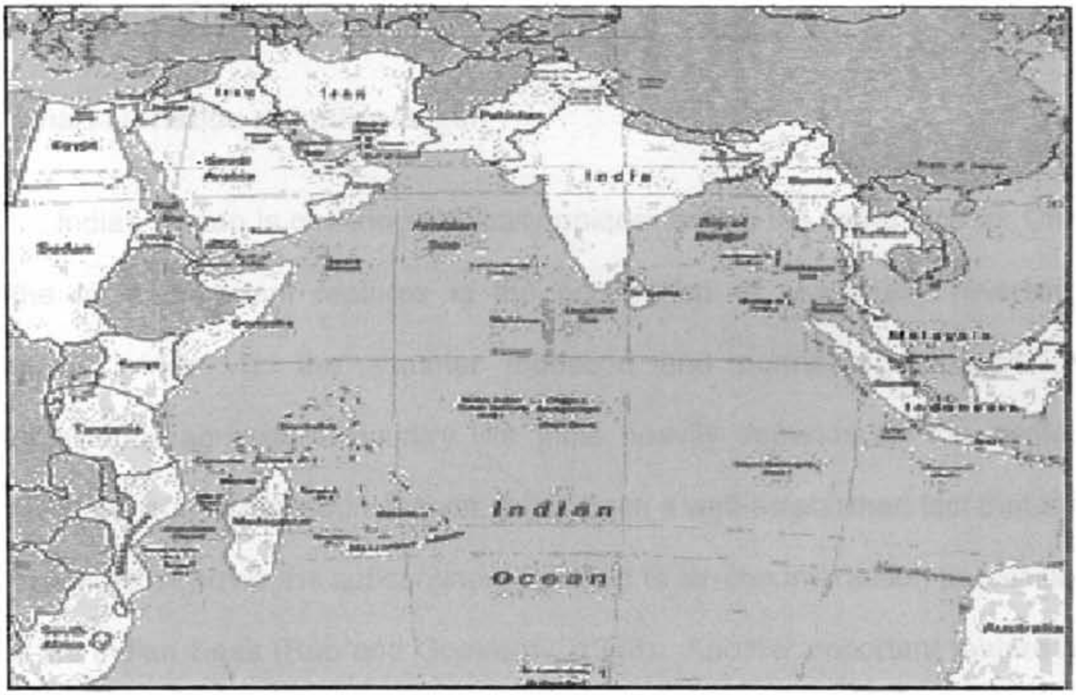


Fig: 1.1 Indian Ocean.

With the advent of remote sensing techniques, collection of data on various parameters over large area become possible. However, lack of observation over the remote oceanic region has been a major draw back in studying the link between air sea interaction over the tropical Indian Ocean and Indian summer monsoon activity. Indian Ocean, compared to other oceans is an inadequately studied and sampled area mainly due to the infancy of real time observing networks in the ocean and the atmosphere in this region and the subsequent shortness of the records. One reason that can be attributed to this situation is most of the countries around Indian Ocean have limited economies and hence less government support for oceanic observation and research. With the advent of satellites, the entire oceanic region of tropical Indian Ocean has been fairly well sampled. The data sets compiled by NOAA, NCDC and NCAR have provided a good database for the study of the ocean atmospheric interaction in the global Ocean.

1.2 Characteristics of Indian Ocean

Indian Ocean is oceanographically unique among the world Ocean. One of the most important features is the occurrence of seasonally reversing monsoon winds viz. the summer monsoon and northeast monsoon. A predominantly agricultural country like India heavily depends on the rainfall received during the monsoon season. It has been a well-established fact that the monsoon activity over the subcontinent is linked to air-sea interaction processes over the Indian Seas (Rao and Goswamy, 1988). Another important feature is the occurrence of warm pool of water in the southeastern Arabian Sea prior to the onset of summer monsoon. Recently another feature called Indian Ocean dipole has been discovered which has got considerable impact on the air-sea interaction process and the monsoon activity (Saji.et al 1999). The circulation of the tropical Indian Seas exhibits seasonal variability in response to the summer monsoon. These circulation features play an important role in governing the dynamics of the tropical Indian Ocean. Added to this circulation feature is the

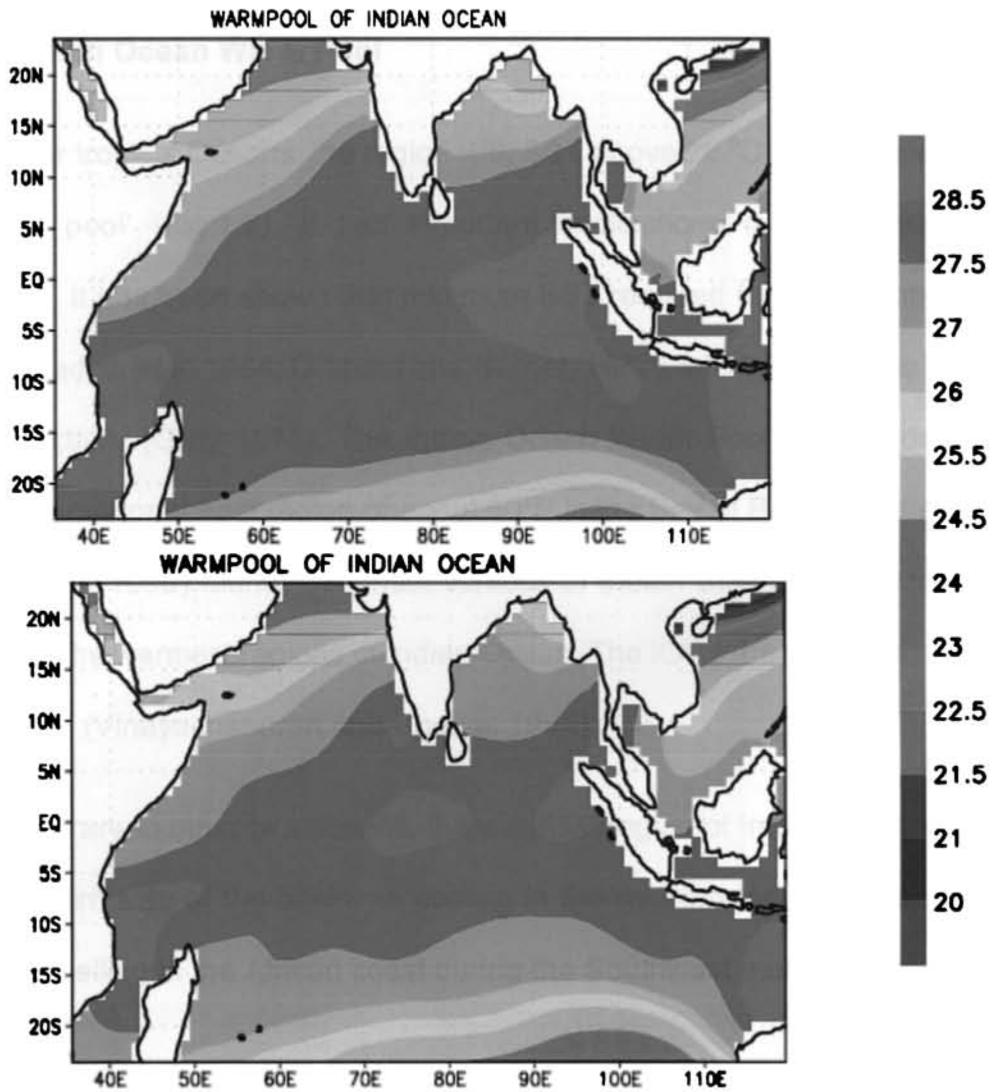


Fig: 1.2 Sea Surface Temperature in the Indian Ocean (a) July (b) August.

Indonesian throughflow (Wyrtki 1987), a flow between the western Pacific and the Eastern Indian Ocean. A detailed description of these features is presented in the following sections.

1.2.1 Indian Ocean Warm Pool

Over tropical Oceans, the region with SST above 28 °C is often referred to as 'warm pool' (Fig:1.2). It has important implications in the atmospheric processes. It has been shown that minimum SST required for active convection is 28°C (Gadgil. et al 1984; Graham and Barnet, 1987), which is prone to tropical cyclonic activity (Gray 1975). The Indian Ocean Warm Pool (IOWP) develops along the equatorial east region of about 50°E in the Bay of Bengal and Arabian Sea. Joseph (1990) found that onset vortices of Indian Summer Monsoon take place over the warmest regions of Indian Ocean. The IOWP has an area of about 2.8x 10 km^2 (Vinayachandran and Shetye, 1991).

Contrary to other two oceans, the warmest region of Indian Ocean occurs in the eastern side of the basin. A cooling in the western Indian Ocean occurs due to up welling in the African coast during the Southwest monsoon season.

1.2.2 INDIAN OCEAN DIPOLE (IOD)

Internal modes of variability that lead to climatic oscillations have been recognized in Indian Ocean recently. As in the case of tropical Pacific Ocean and Atlantic Ocean, ocean atmosphere interaction leading to inter-annual climate variability is noticed in Indian Ocean also. An analysis of observational data over

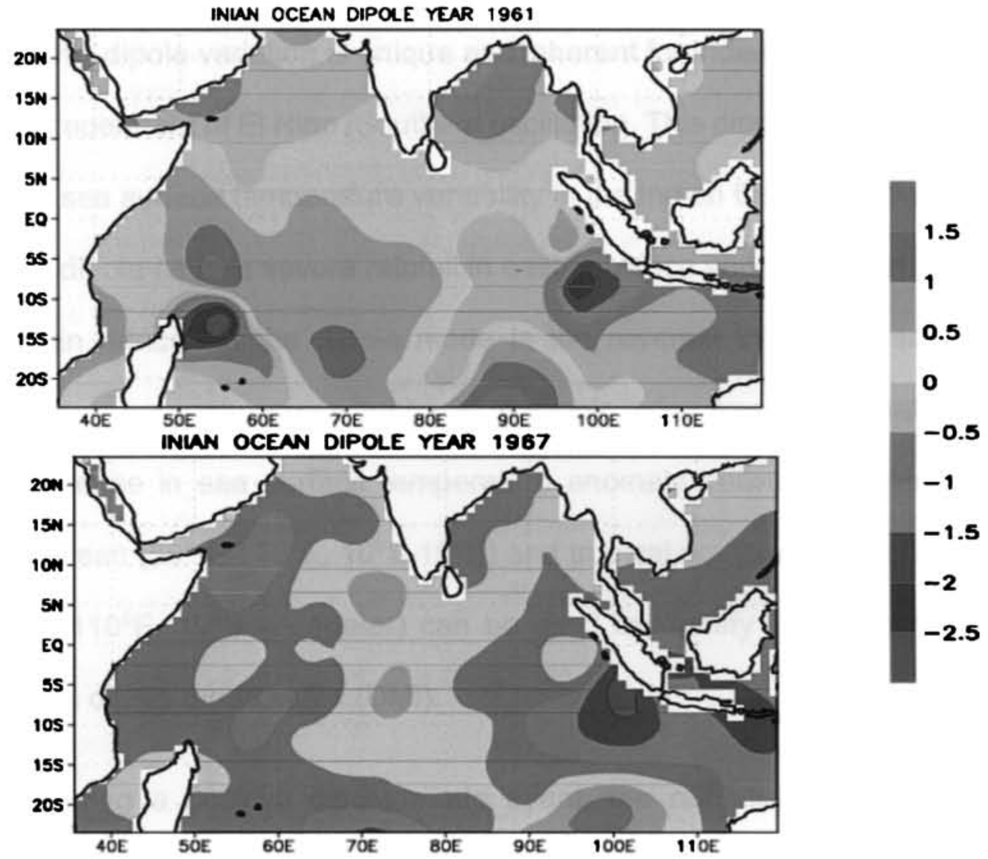


Fig: 1.3 SST anomaly (ºC) in the Indian Ocean during (a) October 1961 (b) October 1967.

the past 40 years, shows a dipole mode in Indian Ocean (Fig:1.3) a pattern of anomalously low sea surface temperature off Sumatra and high sea surface temperatures in the western Indian Ocean, with accompanying wind and precipitation anomalies (Saji et al 1999).

This dipole variation is unique and inherent for Indian Ocean and is found to be independent of El Nino /Southern oscillation. This dipole mode can account for 12% sea surface temperature variability in the Indian Ocean. Active period of positive dipole causes severe rainfall in eastern Africa and drought in Indonesia. The main feature of the dipole mode is the reversal in sign of sea surface temperature anomaly across the basin. A simple index time series that describes the difference in sea surface temperature anomaly between tropical western Indian ocean (50° E– 70°E, 10°S-10°N) and tropical southeastern Indian Ocean (90°E – 110°E, 10°S – equator) can be used to identify dipole mode. This is known as dipole mode index (DMI).

During a positive dipole mode event, the rainfall decreases over the oceanic tropical convergence zone (OTCZ) and increases over the western tropical Indian Ocean. This rainfall is dynamically consistent with the divergence /convergence shift associated with the wind field. The evolutions of dipole mode (DM) events are examined for the period of 1975 – 1998. Positive events are recognized during 1997,1994,and 1982 and negative dipole mode (DM) events during 1996,1984-1985 years. In its positive phase of DM is characterized by warmer than normal SST in western Indian Ocean and cooler than normal SST in eastern Indian Ocean (Vinayachandran et.al 2002).

The DM is phase locked to the seasonal cycle in the Indian Ocean and the peak strength of anomalies generally occurs during boreal fall. The strongest of the known positive DM event occurred in 1994 and1997, coinciding with most intense ENSO of the recent decades. The SST in the eastern Indian Ocean was cooler by more than 2°C during these events (Yu and Reinecker 1999; Webster et al, 1999). The east west contrast in SST was in turn maintained by the wind anomalies for a longer period by coupled ocean atmosphere interaction (Saji et al 1999, Webster et al; 1999). The wind anomalies alter the thermal structure of the equatorial Indian Ocean from its normal conditions. The thermocline in eastern Indian Ocean becomes shallower than usual in the east and deeper in the west. Kelvin waves produced by anomalous easterlies, lifts the thermocline in the eastern Indian Ocean along the shore. Winds off the Sumatra enhance the cooling in the eastern Indian Ocean (Murtugudde et al 2000). This is manifested in the form of a fall in sea level in the eastern Indian Ocean and a rise in western Indian Ocean.

Kelvin and Rossby wave propagations play an important role during the DM event. The wind anomalies excite equatorial Kelvin waves that reflect from the eastern boundary as long Rossby waves that propagate westward. The easterly wind anomalies during 1994 and 1997 (Murtugudde et al 2000) excited an upwelling Kelvin wave in the equatorial Indian Ocean. This Kelvin wave was reflected as an upwelling Rossby wave.

1.2.3 Equatorial Indian Ocean Oscillation (EQUINOO)

The IOD events generally become prominent only after the southwest monsoon. It has been found that over the equatorial Indian Ocean, enhancement of deep convection in the atmosphere over the western part is generally associated with suppression over the eastern part and vice versa during southwest monsoon (Gadgil et al, 2003). This oscillation is also reflected in the pressure gradient and winds along the equator and hence it is named as Equatorial Indian Ocean Oscillation.(EQUINOO). Gadgil et al (2003) suggested that EQUINOO has strong impact on monsoon and the Indian Summer Monsoon Rainfall (ISMR). The ISMR anomlies can be explained by favourable (unfavourable) phases of EQUINOO or ENSO or both. Analysis of the zonal wind anomaly along the equator showed the signal of unfavourable EQUINOO as early by mid April and hence could be used as an indicator of the following ISMR along with ENSO events (Gadgil et al 2003).

1.2.4 Indonesian Through Flow

The horizontal pressure gradient, which is to be maintained between the two sides of Indonesian Archipelago, governs a flow between the western Pacific and the Eastern Indian Ocean (Wyrtki 1987). The upper part of this throughflow is an element of conveyor belt, which links Pacific Ocean and Indian Ocean in the tropical latitudes (Gordon 1986). The Pacific water enters the Indonesian Archipelago following two main routes. (1) The western root enters the Celebes Sea and the Makassar basin. One part of it ends in Indian Ocean through the

Lombok strait while the other part turns through the Flores sea to reach the Banda Sea (2) The eastern route through the Halmahera and Molucca sea, reaches the Banda sea from the north. From the Banda Sea there are two passages to the Indian Ocean. The Ombai strait, between Alor and Timor Islands and the Timor passage between Roti Islands and the Australian continental shelf.

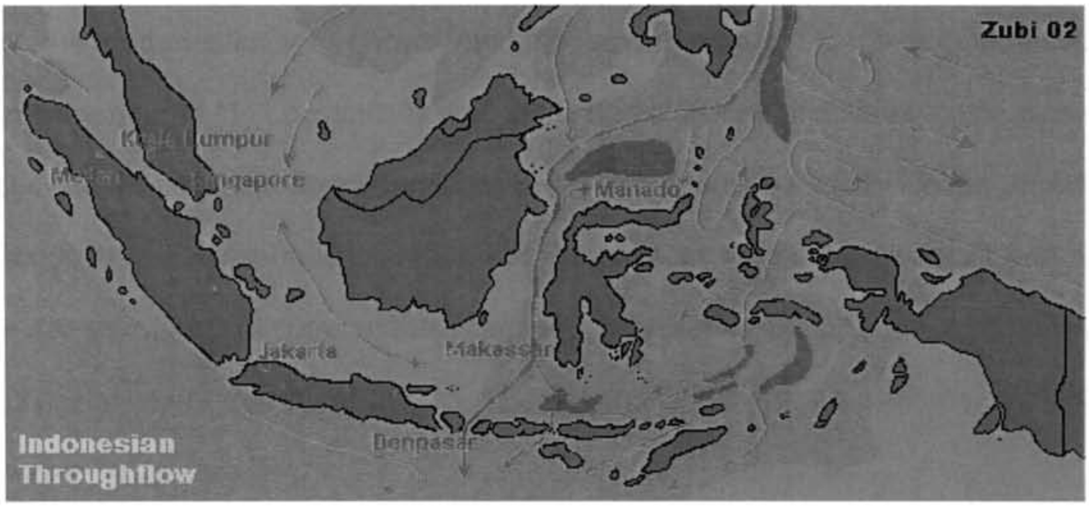


Fig: 1.4 Schematic view of Indonesian throughflow into the Indian Ocean

The Lombok strait corresponds to most direct path is 350m deep and involves the upper layers. The other straits allow deeper fluxes and involve water with a longer residence time in the Indonesian Archipelago. The relative contribution to the through flow of Timor passage is significant (Java Australian Dynamic Experiment JADE '92).

The physical structure of the Pacific and Indian Ocean is substantially affected by inter – ocean transport of excess fresh water from the Northern Pacific Ocean through the Indonesian Seas. The efficiency of this transport is an important regulator of meridional overturning of the oceans and hence perhaps to the global thermo-haline circulations (Gordon & Rana, 1996).

The seepage of warm water out of the Pacific affects the volume of western Pacific warm pool and thus may influence El Nino events. The source, pathways and physical properties of Indonesian Through Flow are not well characterised even today. The salinity, temperature and chemical character data from the Inodonesian seas shows that Indonesian Through Flow is dominated by two components (1) Low salinity well ventilated North Pacific water through the upper thermocline.of Makassar Strait. (2) More saline South Pacific water through the lower thermocline of the east Indonesian seas. Monsoon variation in the range of these components, perhaps modulated by El Nino may exert potentially important feedback to the ocean circulation as well as climate systems.

The warm low salinity Pacific water weaves through Indonesian seas to the eastern boundary of the Indian Ocean. The Indonesian Through Flow water adds fresh water into Indian Ocean as it spreads by advection and diffusion within the Indian Ocean's south equatorial current (SEC). The low salinity Through Flow trace, centred along 12° S stretches across Indian Ocean, separating the monsoon-dominated region of the northern Indian Ocean from the more typical stratification to the south (Gordon et al 1997).

Water mass analysis indicates that the most penetrating root followed by Pacific water occurs within Makassar Strait (Gani Llahude & Gordon 1996).

There exist a lot of uncertainties relating to the magnitude of the flow of deep water into the Indian Ocean associated with the upwelling and the magnitude of the heat gained (lost) by the Indian Ocean from (to) the atmosphere. As the deep inflow increases, the heat gained by the Indian Ocean from the atmosphere should increase to allow cold water to be converted to warm water. As the Through Flow increases more heat should be lost by Indian Ocean to the atmosphere since the Through Flow transports warm water into the Indian Ocean (Banks Helene 2000).

As a consequence of large variation in the estimates of the Through Flow and deep inflow, hydrographic estimates of the amount of heat transferred to the Indian Ocean from the atmosphere vary dramatically between .25 (Fu, 1986) and 1.0PW (Tool and Waren 1993) and between 0.5 (Hasternath and Lamb 1979), and 1.4PW (Hsiung Gartenicht and Schott 1997).

1.3 Interannual variability of monsoon ENSO relationship

Statistical analysis shows that monsoon and coupled ocean atmospheric system undergo interannual variation in three to seven years' time scale. Anamolously warm Pacific Ocean is followed by a weak Indian rainfall in the subsequent years. The correlation between All India rainfall index and southern oscillation index is found to be –5, whereas that with SST and monsoon is –6.4.

Shukla and Paolino (1983) established a very significant relationship between drought and ENSO. Table 1.1 summarise the relationship between Indian summer monsoon and the state of ENSO in equatorial Pacific Ocean.

From it most of the El Nino years are followed by a drought in India. But all Indian drought years were not El Nino years. Out of total 22 El Nino years between 1870 and 1991, only two were seen to have above average rainfall in India. On the other hand, La Nina years are followed by abundant rainfall in India. Only two La Nina years fail to give excess rainfall in Indian region. But all wet years in India are not affiliated with La Nina years. Though the relationship is not very convincing the state of ENSO is related in some fundamental manner with Indian rainfall.

Relationship between the all India summer monsoon rainfalls and the state of ENSO in equatorial Pacific Ocean

All India summer rainfall

Rainfall	Total	El Nino	La Nina
Below average	53	24	2
Above average	71	4	19
Deficient	22	11	2
Heavy	18	0	7

Table: 1.1

The All –India rainfall index analysis (1871-1994) is an extension of Shukla and Paolina (1983). Deficient rainfall is defined as a season in which precipitation was at least a standard deviation less than the mean. Heavy rain refers to a season in which the rainfall is greater than the mean plus a standard deviation.

1.3.1 Process involved in monsoon coupled system

To understand the driving mechanism of planetary scale monsoon, it is essential to know how the atmosphere, land and ocean system communicate. These communications happen through interfacial fluxes of heat, moisture and momentum between the ocean, atmosphere and the land and loss of heat to outer space.

1.3. 2 What controls the SST in the warm pools adjacent to monsoon region?

The warm pool of Indian Ocean in south Asia and North Australia remains between 27°c to 30°c almost throughout the year. The warm pools are integral component of El Nino southern oscillation (ENSO). They directly affects the interannual variability and total mean of monsoon rainfall in India Australia and Africa (Webster 1987b) Small changes in the warm pool have been shown to have major consequences in global climate (Palmer and Mansfield 1984,1986).

The warm pool of western Pacific is of prime importance. Then comes the turn of Indian Ocean warm pool region. The inter annual variability of Indian Ocean is in phase with that of Pacific Ocean. This oceanic region affects the inter-annual variability of monsoon is North Western Pacific region. Many authors have shown that the SST anomalies in Western and North Western Pacific have

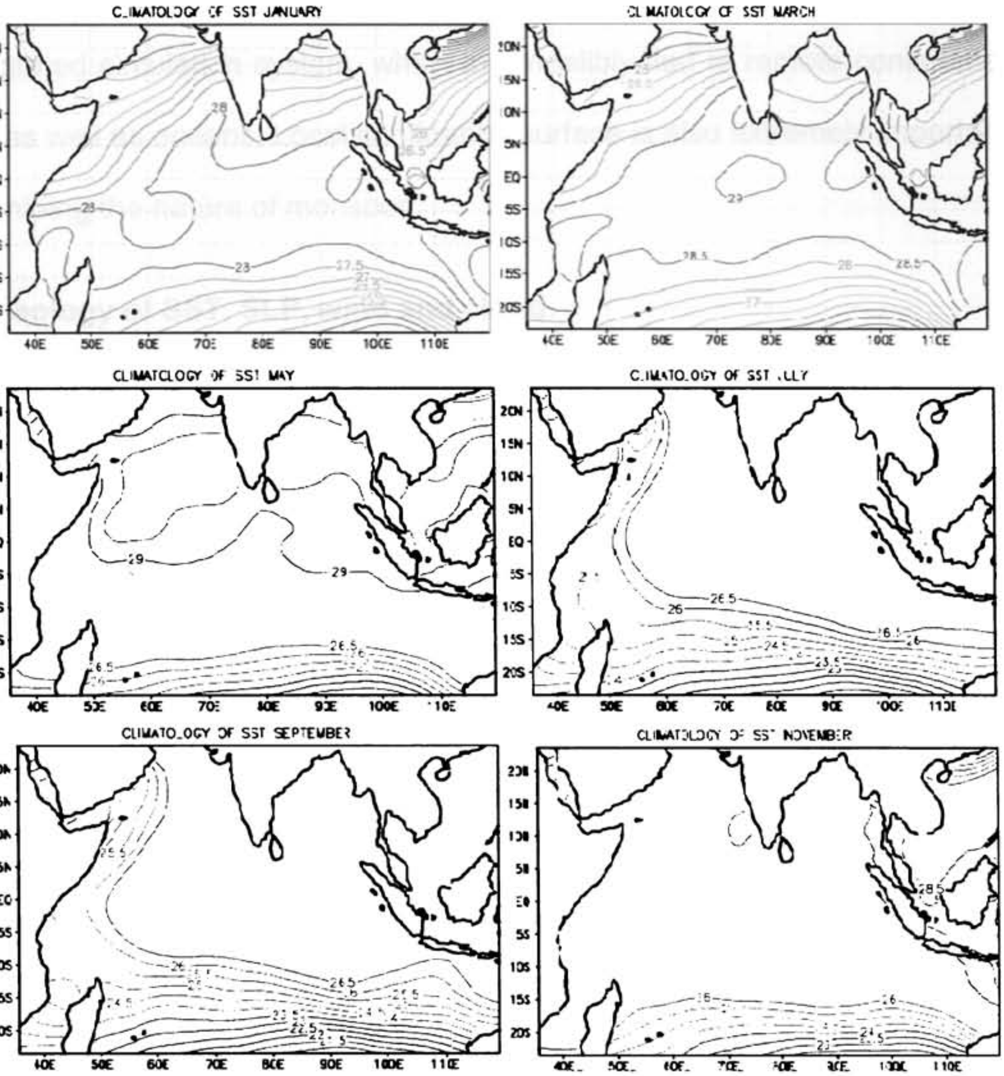


Fig: 1.5 Monthly mean Sea Surface Temperature (ºC) in the Indian Ocean direct effect on the Asian monsoon (Ju and Slingo, 1995). The Indian Ocean through flow is an integral part of heat balance of Indian Ocean and Pacific Ocean. They are modulated by ENSO, the export between the two basins appears to be oceanic teleconnection on the inter-annual time scales. Monsoons are integrated circulation system, which are indelibly tied to remote continental regions as well as oceans. Local continental surface is also extremely important in determining the nature of monsoon.

1.4 Climaology of SST, SLP, wind and cloud

1.4.1 Climatology of Surface temperature distribution of Indian Ocean

SST over major part of Bay of Bengal and Indian Ocean east of 75°E shows a significant increasing trend during the post monsoon season (October – December). The warmest sea surface temperature of 28°C and above is noticed mostly in the south of the equator in January. This region extends upto 11°S to 17°S (Fig.1.5). During February – April, the warm region continues to expand towards south as well as north of the equator. In February, we have more of a southward expansion of warm region. In March and April, the northward expansion of the warm region is more pronounced. By April it covers a large region from 18°S – 18°N of the Indian Ocean and most of the region around equator displays a SST of 29°C or more.

May experiences an enhanced expansion of warmer SST over the North Indian Ocean. We have an SST of 29°C or above from 2°S upto even 18°N at this time. Both Bay of Bengal and Arabian Sea experience this warming. With the onset of monsoon in June, the northwest Indian Ocean, particularly the Arabian

Sea experiences a rapid cooling of SST. It continues till September. By August, the warm region contract to the central and east Indian Ocean and Bay of Bengal. By September, the warmest part of Indian ocean of 28°C or above lies in the Bay of Bengal and equatorial region east of 58° E which coincides with the precipitation zone as derived by Krishnamurthy et al (2001). During other seasons and during the months from June to September, there are some pockets in Bay of Bengal, Arabian Sea and Indian Ocean where increasing trends of SST is noticed.

With the reversal of wind, during the end of southeast monsoon, October witnesses a seasonal change in SST of the Indian Ocean. This results in the shifting of warm SST region more to the south of the equator. This southward expansion continues through November and December and is concentrated over the Indonesian region. The warm SST in the eastern Indian Ocean region during October and its enhancement through November and December demonstrates the significance in establishing the Indian Ocean SST mode and the monsoon ENSO link and the variability of Indian Ocean with respect to ENSO variability.

A remarkable feature of SST in the northern Indian Ocean is that SST maximum does not occur during the summer but during the spring transition from northeast to Southwest monsoon. SST values in May are above 28°C everywhere in the north of the equator and north of 10°S in the East. Summer SST values vary from 25°Cto 27°C.

The Sea Surface Temperature of Indian Ocean (30°N - 40°S) varies between a minimum of 12°C in the winter at subtropical convergence and maximum of about 32°C in the northern part of Arabian Sea. In the Persian Gulf, temperature goes up to 35°C is possible. Between the subtropical convergence and the Tropic of Capricon a seasonal variation of temperature of the order 5°C to 7°C is encountered. But it is only 1°C or 2°C near 10°S. Alternating monsoon winds and currents regulate the Sea Surface Temperature of the region 10° S to the Tropic of Cancer.

During April and May, the thermal equator shifts towards 10°N and everywhere between 10°S and the Tropic of Cancer, surface temperature is high (>28°c). During June, August under the influence of South West monsoon the thermal equator descends southwards and lie near 3°S or 4°S. North of 10°S the water temperature is higher than 25°C except for the belt along the Somali coast and the south Arabian Sea. Due to the strong upwelling along the Somali coast the temperature fall from 26°C offshore to around 15°-18°C along the coast (Swallons et al 1968). The Red Sea and the Persian Gulf attain their maximum surface temperature of 30°C in August. In November everywhere north of 20°S the surface temperature are higher than 25°C and the thermal equator slightly south of geographic equator. Among the world oceans, Indian Ocean has the highest mean annual surface temperature of 25.3°C.

The salient features of the vertical thermal structure are the mixed layer, the upper thermocline (in part will be seasonal), the lower thermocline, the main oceanic thermocline and the deep layer (where the temperature decreases very slowly with depth).

In the tropical and subtropical regions the depth of the mixed layer varies between 40m and 100m approximately. But during upwelling season, along the coast of Arabia, and Somalia the mixed layer is shallower than 20m. It goes deeper than 100m during the SW monsoon in the Arabian Sea and off Sumatra.

The recent GCM experiments by Krishnan et al (2003) showed that tropical Indian Ocean SST anomalies could exert considerable influence on the seasonal mean monsoon rainfall, by producing persistent changes in the regional convection. Thus, on the one hand, the Indian Ocean dynamics is regulated by the seasonal monsoon cycle, while on the other; the monsoon circulations and the associated rainfall distributions are influenced by the temperature variation in the tropical Indian Ocean. Recent studies also raised the possibility of ocean - atmosphere coupled phenomena in the tropical Indian Ocean which is analogous to El Nino / Southern Oscillation in Pacific Ocean (Webster et al 1999, Saji et al 1999,Murtugudde et al 2000).

The warmest water of 31° C is found in Laccadive High, prior to the onset of monsoon and the sea surface height is also very high there due to the presence of Rossby wave in this region during December and February. The monsoon vortex is believed to have its genesis from this warm pool found in Laccadive High region. The low salinity and high temperature of the warm pool will provide good input into the monsoon rain. Though the Indian Ocean SST anomaly is very less (almost within 1°C.) its impact on the climate is very high compared to the Pacific Ocean.

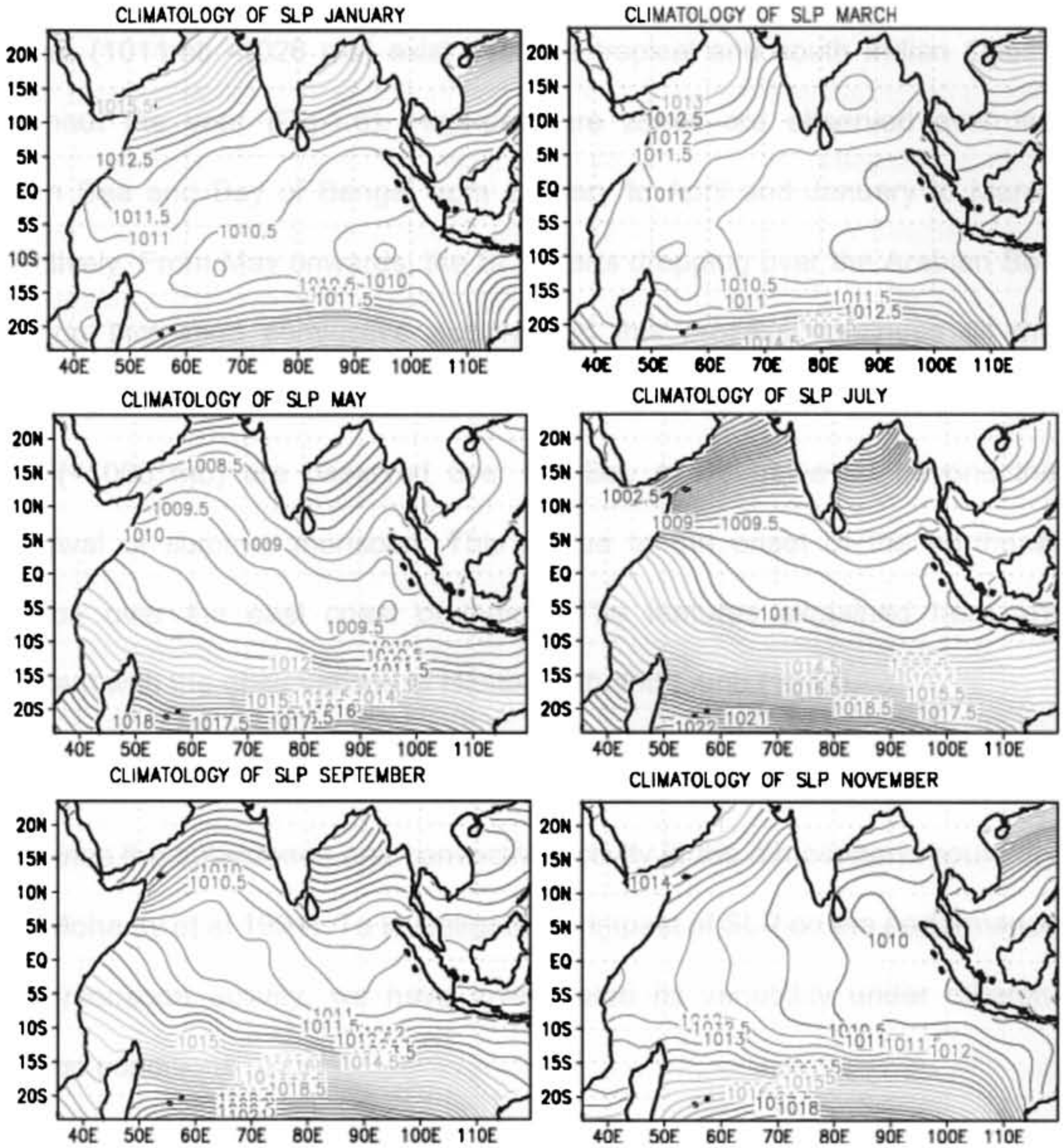


Fig: 1.6 Monthly mean sea level pressure in (hPa) in the Indian Ocean

1.4.2 Climatology of the Sea Level Pressure of Indian Ocean

The monthly climatology of SLP variability indicates that the zones of high pressure (1011mb -1026 mb) exist over the tropical and south Indian Ocean throughout the year (Fig.1.6). High-pressure zones are observed over the Arabian Sea and Bay of Bengal from January to April and January to March respectively. From May onwards, the SLP starts dropping over the Arabian Sea indicating increased convective activity and this process continues till the withdrawal of the summer monsoon in September. On the contrary, low-pressure zones (<1008 mb) are observed over the Bay of Bengal even beyond the withdrawal of summer monsoon. This is due to the onset of the northeast monsoon over the east coast of India. The features explained here are consistent with the observations of Hasternrath and Lamb (1979).

It has been an established fact that variability of sea level pressure (SLP) determines the subsidence and convective activity in the atmospheric boundary layer (Mohanty et al 1994). To investigate the impact of SLP on the performance of the monsoon activity, we have investigated its variability under different monsoon conditions.

The air-sea interaction processes over the tropical Indian ocean show that sea level pressure in the Arabian sea during the pre-monsoon (April and May) and monsoon (June –September) months play a significant role in deciding the fate of ensuing summer monsoon activity over the Indian subcontinent. Sea level pressure (SLP) shows a decreasing tendency from the month of April onwards over the Arabian Sea. A maximum difference in SLP of (upto 0.5mb) is observed in the northern Arabian Sea and northwestern Bay of Bengal in September. This large pressure gradient between the northern Arabian Sea and the southern Indian Ocean is conducive for the strong cross - equatorial flow during an excess monsoon season. On the other hand, the small or weak pressure gradient between the Arabian Sea and south Indian Ocean results in a weaker cross equatorial flow and deficit monsoon (Ramesh Kumar, 2000).

1.4.3 Climatology of Wind vector variation in Indian Ocean

There is a belt of westerlies and easterlies centred around 8°N and 8°S respectively, right across the Indian Ocean during the northern summer (June - August), The winds are reversed during winter (December – February) (Fig.1.7). The zonal wind pattern during the June – August period represents the monsoon wind which has a strong easterly component in the southern hemisphere and changes direction from south easterly to south westerly when it approaches the Indian subcontinent. The Arabian Sea is warmer during this period associated with higher sea level.

The low SST and the low sea level along the southwest coast of India indicate upwelling while the coastal waters of the east coast are abnormally warm with positive Sea Surface Height anomaly, a characteristic normally associated with sinking. The wind pattern over the tropical Indian Ocean during winter is analogous to the other oceans, with North East and South East Trade Winds on either side of the equator. The Equatorial current system of the Indian Ocean appears to be displaced to the south during this season. The SST of

Arabian Sea and Bay of Bengal decreases when winter sets in (November - February) along with a warming at the southern ocean (southern summer) (Muraleedharan et al., 2000).

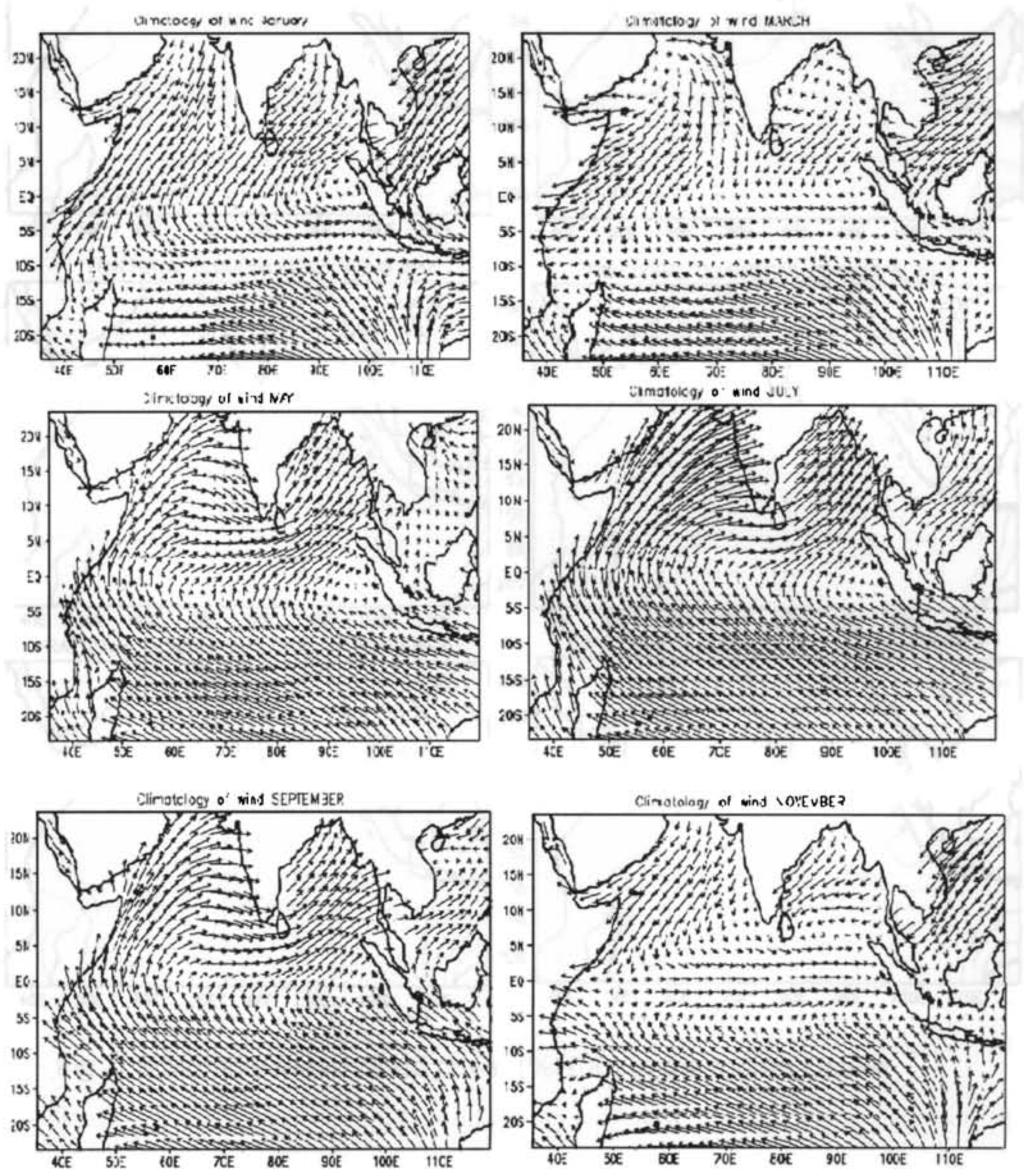


Fig: 1.7 Monthly mean surface wind of alternate months in the Indian Ocean

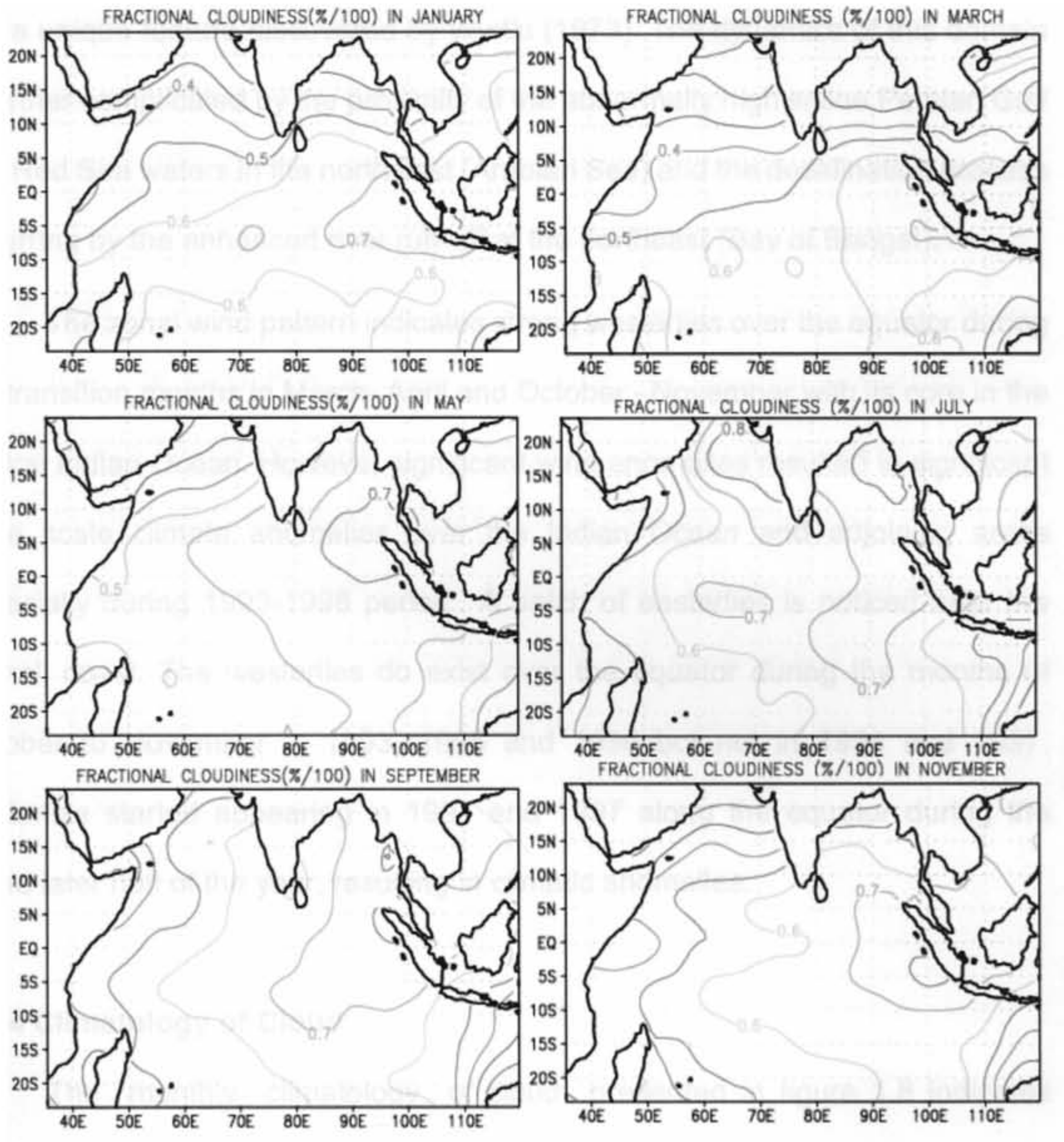


Fig: 1.8 Monthly mean cloud cover (%/100) in the Indian Ocean

The frequent reversal of wind (Monsoon) paves way for the formation of monsoon drift current during the monsoon season and Inter-monsoon Equatorial Jet, a unique feature discovered by Wyrtki (1973). The dynamics of this domain is further complicated by the proximity of the abnormally high saline Persian Gulf and Red Sea waters in the northwest (Arabian Sea) and the desalination process occurring by the enhanced river run-off at the northeast (Bay of Bengal).

The zonal wind pattern indicates strong westerlies over the equator during the transition months in March- April and October –November with its core in the central Indian Ocean. However significant wind anomalies resulted in significant large scale climate anomalies over the Indian Ocean and adjoining areas especially during 1993-1998 period.. A patch of easterlies is noticed near the Somali coast. The westerlies do exist over the equator during the months of October to November in 1993, 1995 and 1996 but not in 1994 and 1997. Easterlies started appearing in 1994 and 1997 along the equator during the entire later half of the year, resulting in climatic anomalies.

1.4.4 Climatology of Cloud

The monthly climatology of cloud presented in figure 1.8 indicates strong seasonal variability. It is observed that skies are cloudy with cloud cover varying between 55% and 75% over the south Indian Ocean during most of the months suggesting increased convective activity. However, the cloud cover is scanty in the Arabian Sea (< 30 %) during the northern winter (Jan- May). The picture is different in the southern Bay, where we can observe partially overcast sky conditions (>50 %) caused due to increased air-sea interaction by the northeast monsoon winds. The pre-monsoon months (April and May) are characterized by overcast skies over the Bay of Bengal, mainly due to the conducive conditions for the formation of tropical cyclones. With the commencement of summer monsoon in June, the intense convective activity leads to overcast skies (>60 –80 %) over the Arabian Sea and Bay of Bengal. This feature is observed until the withdrawal of the summer monsoon during September. The skies clear up over the Arabian Sea with cloud coverage less than 30%. But the Bay of Bengal is characterised by cloudy conditions caused due to the prevailing northeast monsoon even after the withdrawal of summer monsoon clearly indicates that clouds exhibit drastic seasonal variability in the Arabian Sea and Bay of Bengal compared to the south Indian Ocean. The features explained here are consistent with the observations of Hasternrath and Lamb (1979).

CHAPTER II

OBJECTIVES, DATA AND METHODS

2.1 Introduction

The discussion presented the chapter I clearly indicated that the tropical Indian Ocean is unique among the world oceans owing to the presence of number of oceanograhic features. These features make the prediction a challenging problem for the modellers. It is essential to understand the variability of these features both on temporal and spatial domains before attempting modelling studies. The best way to tackle this problem is to analyse the surface marine met fields. The very objective of present study is to document various sea surface parameters of the Indian Ocean and to examine the anomalies found in them. Hence to relate the anomaly to the variability of monsoon over India, highlighting the occasion of contrasting monsoon periods.

Since atmosphere and ocean act as a closely coupled system it is essential to study their behaviour in a detailed manner to document the variability both on temporal and spatial domains. From the above discussions it is observed that tropical Indian Ocean poses a major problem for modellers with all the complex features described in the above sections. It is felt that the variability of the fields has to be addressed in a comprehensive manner to understand the air-sea interaction processes over the tropical Indian Seas. In this thesis attempts have been made to study the variability of surface marine meterological and heat fluxes and their possible link with summer monsoon activity.

The thesis is presented in six chapters. Besides this introductory chapter, the thesis contains five other chapters.

Chapter II deals with the objective of study, data used, methodology followed and the limitations of the data.

The anomalies of the sea surface temperatures, sea level pressure and wind speed, wind direction and cloud are the four parameters involved in the investigated in chapter III. Here we are interested in the following main questions.

(a) What is the nature of variability of sea surface temperature, sea level pressure, surface wind and cloud in the Indian Ocean area?

(b) Do these variability in Indian Ocean coincide with the occurrence of Dipole events, Equatorial Indian Ocean Oscillations and ENSO related features?

(c) How do they feed back on the variability of monsoon over India?

Chapter IV describes the latent heat radiation, short wave radiation, net heat radiation, evaporation and heat budget anomalies of the Indian Ocean for the deficit and surplus monsoon years. Also investigate whether they can explain the occurrence of good and bad monsoons of India.

Chapter V depicts the relation between Indian Summer Monsoons and the occurrences of Equatorial Indian Ocean Oscillation and ENSO related features.

The variability of Indian summer monsoon is widely discussed in this chapter.

Chapter VI summarises the results and portray the conclusions of the work.

2.2 Data

Since this thesis is concerned with the variability over the tropical Indian Seas, we have used Comprehensive Ocean Atmospheric Data Set (COADS) processed and averaged over 1° known as Surface Marine Data Set (Da Silva et al 1994). Among many data fields present in the data set we make use of objectively analysed fields of surface marine anomalies of sea surface temperature, sea air temperature, sea level pressure, surface wind, fractional cloud cover, latent heat radiation, short wave radiation, net heat radiation and evaporation over most of the global ocean. These revised monthly mean fields are derived from individual observations in the Comprehensive Ocean-Atmosphere Data Set (COADS) January 1945 to December 1993 and are analysed on a 1° by 1° global grid. Corrections have been made to reduce wind speed bias associated with an erroneous Beaufort equivalent scale, and to quality control night-time fractional cloud cover observation according to the brightness of the sky. Various anomaly fields from January 1945 to 1993 December have been used for the analysis.

The availability of this data set has significantly contributed to advancing our understanding of the atmosphere-ocean climate system. One of the main contributions of the COADS project was to unify several historic data sets in a single and consistent format and to subject the ship reports to the same quality control procedures. The primary data source for this study is the Release I of the Comprehensive Ocean-Atmosphere Data Set (COADS), which cover the period of 1854-1979, and subsequent releases cover upto 1993.

Anomalies are the departures of a parameter from its mean value. It is computed by subtracting the monthly climatological mean derived from the whole length of data, from the individual observations. Positive values of the anomalies indicate high values compared to the climatology and vice versa.

The revised anomaly fields are derived from individual observations in the Comprehensive Ocean Atmospheric data set (COADS, Da silva et al, 1944) from January 1945 to December 1993 are analysed on 1° by 1° global grid. Objective analysis has been done on the data by successive correction method with Barne's weight function same response functions as Levitus (1982). Climatological Atlas of world oceans has been used by the developers of COADS data set.

The data used is in the netCDF (Network Common Data Form) format, a data format developed at the University centre for Atmospheric Research, University of Boulder, Colorado, and U. S.A. It is an interface for array – oriented data storage and access, which is widely used in climatological research and other fields of Geosciences. Many important data sets and model output files are distributed and shared among the scientists in the netCDF format.

Quality control in COADS is implemented by means of multiple statistical procedures to identify outliers. The first step is to generate *Decadal Summary Untrimmed Limit* for six variables. A detailed description of the statistical procedure can be found in Slutz et al.(1985). In general an observation is rejected if it differs from the smoothed median by more than 3.5 standard deviation

2.3 Limitations and shortcomings of the data set

A description of the marine data set is not complete without a discussion on the problems and shortcomings of the data set. Ship reports are the only record available for surface marine climate. This data set undoubtedly continued to be used in studies of the Ocean-Atmosphere system. An analysis of surface marine data is an evolving system. Constant improvements are necessary as more data or metadata becomes available, physical parameterisations are improved or advances of data assimilation techniques become available. The corrections reported here are the best in the right direction but much remains to be done.

The first and the most uncontrollable errors exist in the ship report themselves. Marine and atmospheric measurement techniques and the recording of these observations are not perfect. Some errors are introduced through poor instrumentation. Surface air temperature can be biased due to heating of the ship's superstructure during the day and inadequate surface ventilation. (Isemer and Hasse 1987; Ramage1984; Kent et al 1993a, b). Sea Surface

Temperature observations taken via ship intake will be biased compared to those taken by canvas bucket.

A proper homogenisation of the measured wind speed requires the knowledge of the precise anemometer height information, which is not included in COADS. An average anemometer height of 20 m has been assumed in our calculation. The discrimination between measured and estimated winds is not entirely reliable (Slutz et al. 1985, Cardone et al.1990. Our wind speed bias correction producer takes the WI flag at face value and consequently is affected by this uncertainty.

Even when an observation is considered accurate and unbiased, the recording practices may result in errors (Slutz et al. 1985). Some observations are taken in different units (such as Fahrenheit vs. Celsius) without documentation of the fact. All observers do not implement coding practice changes on the same date. Ship position may be erroneous. The transfers of written record to digital also introduce possible errors. Little can be done to miscoded or other human errors. A certain amount of recording errors were repaired by the compilers of COADS, but some miscoded errors undoubtedly remain. The tendency of ships to avoid the stormy weather bias into observations when considered as a whole, although this bias is less problematic in the tropics.

After we deal with correctable biases and calculate raw fields, errors can be introduced by the analysis scheme as well. One problem with the successive correction method used in this analysis is the introduction of unrealistic features (Levitus 1982) One of the most severe problems with surface marine data is the poor coverage in some regions especially away from shipping lanes. Objective analysis of the sort used here fills in the gaps by interpolating/extrapolating smooth data from remote region, but cannot compensate entirely for poor sampling. This effect is particularly troublesome in the tropics and southern oceans where observations are clumped along ship tracks with data void area in between. As a result, wrong features sometimes remain in analysed fields, particularly in anomaly fields.

The analysis of anomaly values rather than the observed values can result in spurious extrema. For example, the analysis has problems handling positive definite quantities such as fractional cloud cover, which only takes values between 0 and 1. The analyses procedure can produce full field values (climatology plus anomalies) less than 0 in broad clear areas or values greater than 1 in broad overcast region. It should also be kept in mind that although the analysed fields are given on a 1° by 1° grid; only features with wavelength greater than 770 km. are retained. The reader is referred to Levitus (1982) or Dailey (1991) for additional discussions of the limitations of the objective analysis technique.

Despite these limitations the COADS data set remain as the best data set to study the broad scale features occurring over the oceans especially during the present study period of 1945 – 1993.

CHAPTER III

ANOMALIES OF SURFACE TEMPERATURE, WIND, PRESSURE AND CLOUD OVER THE INDIAN OCEAN

3.1 Introduction

The heat storing capacity of the ocean and time involved in the ocean circulation play major role in determining regional as well as global climate. After decades of research, it came to know that El Nino Southern Oscillation (ENSO), a phenomena chiefly associated with Pacific Ocean with its root in Indian ocean grows globally and upsets the weather pattern all over the world (Meehl,1989, 1994; Yasunari and Seki,1992: Yasunari 1996). A rise of sea surface temperature in the central Pacific Ocean evidences its effects, some times months later, across the globe. The oceans can no longer be considered as a passive pattern in climate control.

Ocean is a major contributor to total planetary heat transport and is a major source of atmosphere water vapour. Its interaction with the atmosphere occurs through the SST. The amount of energy transfer to the atmosphere is primarily decided by the Sea Surface Temperature. In other words, the rate of evaporation near the sea surface is determined mainly by SST and also by the wind speed.

The role of Indian Ocean in determining the summer rainfall over Indian subcontinent has been worked out by many authors. Krishnamurti et al (1985) came out with a suggestion that interannual variation in the activity of summer monsoon is related to the variation of sea surface temperature (SST) over the Indian Ocean. Webster et al (1998) have ascertained that the Indian Ocean region potentially influences the monsoon circulation. Tourre and White (1995) explained an eastward propagation of ENSO signal from Indian Ocean to Pacific. Indian Ocean is also now recognised as a source of other modes of climate variability like the dipole mode (Saji et al 1999, Webster et al 1999, 2001; Behera and Yamagata, 2003).

In order to study the effect of sea surface anomalies of meteorological parameters on the process of Indian summer monsoon, first we have to categorize the summer rainfall as deficit years of rain, surplus years of rain or as years of normal rainfall. In this chapter we look into special features of sea surface anomaly fields (sea surface temperature, sea level pressure, wind speed, wind direction and cloud) occurred during contrasting very good and very bad monsoon periods of India during 1945 to 1993. Different methodologies are followed for finding out the deficit/surplus rainfall years. The classification that is considered to be rational for the tropical regions with high spatial and temporal variations of rainfall is followed by Parthsarathy et al (1987). Similar classification is carried out in this work also.

The new update of All India Monthly and Seasonal Rainfall Series: 1871 – 2002 has been downloaded from the web site of IITM (Indian Institute Of Tropical Meteorology – Pune). This long homogeneous rainfall series of All India has been prepared based on a fixed and well-distributed network of 306 rainguage stations over India by giving proper area weightage.

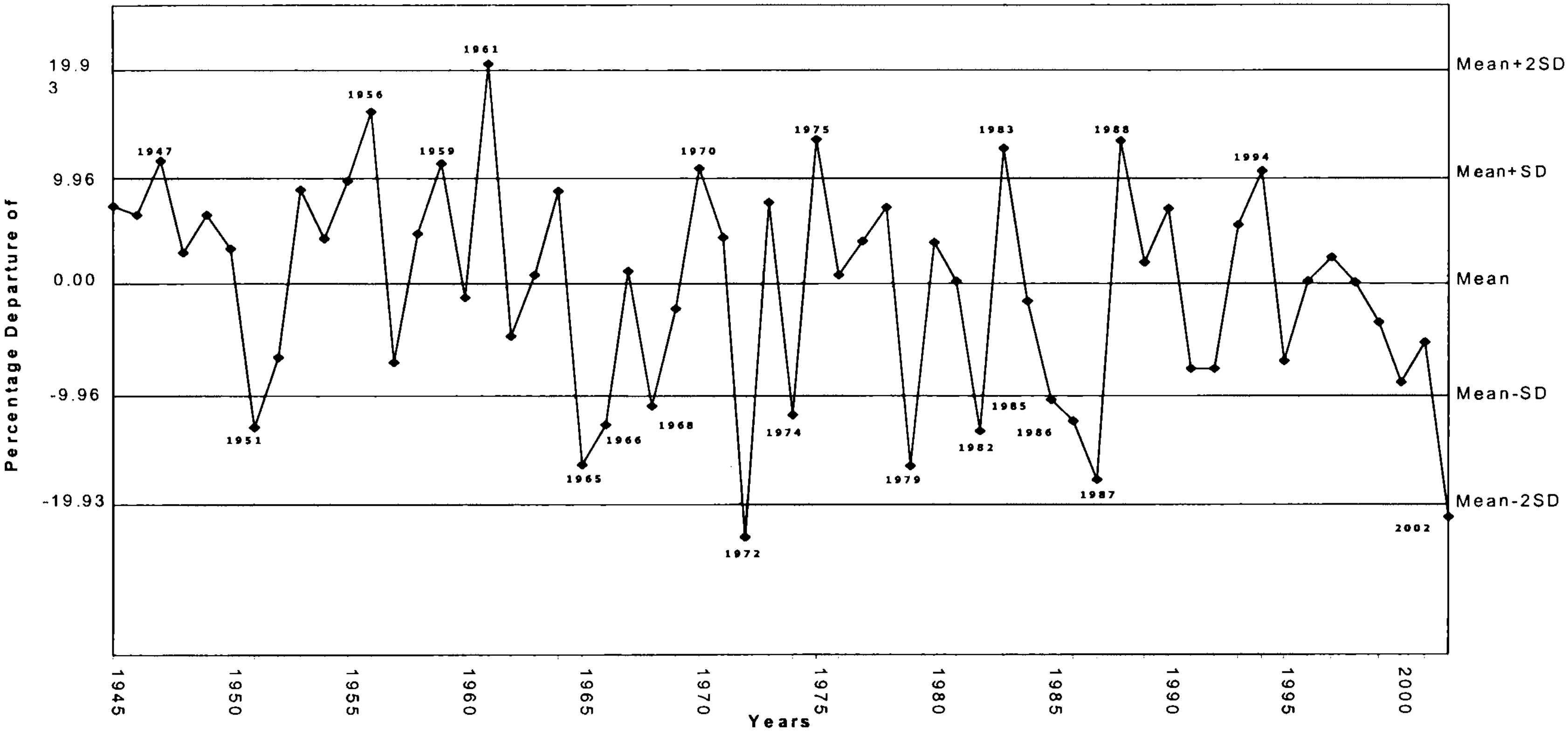


Fig.3.1 Percentage departure of Indian Summer Monsoon Rainfall (1945 - 2002)

It is seen from the Fig. ~~Table~~3.1 that the monsoon months of June, July, August and September are the main rainy months and they altogether contribute about 78.2% of the total annual rainfall. The All India Monsoon Rainfall (ISMR) series expressed as percentage departures from the long term mean of 1871-2002. A year has been classified as excess (wet) monsoon rainfall year when R_i is above R+S and as deficient (dry) when R_i is below R-S, where R_i is the monsoon rainfall of the i^{th} year. R the mean and S the standard deviation of the series. This classification is considered to be rational for the tropical regions with high spatial and temporal variations of rainfall with crop growth and water requirement and management tuned to the local mean conditions (Parthasarathy et al., 1992b).

The percentage departure of all India summer monsoon rainfall for a period of 1945 –1993 (Fig.3.1) suggests the rainfall activity has been quite variable. Excess rainfall from the normal was observed in 1956, 1961,1966,1970, 1975,1983 and 1988. Severe drought conditions occurred in 1951, 1965, 1966, 1972, 1979, 1982 and 1987.

3.2 Sea Surface Temperature Anomaly

It is worth noting that the SST anomalies in Tropical Indian Ocean are very small and often less than 0.5°c. Hence the accuracy of observations is a crucial factor in arriving at conclusions regarding the relationship between SST anomalies and monsoon rainfall. The simple averaging of such sensitive values may misguide the real situation in anomaly studies.

Earlier studies based on COADS data have observed a zone of positive sea surface temperature anomalies in the northern and western Arabian Sea during surplus monsoon years. On the contrary, negative sea surface temperature anomalies are noticed over the same region during deficit monsoon years (Mohanty et al 1994; Ramesh Kumar et al, 2000). These analyses were based on composite sea surface temperature fields of the entire good monsoon years and bad monsoon years for long periods.

However, while similar trends are found in western tropical south Indian Ocean moderately negative anomalies are noticed in the eastern tropical Indian Ocean during surplus monsoon years and vice versa in deficit monsoon years (Mohanty et al 1994).

In the present analysis, the SST anomaly is looked for individual years of both good and bad rainfall. The years of 1956, 1961,1975,1983 and 1988 are chosen from among excess rainfall years for sea surface temperature analysis whereas among the drought years, 1951,1965,1972,1979,1982 and 1987 are considered. Out of these severe drought conditions were experienced in 1972 and 1987, both El Nino years. Major findings for pre monsoon and monsoon months are discussed below.

3.3 SST anomalies during good monsoon years

Obviously, the analysis shows contrasting features of SST anomalies

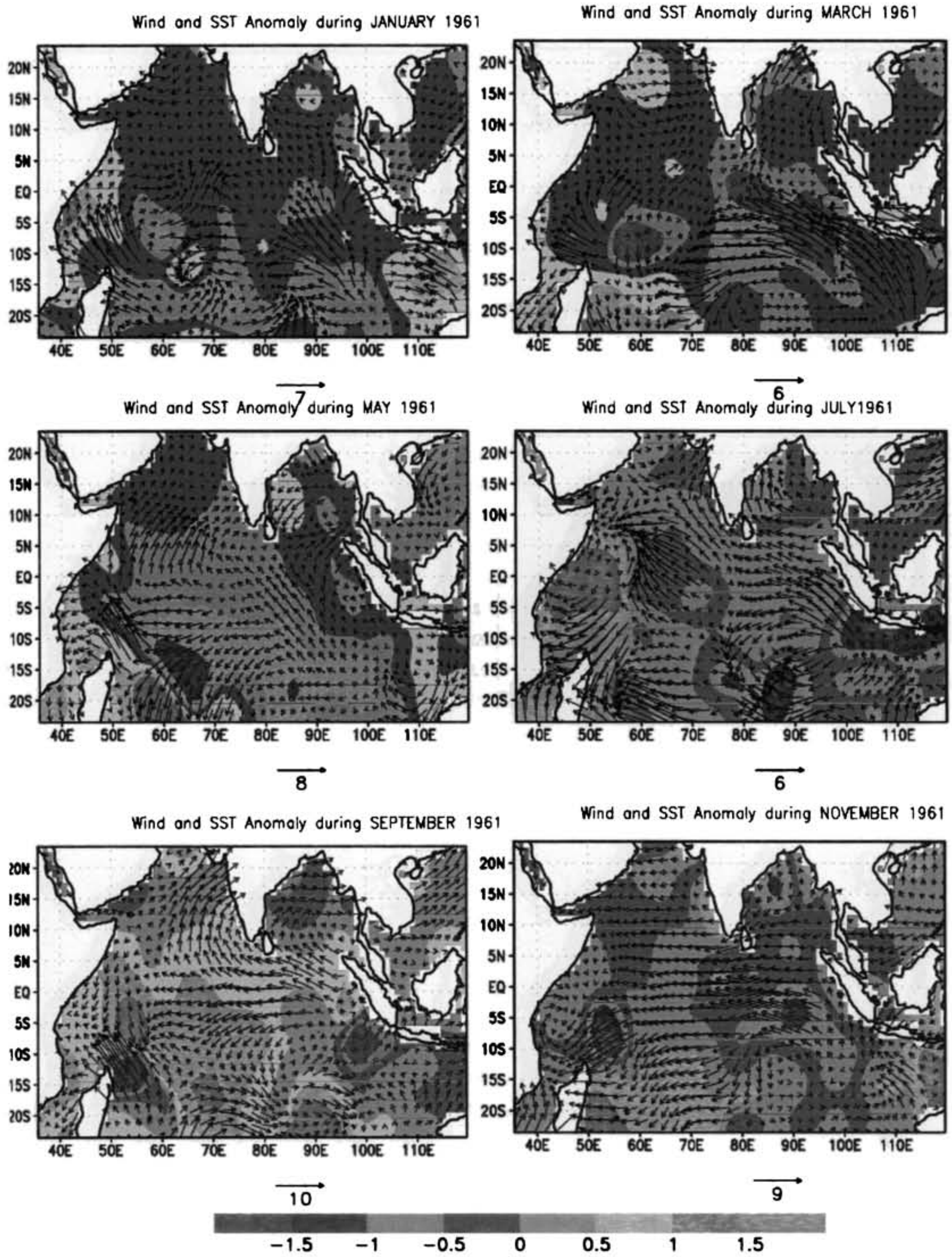


Fig: 3.3 WIND (m/s) AND SEA SURFACE TEMPERATURE (°C) ANOMALY 1961

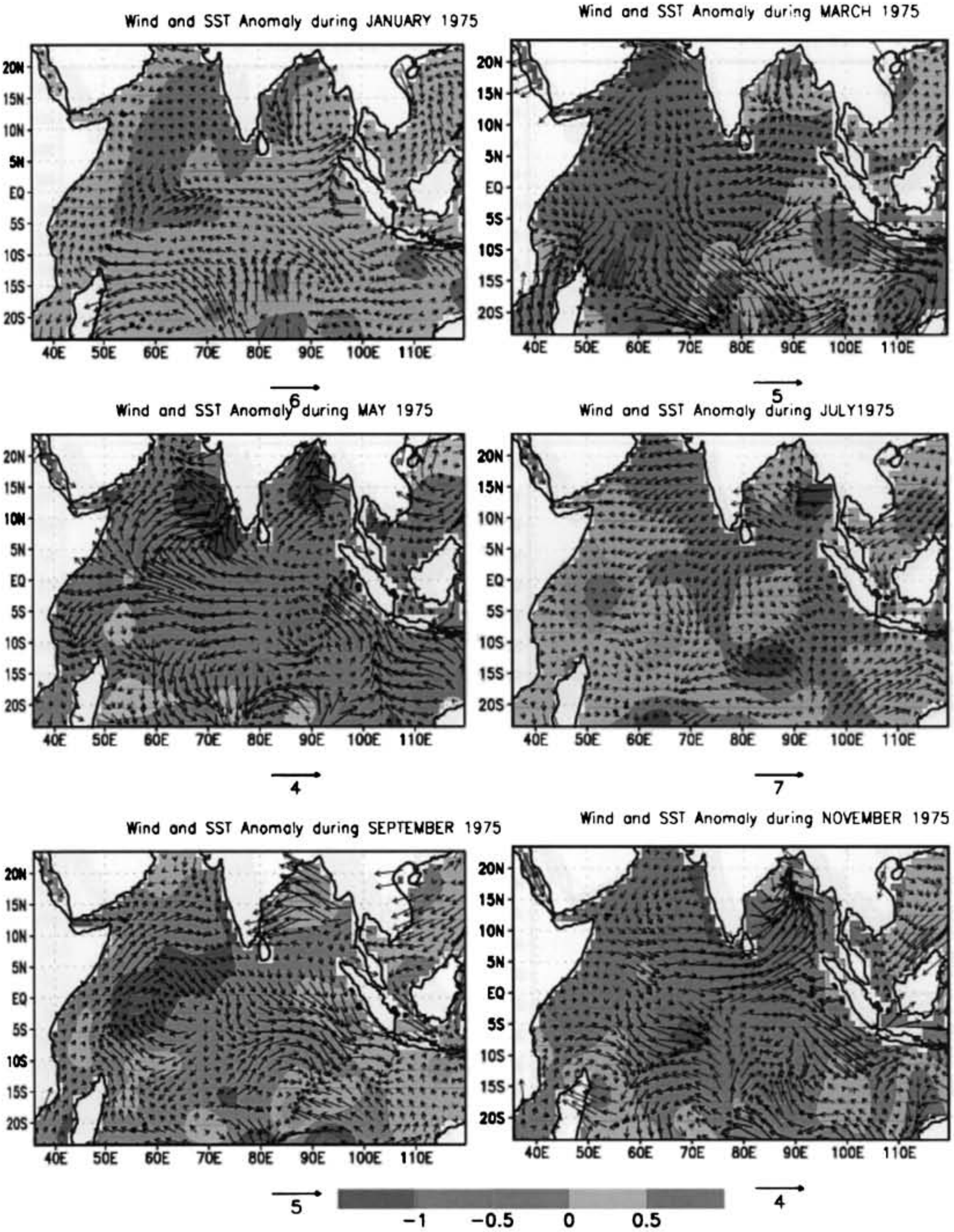


Fig: 3.4 WIND (m/s) AND SEA SURFACE TEMPERATURE (°C) ANOMALY 1975

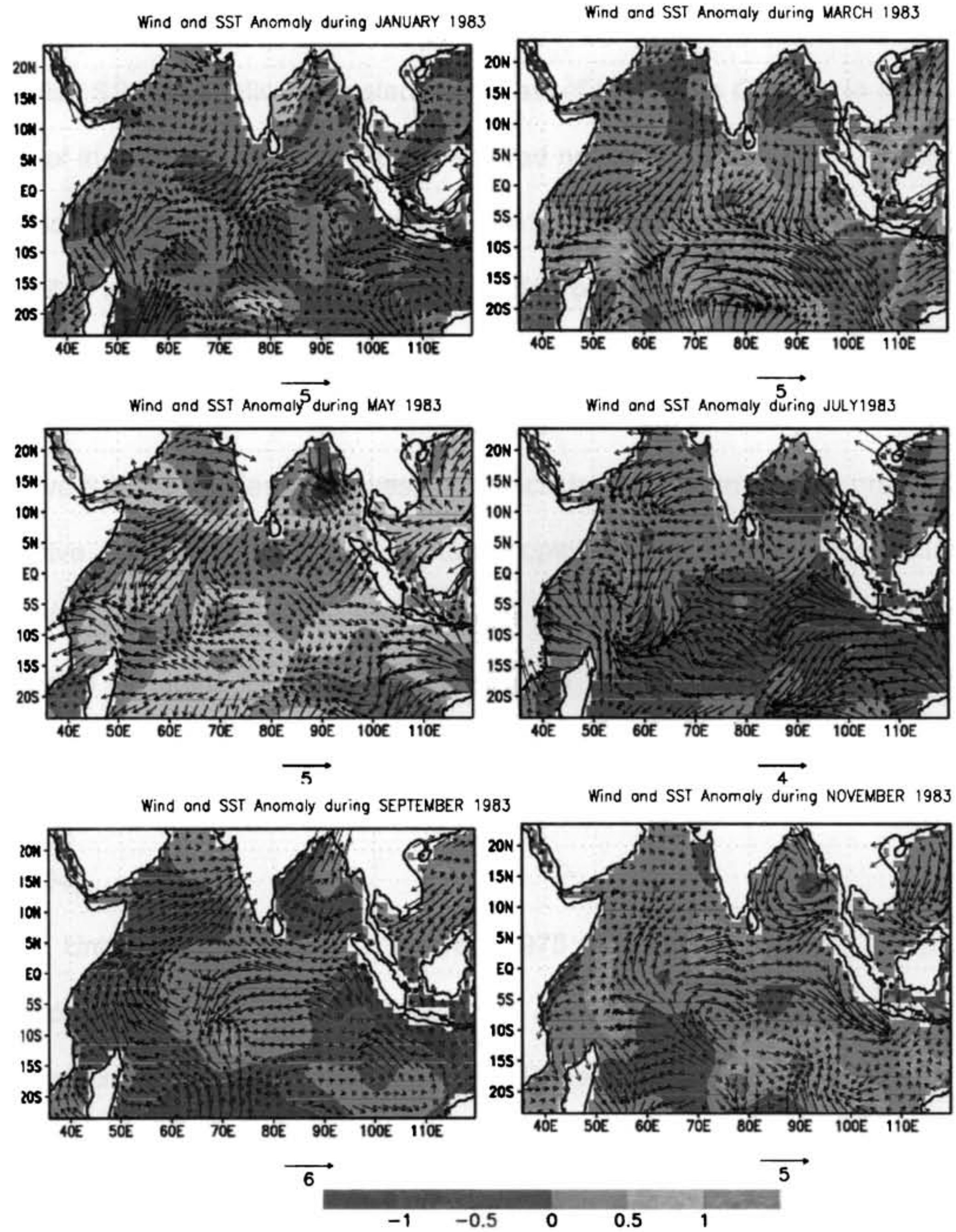


Fig: 3.5 WIND (m/s) AND SEA SURFACE TEMPERATURE(°C) ANOMALY 1983 during the pre monsoon period of good monsoon years. For example, during 1956 (Fig: 3.2), positive SST anomalies can be observed only in the northern Arabian Sea and eastern tropical Indian Ocean in both March and April whereas negative SST anomalies persisted in the rest of the Indian Ocean. In that year, most of the regions of Indian Ocean showed negative SST anomaly throughout the monsoon season. But as regards 1961, (Fig: 3.3) positive SST anomalies prevailed in the northern Arabian Sea during both pre monsoon and monsoon periods. Positive SST anomalies were also observed in the Bay of Bengal from March to June. Another important feature in 1961 was the occurrence of strong positive SST anomalies in the western tropical Indian Ocean and the presence of negative SST anomaly in the eastern tropical Indian Ocean. This condition prevailed from July and became stronger by October. This peculiar feature is attributed to Indian Ocean Dipole (Saji et al 1999; Yu and Reinecker 1999; Webster et al 1999 Murtugudde et al 2000). This led to severe drought in the Indonesian region and anomalously large rainfall over eastern Africa (Kapala et al 1994).

Unlike in 1956 and 1961, the year 1975 showed negative SST anomaly in the northern Arabian Sea from March to August (Fig: 3.4). A negative SST anomaly in the western equatorial Indian Ocean and a positive anomaly in the eastern tropical Indian Ocean during September to October indicate a strong negative Indian Ocean Dipole. Further year a major part of northern Indian Ocean showed negative SST anomaly during premonsoon and monsoon.

Positive SST anomalies (Fig. 3.5) existed in the major part of the Indian

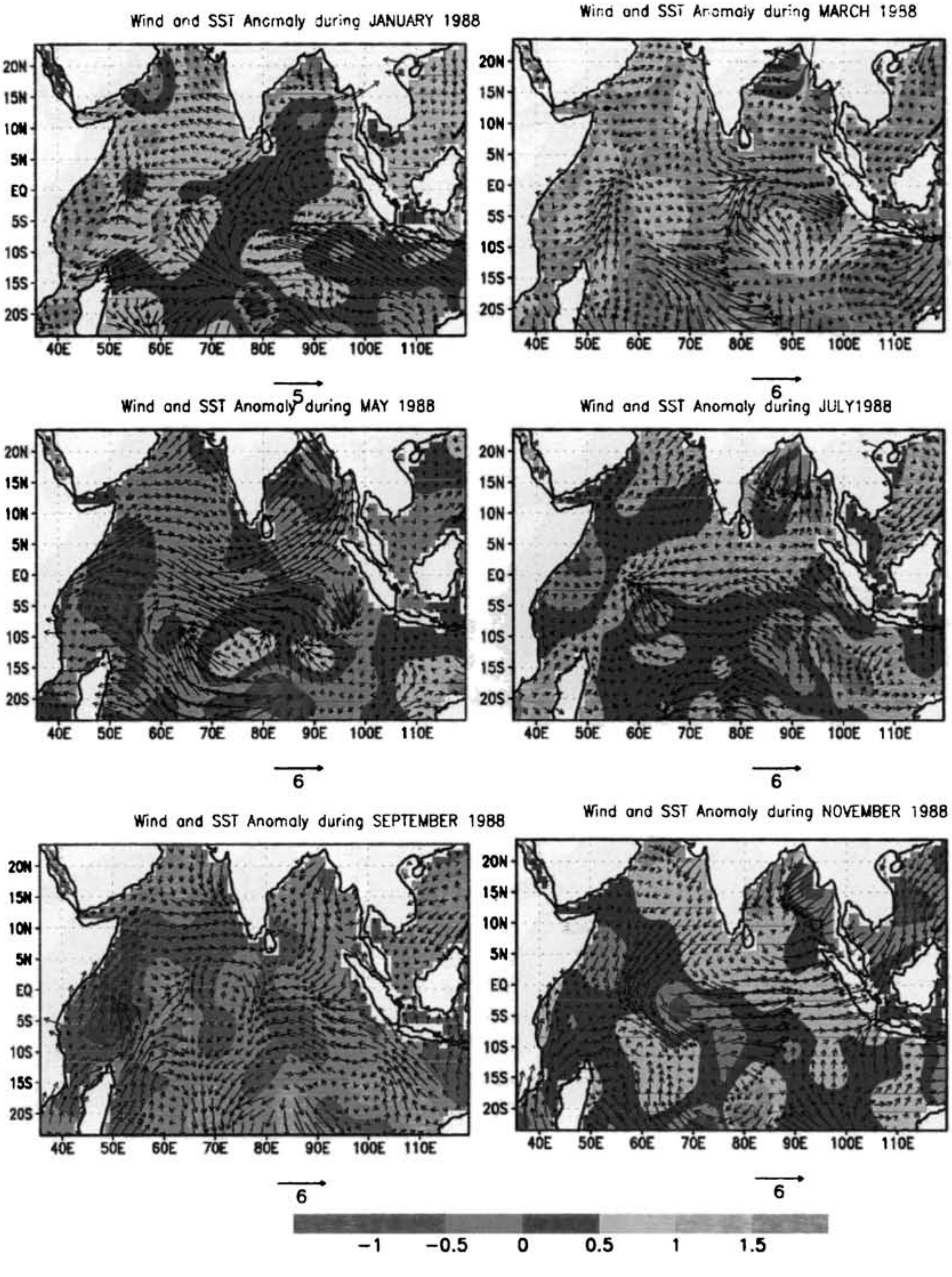


Fig: 3.6 WIND (m/s) AND SEA SURFACE TEMPERATURE (°C) ANOMALY 1988

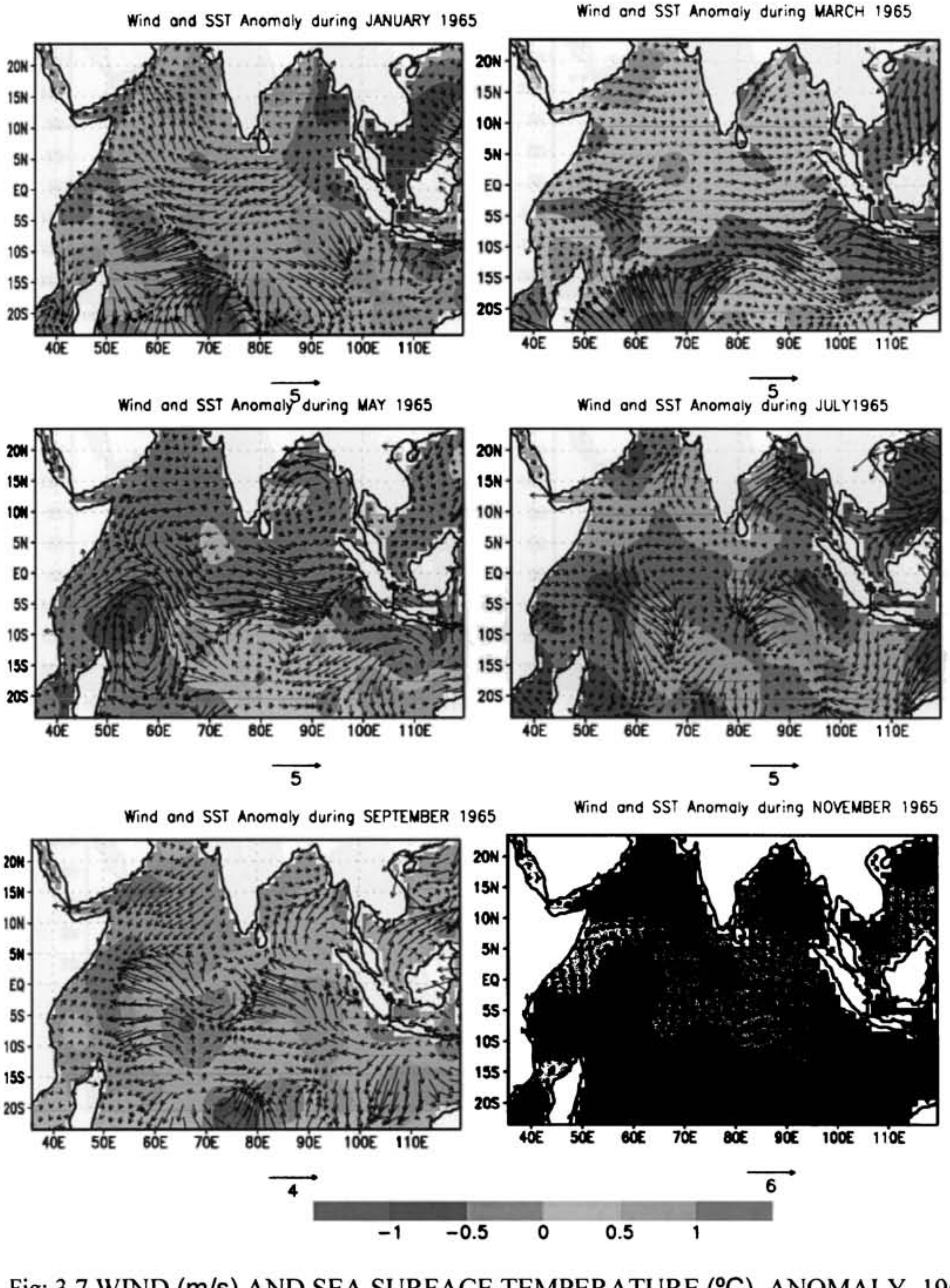


Fig: 3.7 WIND (m/s) AND SEA SURFACE TEMPERATURE (°C) ANOMALY 1965

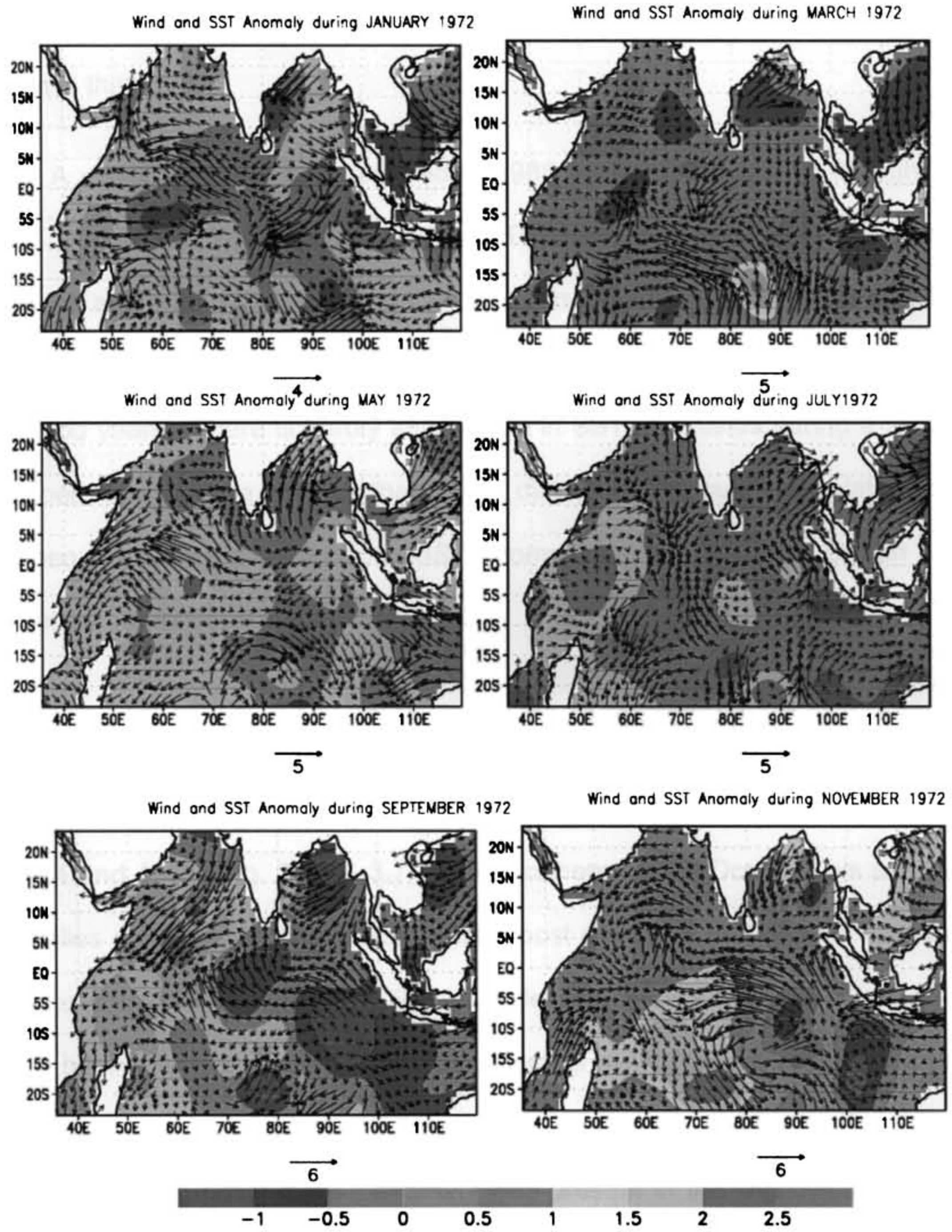


Fig: 3.8 WIND (m/s) AND SEA SURFACE TEMPERATURE (ºC) ANOMALY 1972

Ocean during 1983 another year of good monsoon. A moderate El Nino occurred in 1982-83. This could be the reason for higher than normal SST in the Indian Ocean in that year.

A very good monsoon occurred in 1988 following the severe drought of 1987. During 1988 (Fig 3.6) positive SST anomalies persisted during pre monsoon and during the course of the monsoon period. This suggests that SST anomaly patterns just before the onset of monsoon are very similar for drought and flood years. There is hardly any pattern in SST anomalies during a strong monsoon year. In the light of these it is difficult to speculate a relationship between premonsoon or monsoon SST anomalies, over the Indian Ocean and the nature of monsoon in that year.

3.4 SST anomalies during bad monsoon years

Very strong positive SST anomalies could be absorbed in the premonsoon of 1951 and 1965 (Fig. 3.12 & 3.7) in the northern Indian Ocean. This positive anomalies continued during monsoon and post monsoon though with reduced magnitude. In general, negative SST anomalies were noticed in 1972 (Fig: 3.8) in March but in May maximum negative anomaly of over 1°C was observed in the equatorial western and eastern Indian Ocean. It may be noted that during July – October a strong positive SST anomaly was present in the western equatorial Indian Ocean and a mild negative anomaly in the eastern tropical Indian Ocean. It indicates a weak positive dipole year. By September - October, the positive anomaly spreads to major portion of north Indian Ocean.

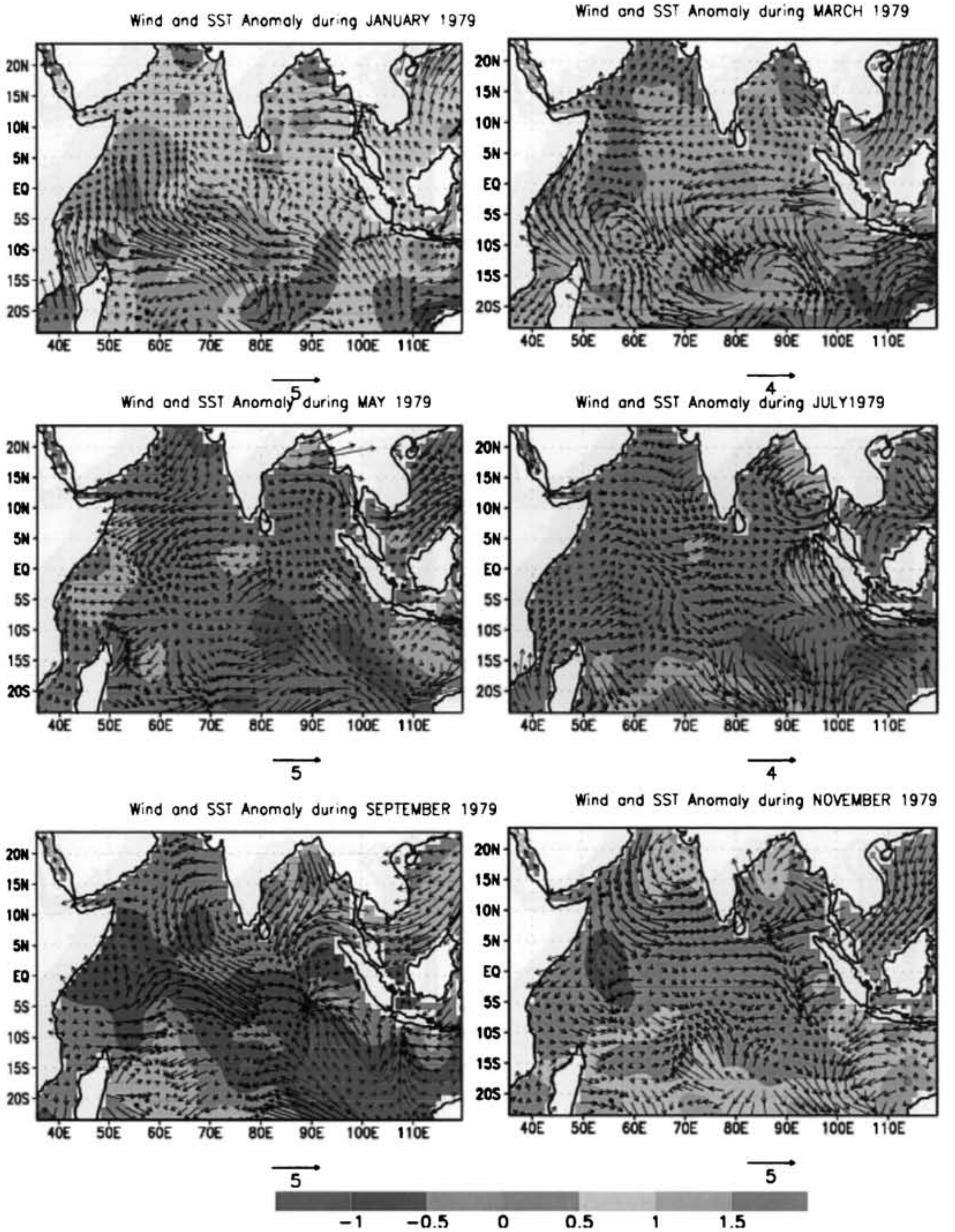


Fig: 3.9 WIND (m/s) AND SEA SURFACE TEMPERATURE (°C) ANOMALY 1979

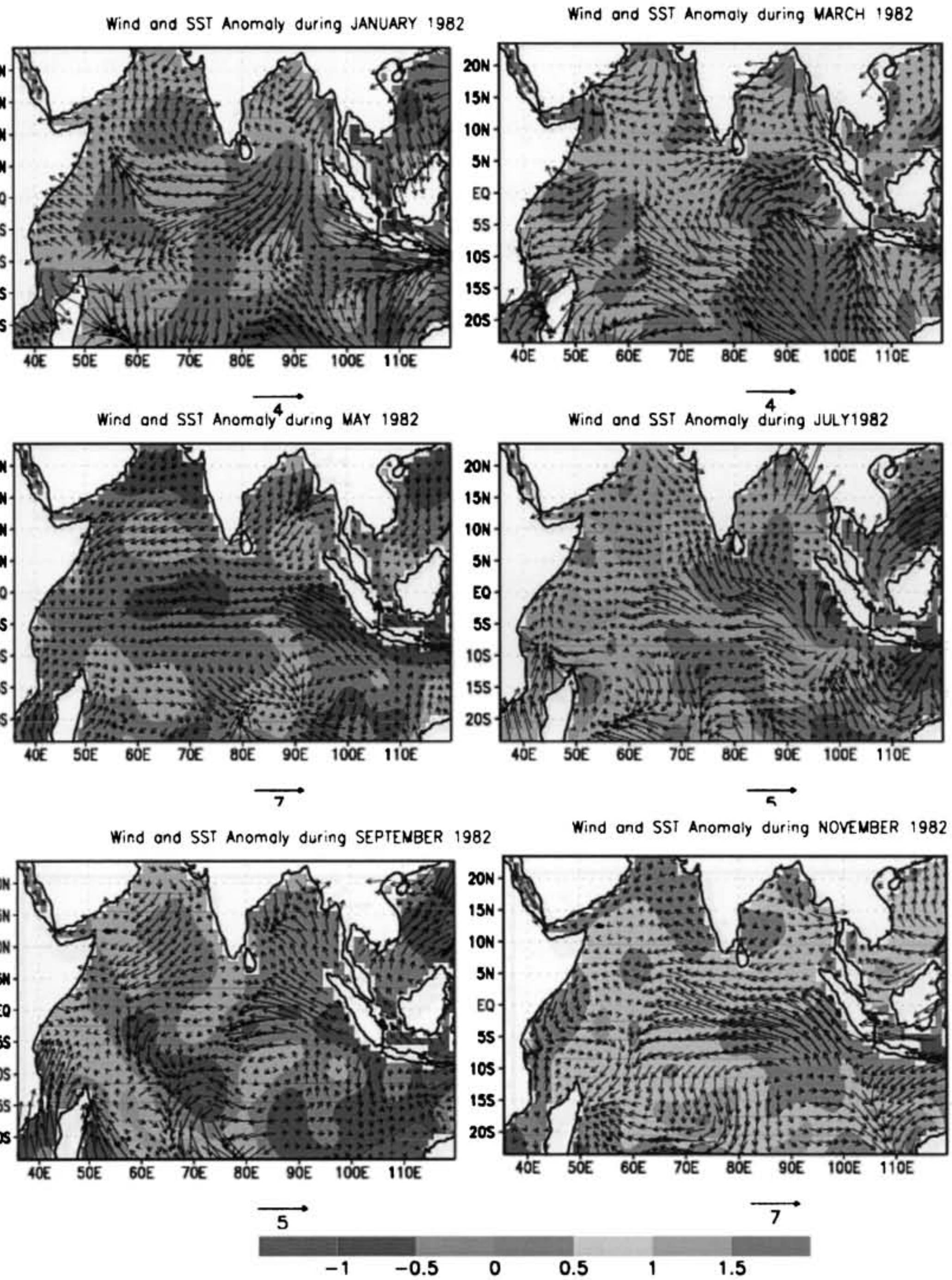


Fig: 3.10 WIND (m/s) AND SEA SURFACE TEMPERATURE (°C) ANOMALY 1982

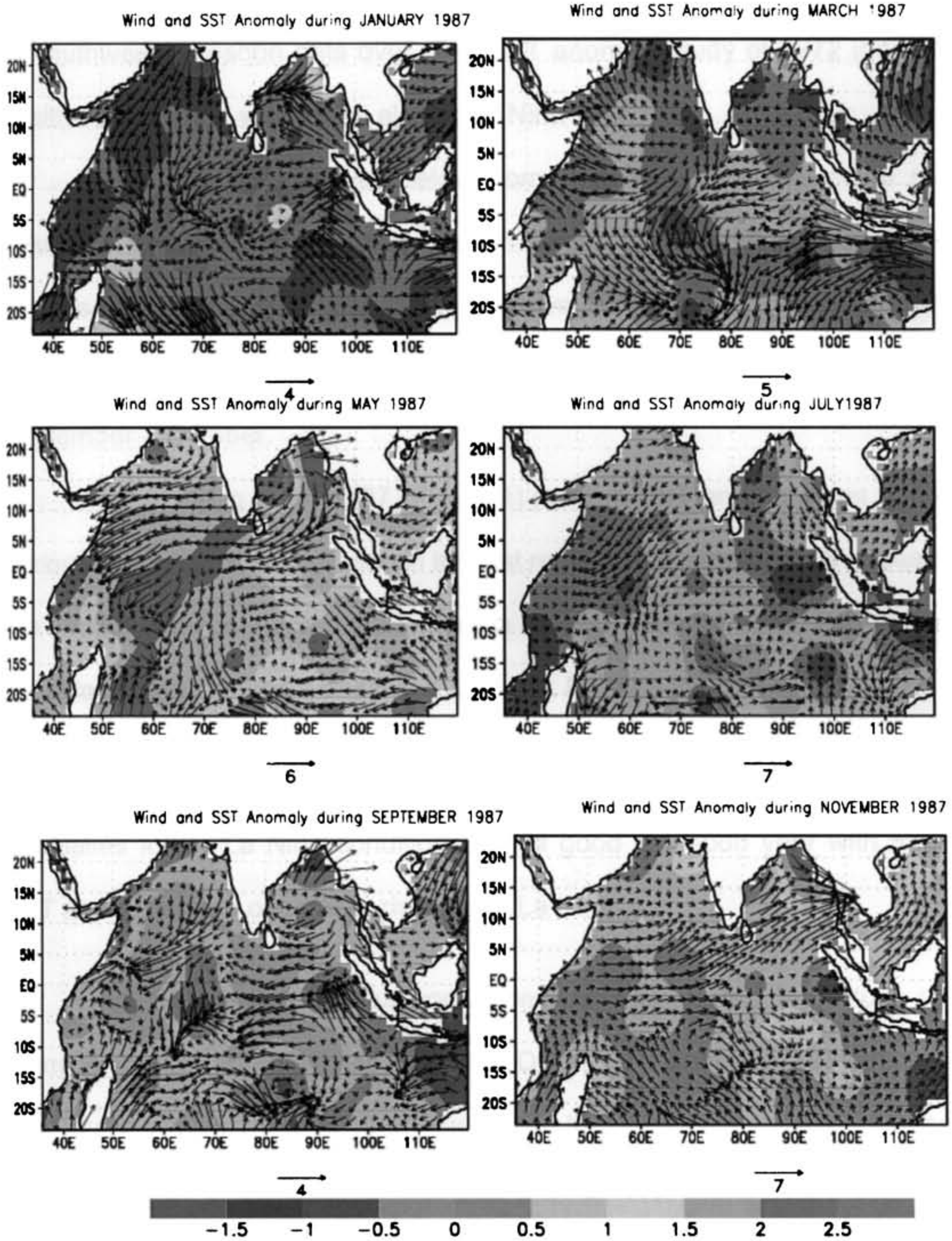


Fig: 3.11 WIND (m/s) AND SEA SURFACE TEMPERATURE (°C) ANOMALY 1987

In 1972, the negative SST anomaly observed in the major part of Indian Ocean is replaced by positive anomaly by May – June (Fig: 3.8) and it continues till southwest monsoon gets over. The SST anomaly study of 1972 indicates a positive IOD event, which was also an El Nino year.

The monsoon was well below normal in 1979, which showed mostly positive anomalies in the northern Indian Ocean during both the pre monsoon and the monsoon periods (Fig: 3.9). Negative SST anomaly was dominant only in the western tropical Indian Ocean and northwest off Australia during September – October.

In 1982 (Fig 3.10) 1987 (Fig 3.11) positive SST anomaly was dominant during pre monsoon and monsoon in most regions of northern Indian Ocean and the Indian subcontinent experienced a severe drought. The SST anomaly conditions also showed a positive IOD event. It was a strong El Nino year.

It is seen from this study that a drought year with generally negative SST anomalies follow La Nina conditions and a good monsoon year with negative SST anomalies are often associated with La Nina.

The SST anomaly was considered as one of the indicators of the monsoon rainfall anomaly over the Indian Ocean (Krishnamurti, 1985). However, later studies (Webster et al 1998) suggested that the SST anomaly could hardly be an indicator of the summer monsoon rainfall over India. The present study suggests that there are many good monsoon years with predominantly large negative SST anomalies over the northern Indian Ocean (eg. 1956) often associated with La Nina and predominantly positive SST anomalies with good monsoon rainfall (eg. 1983,1988) following a severe El Nino and bad monsoon in the previous year. The bad monsoons of 1951, 1965, 1972, 1982 and 1987 where all associated with El Nino and with predominantly positive SST anomalies in the Indian Ocean but the bad monsoon of 1979 was independent of ENSO. The large scale Indian Ocean SST anomalies are closely linked ENSO events (Chambers et al 1999: Klein et al, 1999 and Kawamura et al, 2001). It has been observed that, unlike in major parts of the world oceans, positive SST anomalies are generally not associated with deep convection in the Indian Ocean region (Webster and Young 1992). Hence regions of warm SST may not be regions of negative OLR in the Indian Ocean (Webster et al 1998). This could be one of the major reasons for any one to one direct relationship between Indian summer monsoon rainfall (ISMR) anomalies and SST in the Indian Ocean.

3.5 Analysis of Surface Wind Anomaly

The composite of surface wind speed anomalies for all drought years and flood years taken respectively showed considerable weakening of wind during drought years (Mohanty et al 1994). The maximum weakening is observed in western Arabian Sea during May and in the central and western Indian Ocean during monsoon (Mohanty et al 1994). This study also showed stronger wind during drought years in the northern Bay of Bengal and southeastern Indian Ocean.

The wind anomaly (vector) clearly shows a substantial reduction in the

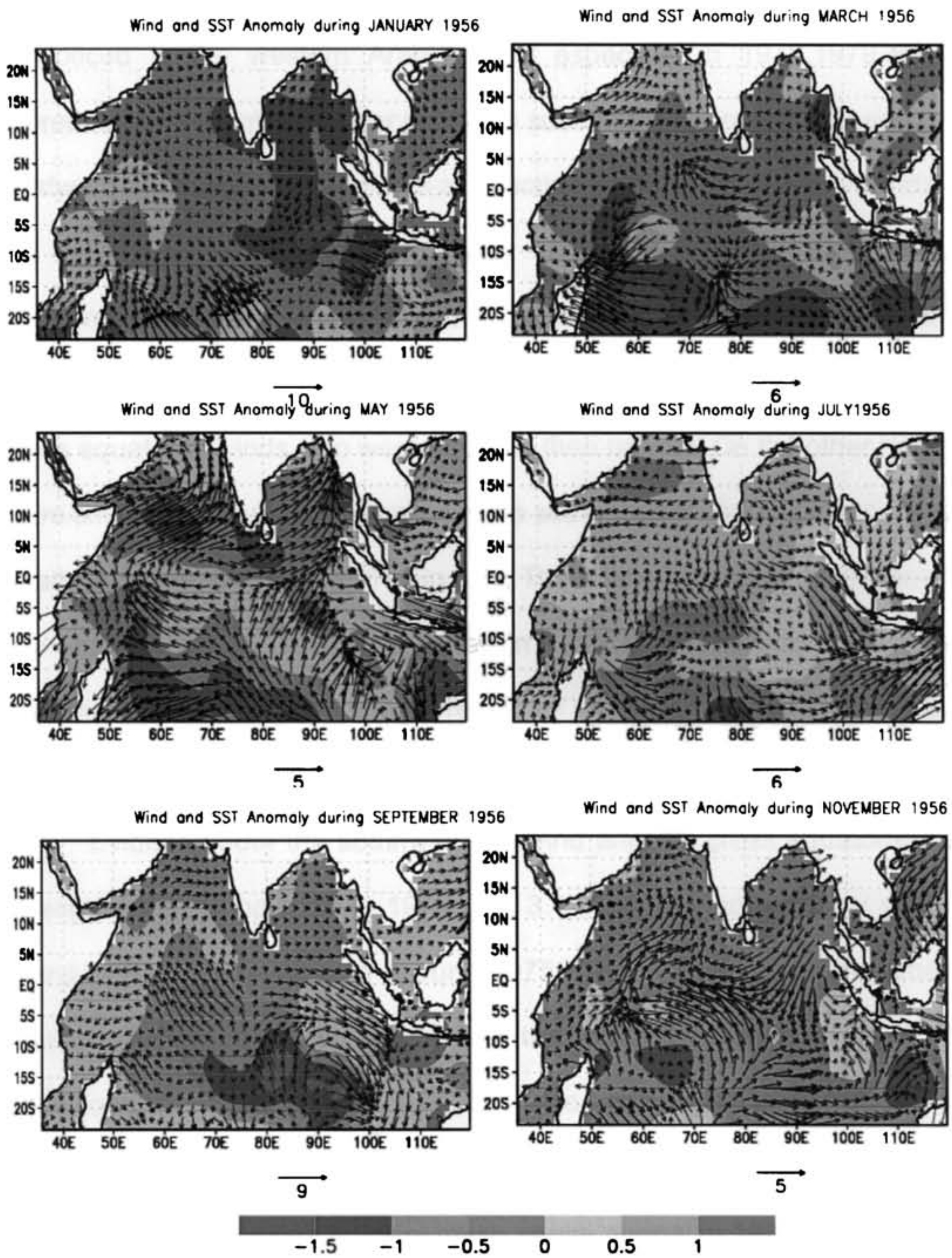


Fig: 3.2 WIND (m/s) AND SEA SURFACE TEMPERATURE (°C) ANOMALY 1956

wind speed in May both in Arabian Sea and western part of Bay of Bengal during all drought years (1951, 1965, 1972, 1979, 1982, 1987). The maximum reduction is noticed in the western Arabian Sea especially in 1972,1979,and 1987. Moreover wind anomaly vector suggests substantial decrease in speed (a south eastward anomaly vector indicate a reduction in the south westerly winds.

During the good monsoon year 1956 (Fig 3.2), the strengthening of southwesterlies is obvious from May onwards in the Arabian Sea and this continued through the entire monsoon. During major part of the monsoon, the cross equatorial winds also were stronger than normal. On the other hand, there were southeasterly wind anomalies in the southeastern Indian Ocean, which was totally opposite in sign in 1961(Fig 3.3). The strengthening of southwesterlies in the northern Indian Ocean continued in 1961 also. The strong easterly wind anomalies exhibited in southeast region of equatorial Indian Ocean especially after the monsoon could be the outcome of IOD event, which occurred that year.

Evidently both the southwesterly wind and the cross equatorial flow got strengthened during May in 1975 (Fig 3.4). Large wind anomaly vectors are found in Bay of Bengal also in June 1975. Except in July the strengthening of southwesterlies continued throughout the monsoon period in 1975. Similar conditions prevailed in 1988 (Fig 3.6) also, which was exceptionally a good monsoon year.

Though the monsoon was much above normal in 1983 (Fig 3.5) the wind anomaly patterns were totally different unlike any other monsoon years. Right from May till July this pattern continued whereas in August and September the

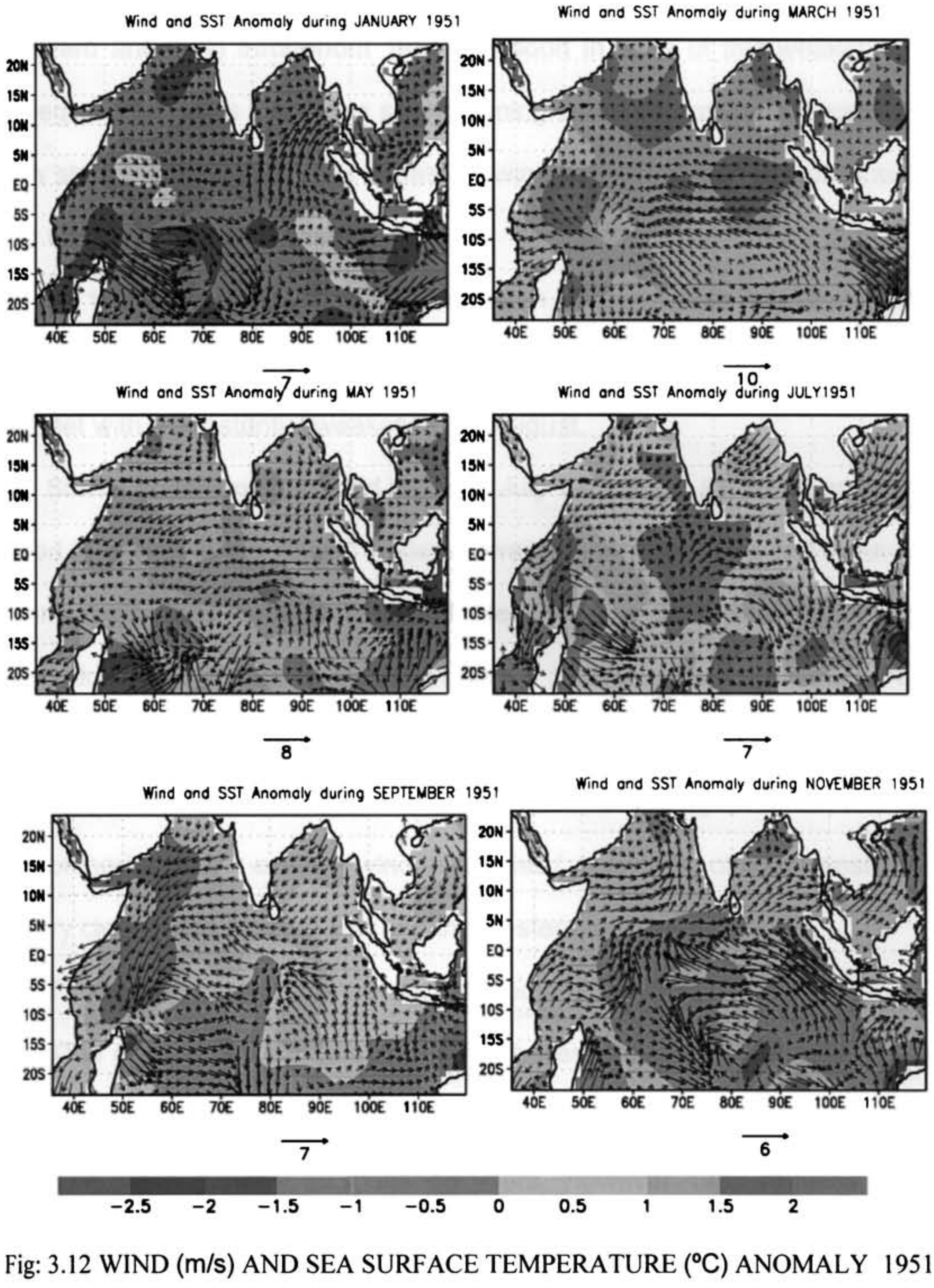


Fig: 3.12 WIND (m/s) AND SEA SURFACE TEMPERATURE (ºC) ANOMALY 1951 southwesterlies strengthened in the northern Indian Ocean.

The wind anomaly displayed weakening of cross equatorial winds (southward anomaly) throughout 1951 monsoon In spite of this weakening of cross equatorial winds there was strengthening in the central Indian ocean in a narrow strip (Fig.3.12). The cross equatorial wind flow strengthened in the central Indian Ocean during 1956 also.

In 1965 (Fig 3.7) all over the Indian Ocean cross equatorial wind weakened. Surprisingly it strengthened in the Bay of Bengal in June - July and again met with a substantial weakening by August.

Similar conditions persisted in June - July 1972 (Fig 3.8). Whether this has resulted in a shift in the rainfall pattern towards northeast India or Myanmar – Thailand region has to be investigated. However, the southwesterlies blowed very weak in Bay of Bengal in 1979 (Fig.3.9) which ultimately resulted in a drought over India. The wind anomalies in 1982 (Fig. 3.10) and 1987 (Fig. 3.11) where similar on the Arabian Sea and equatorial during premonsoon and monsoon period. North easterly wind anomalies (weakening of south westerlies) was very clear in the Arabian sea and the western Indian Ocean.

The major observations can be solicited as

1) In general, a strengthening of south westerlies (of over 3-5m/s) in the Arabian Sea occurs during good monsoon years and a weakening of SW take place during bad monsoon years. However 1983 remains as an exceptional case during early part of the monsoon.

2) Weak cross equatorial flow in the western Indian Ocean in bad

monsoon years compared to good monsoon years.

3) Some bad monsoon years are distinguished with the strengthening of south westerlies in the eastern part of Bay of Bengal (example 1965).

4) The wind anomalies are strongest in the eastern equatorial Indian Ocean.

3.6 Study of Sea Level pressure Anomaly

It has been an established fact that variability of sea level pressure determines the subsidence and convective activity in the atmospheric boundary layer (Mohanthy et al 1994). Sea level pressure is an important parameter governing the convection activity at the air – sea interface. Normally, fair weather /less air-sea exchange is experienced over the high-pressure zones and the vice versa in low - pressure zones. Though there are a number of studies relating SST anomaly in the Indian Ocean and monsoon variability (Shukla, 1987; Rao and Goswami, 1988 and Krishnamurti, 1989a, 1989b) studies relating sea level pressure anomaly in the Indian Ocean and summer monsoon is fragmentary (Mohankumar 1991). Mohanty et al (1983) indicated that prior to the onset of monsoon moisture builds up in the lower atmosphere boundary layer and there is an increase of latent heat flux and moist static energy. Study carried out by Shukla (1987) suggested that above normal monsoon rainfall years are associated with strong pressure gradients, strong winds and drastic cooling of the Arabian Sea in comparison to deficient monsoon years.

To elucidate the impact of Sea level pressure on the performance

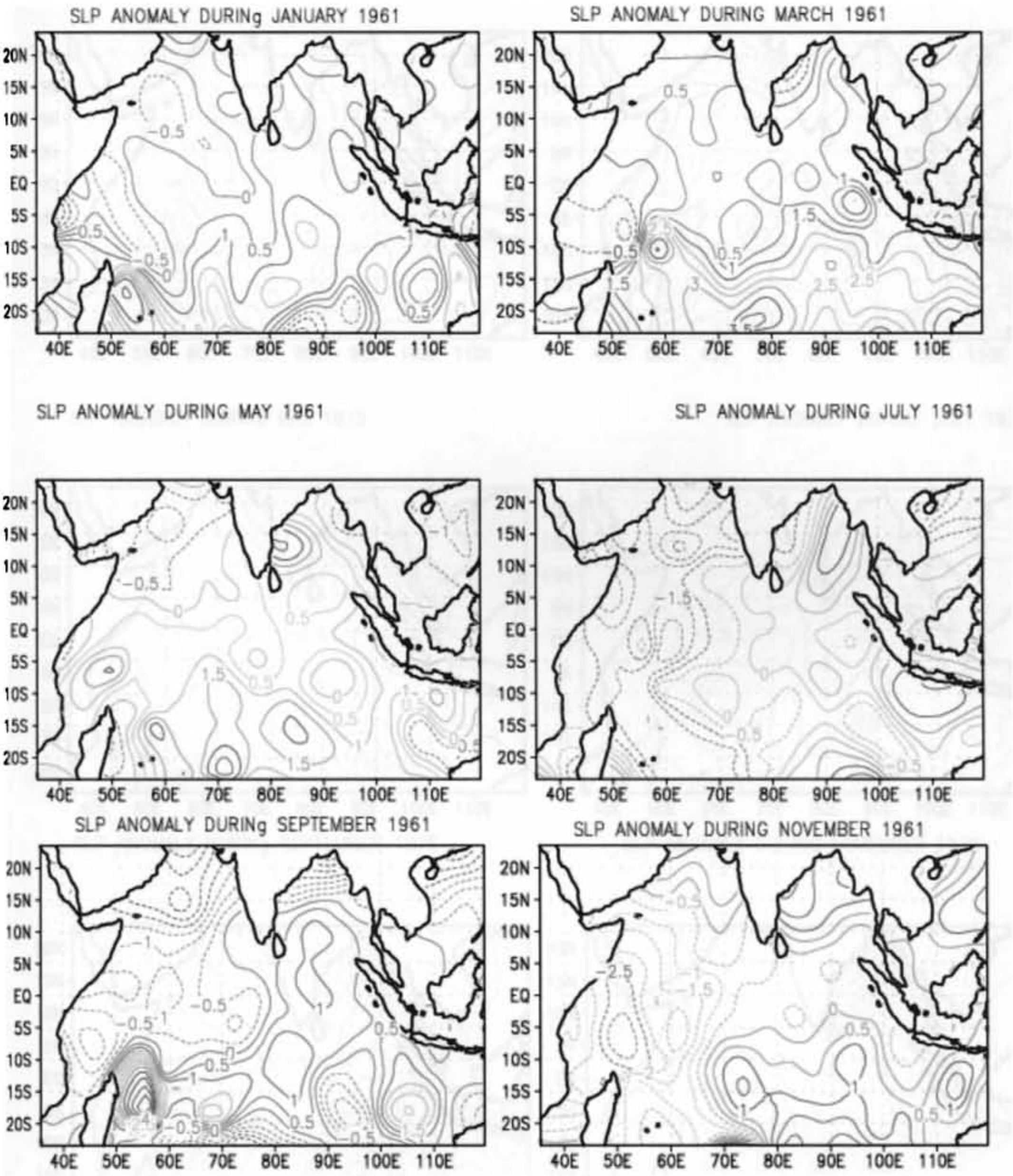


Fig: 3.14 SEA LEVEL PRESSURE (hPa) ANOMALY 1961

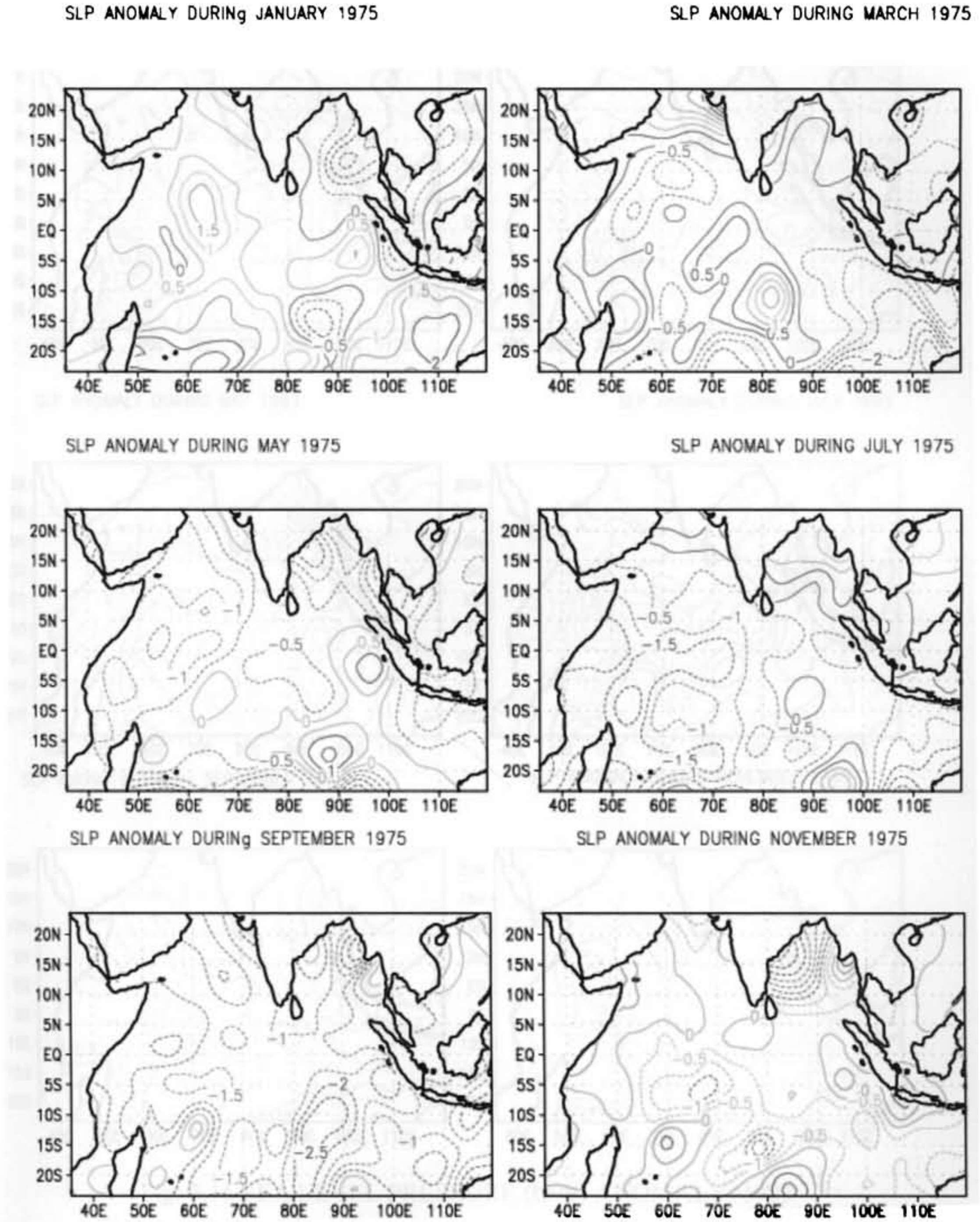


Fig: 3.15 SEA LEVEL PRESSURE (hPa) ANOMALY 1975

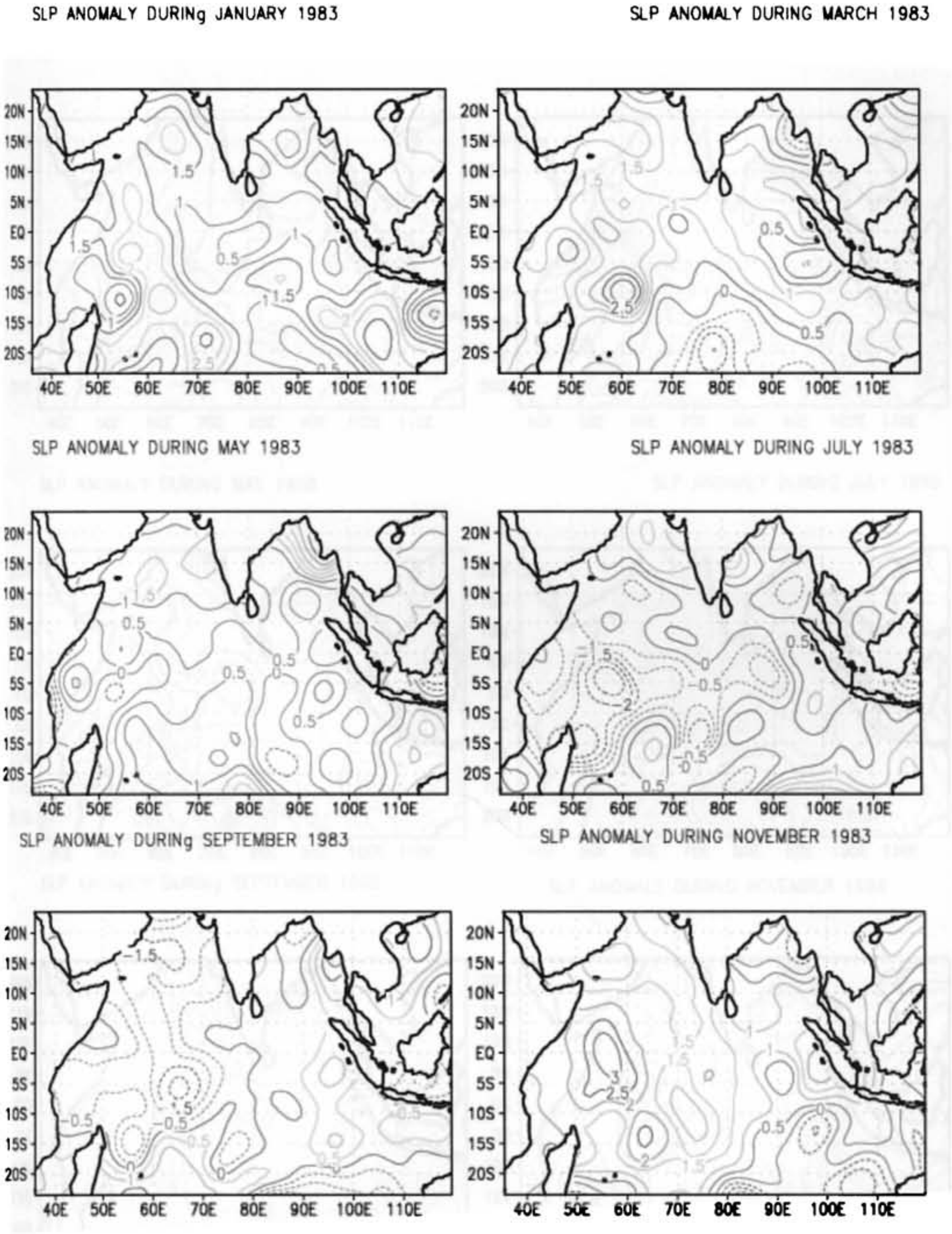


Fig: 3.16 SEA LEVEL PRESSURE (hPa) ANOMALY 1983

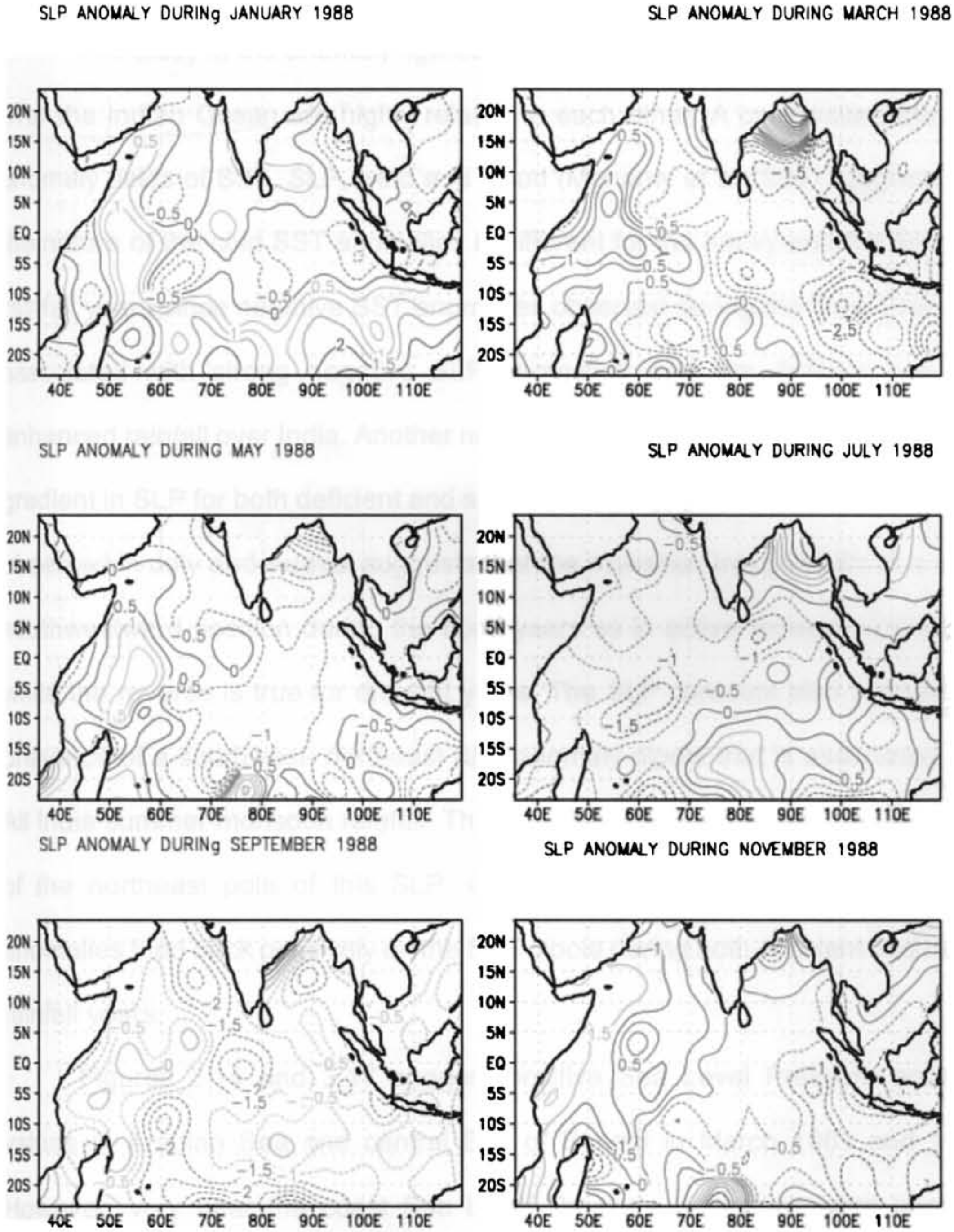


Fig: 3.17 SEA LEVEL PRESSURE (hPa) ANOMALY 1988

of the monsoon activity, its variability under the different monsoon conditions was investigated and the results are presented in the following sections.

The study of the anomaly figures of SST and SLP reveals that parameters over the Indian Ocean are highly related to each other. A composite picture of anomaly fields of SST, SLP, wind and cloud (Mohanty et al 1994) suggests that the nature of the cold SST anomalies is different for the heavy and the deficient rainfall years since negative SST anomalies observed during the flood years are associated with strong negative SLP anomalies over the Arabian sea and enhanced rainfall over India. Another noticeable feature is the existence of strong gradient in SLP for both deficient and surplus rainfall years. SLP anomaly pattern observed in July and August suggests that the monsoon trough shifts to a more southwestward position during the flood years as in active monsoon condition, while the reverse is true for drought years. The SLP contours also suggest the presence of a southwest- northeast SLP anomaly dipole that is associated with All India summer monsoon rainfall. There is also a southeastward displacement of the northeast pole of this SLP. On the other hand, the convergence of anomalies feed back positively on the SST dipole during both deficient and heavy rainfall years.

Figures 3.14 and 3.17 suggest positive Sea Level Pressure anomaly values in Arabian Sea and central Bay of Bengal in March 1961 and 1988. However, very near the coast Sea Level Pressure anomalies were negative suggesting the occurrence of low - pressure systems. However, the entire scenario changed by May. The entire North Indian Ocean with the exception of

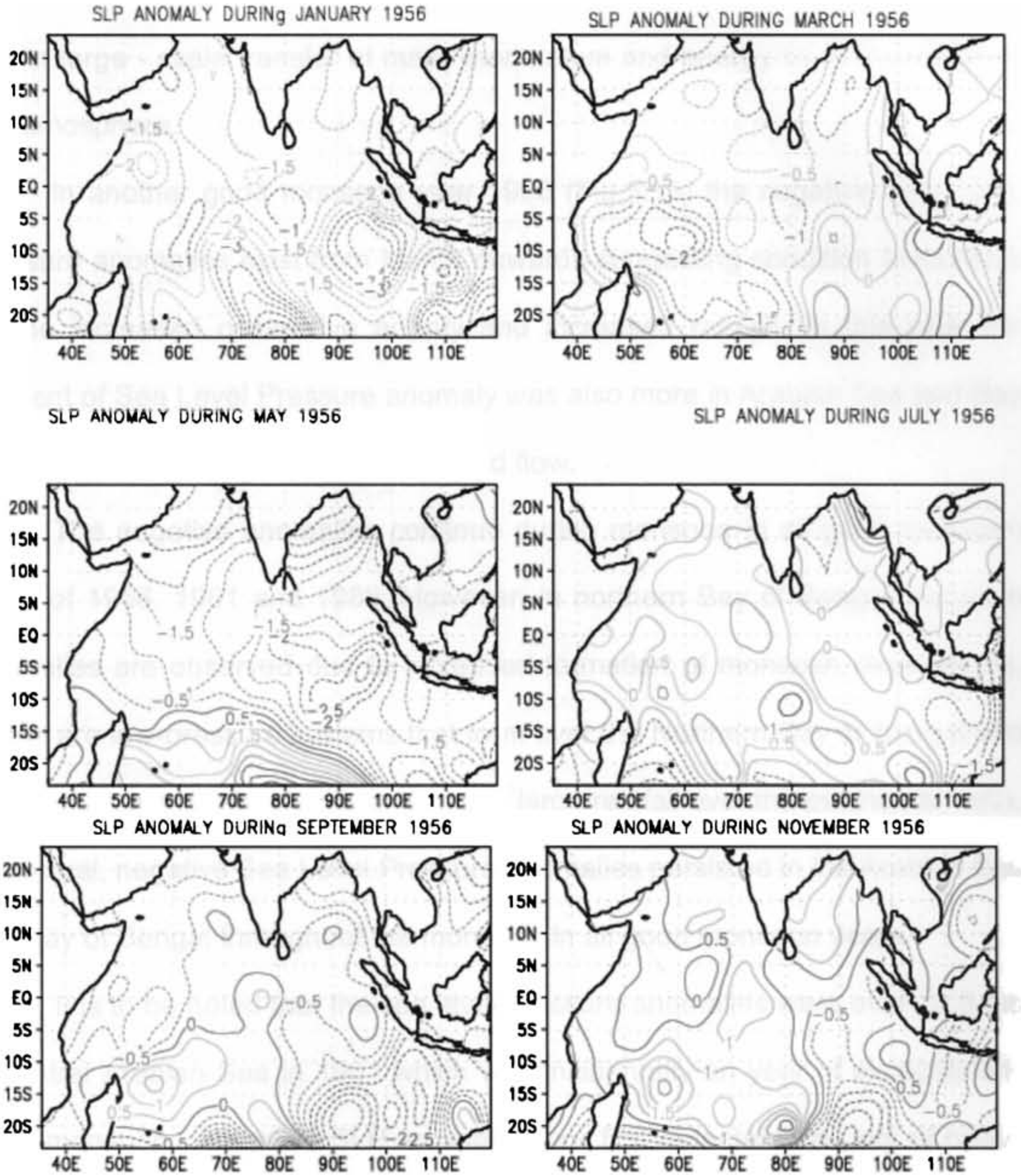


Fig: 3.13 SEA LEVEL PRESSURE (hPa) ANOMALY 1956

Bay of Bengal in 1961 (Fig. 3.14), experienced negative Sea Level Pressure anomalies. These low-pressure anomalies could lead to increased convective activity prior to a good monsoon. As the convective activity increases one can expect large - scale transfer of mass momentum and energy from the ocean to the atmosphere.

In another good monsoon year 1956 (Fig.3.13) the negative Sea Level Pressure anomalies exist from March onwards-suggesting condition favourable for the increased convective activity and increased rainfall. In this year the gradient of Sea Level Pressure anomaly was also more in Arabian Sea and Bay of Bengal, favouring for accelerated wind flow.

The negative anomalies continue during monsoon in all good monsoon years of 1956, 1961 and 1988. However, in northern Bay of Bengal, negative anomalies are observed due to increased formation of monsoon depressions. These are low-pressure systems that form over the Northern Bay of Bengal and move along the monsoon trough bringing large rainfall over the peninsular India. In general, negative Sea Level Pressure anomalies persisted in the Arabian Sea and Bay of Bengal throughout the monsoon in all good monsoon years.

It is to be noted that the sea level pressure anomalies were over –2.5 mb in central Arabian Sea in 1961 which was incidentally an year of exceptionally good monsoon. Negative SLP anomalies are first detectable in July of heavy rainfall years over Arabian Sea and Bay of Bengal close to the peninsula. These negative SLP anomalies progressively grow and expand to cover the entire domain in September of heavy rainfall years. Such broad scale evolutions of

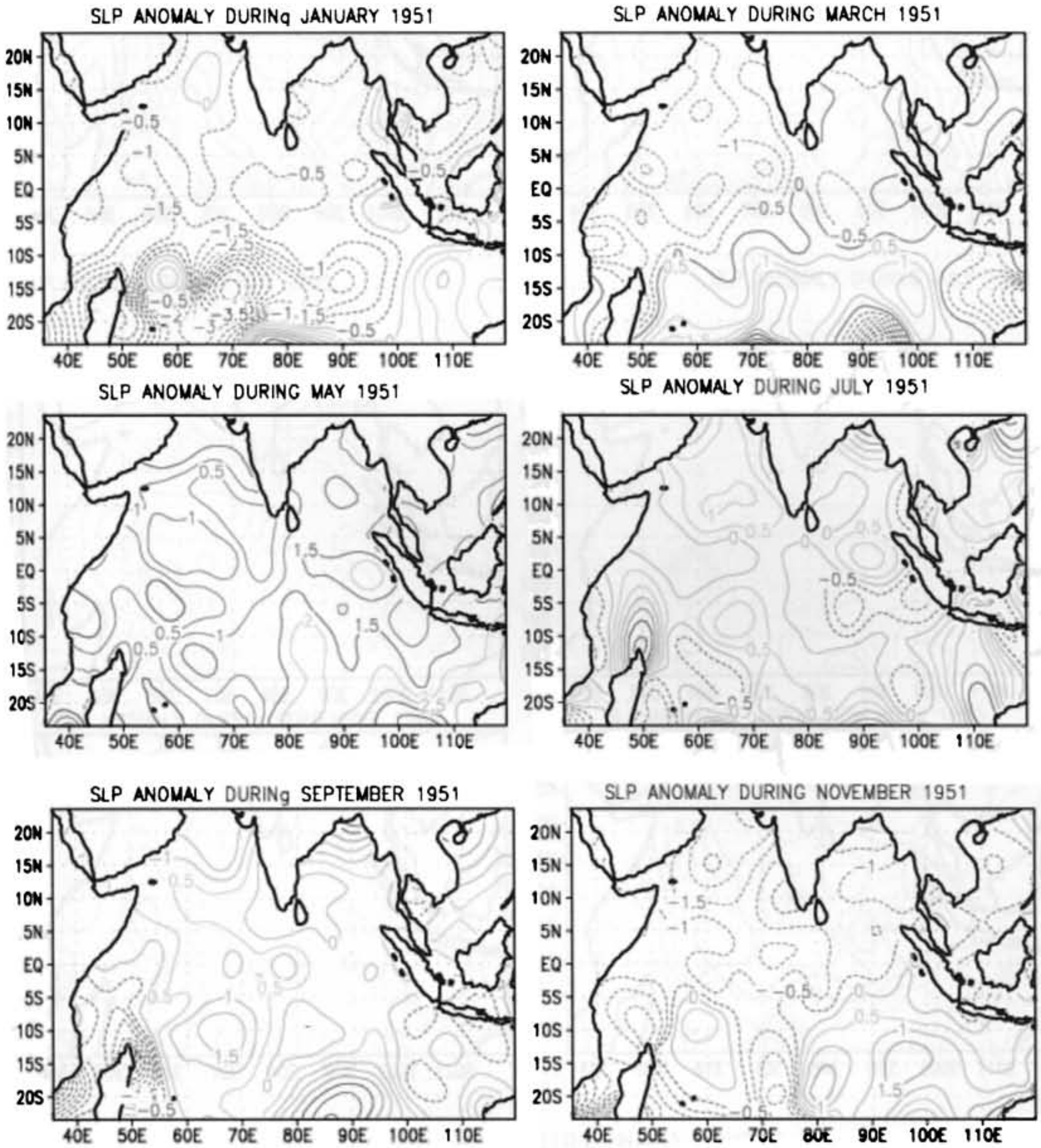


Fig: 3.18 SEA LEVEL PRESSURE (hPa) ANOMALY 1951

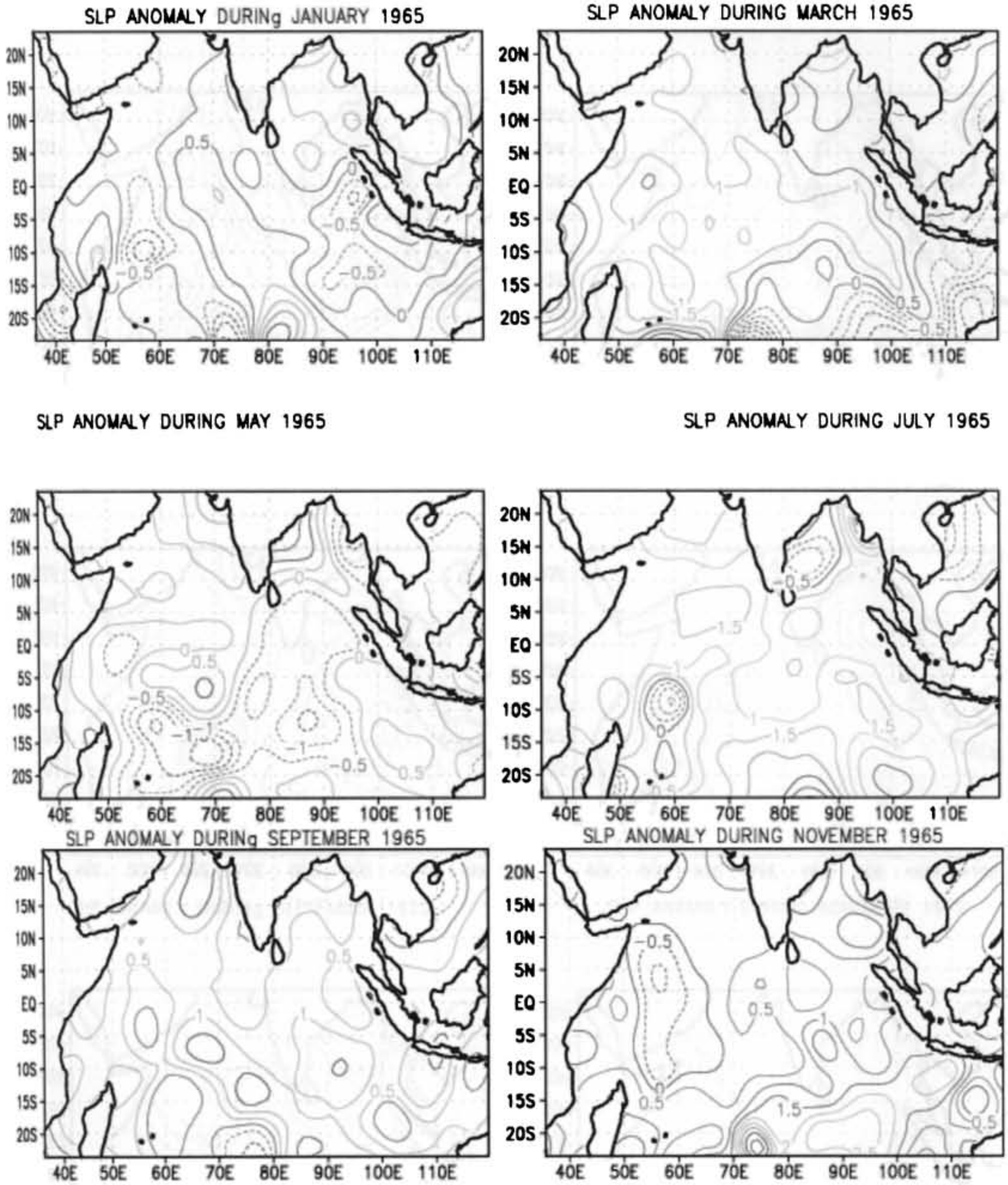


Fig: 3.19 SEA LEVEL PRESSURE (hPa) ANOMALY 1965

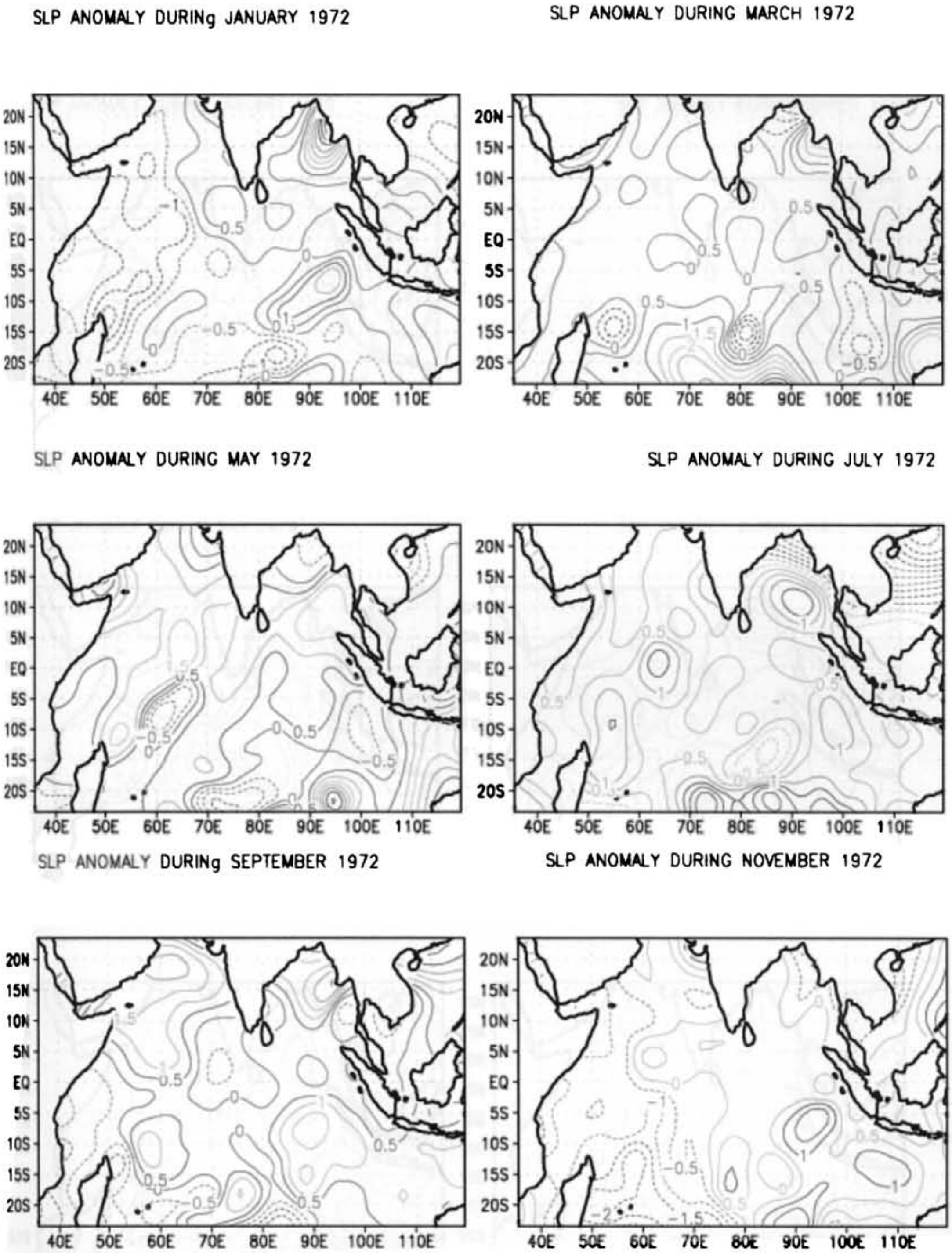


Fig: 3.20 SEA LEVEL PRESSURE (hPa) ANOMALY 1972

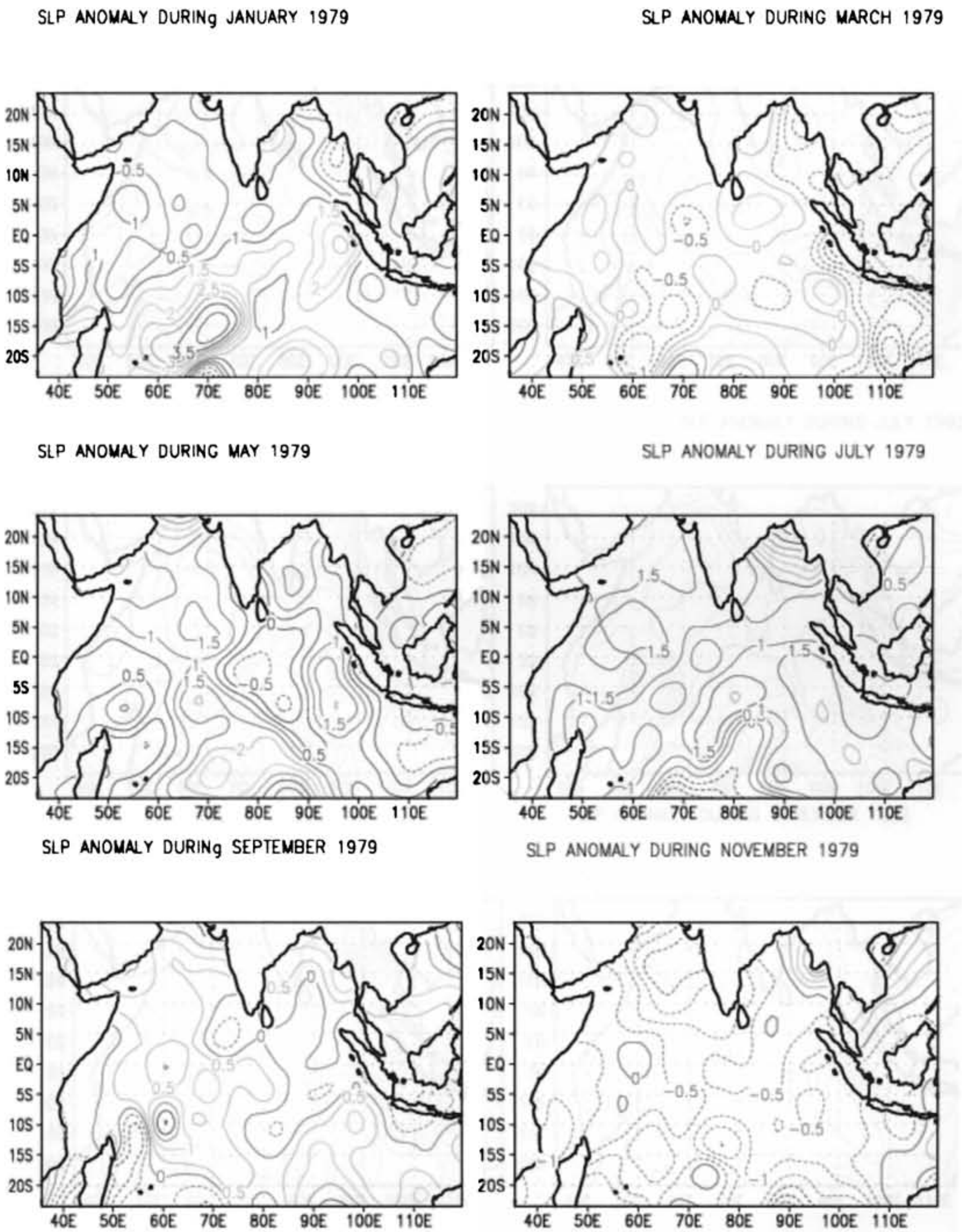
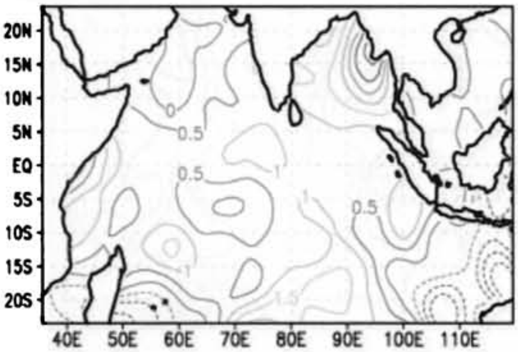
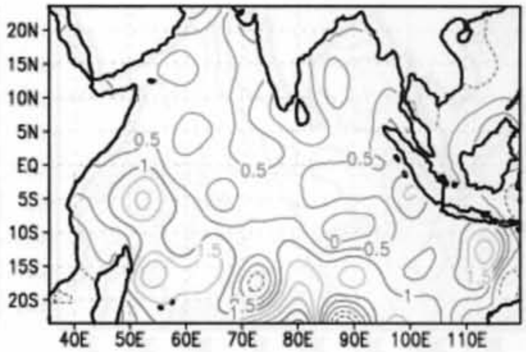


Fig: 3.21 SEA LEVEL PRESSURE (hPa) ANOMALY 1979

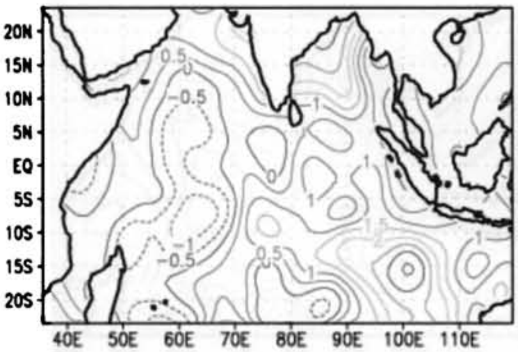
SLP ANOMALY DURINg JANUARY 1982



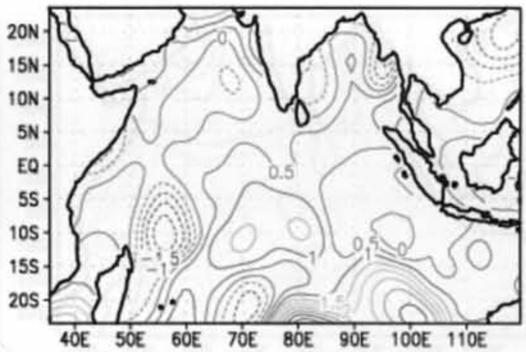
SLP ANOMALY DURING MARCH 1982



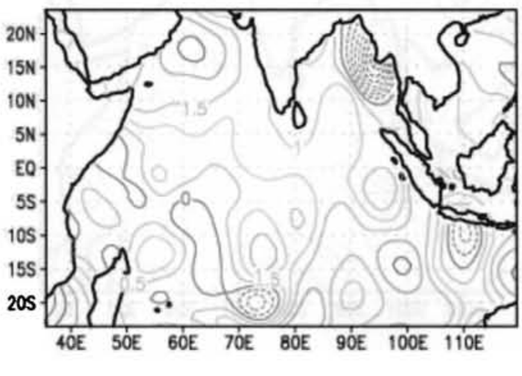
SLP ANOMALY DURING MAY 1982



SLP ANOMALY DURING JULY 1982



SLP ANOMALY DURINg SEPTEMBER 1982



SLP ANOMALY DURING NOVEMBER 1982

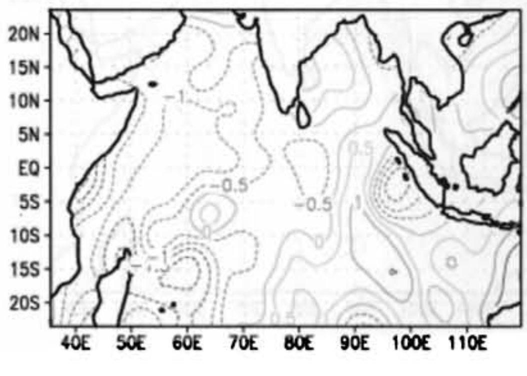


Fig: 3.22 SEA LEVEL PRESSURE (hPa) ANOMALY 1982

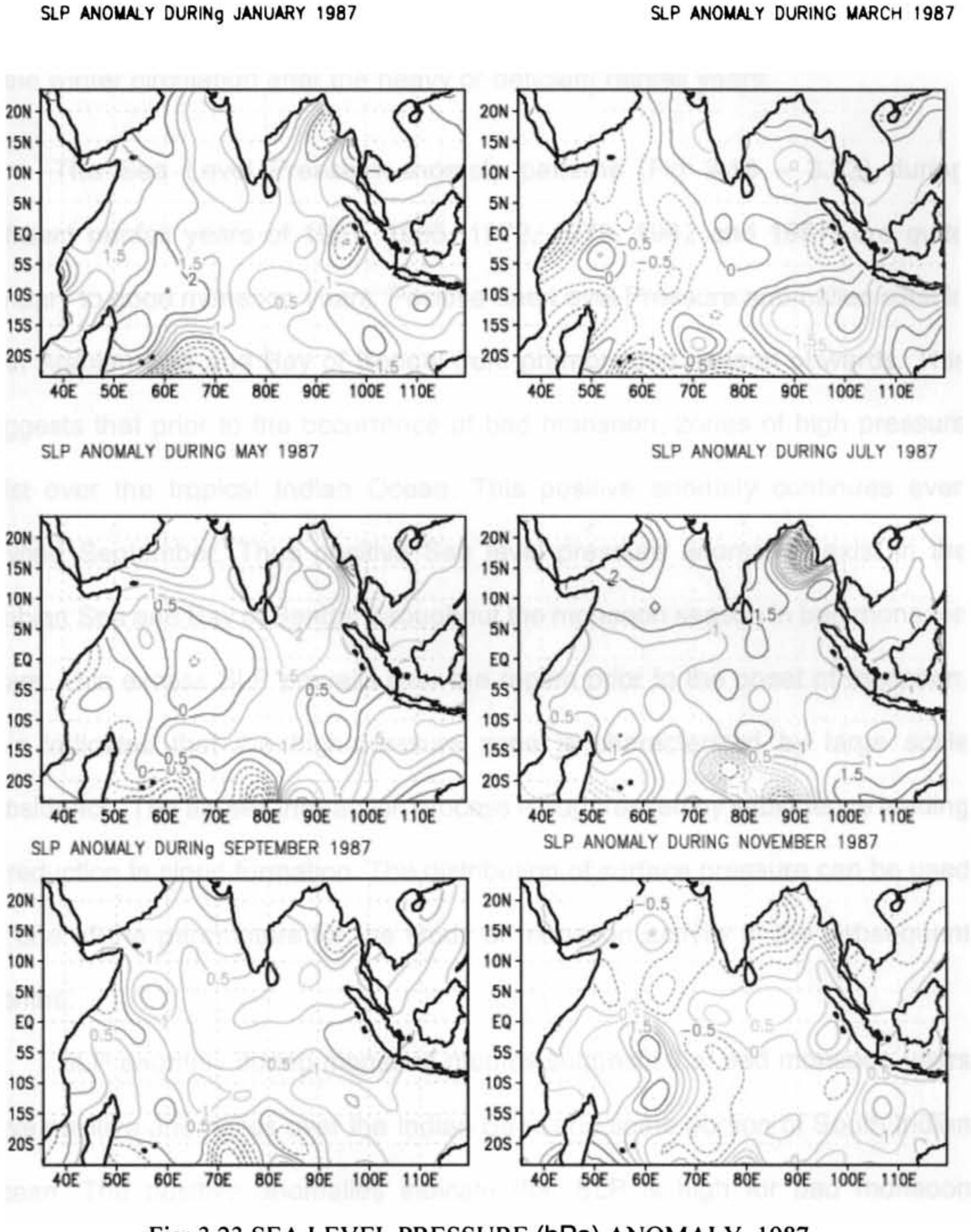


Fig: 3.23 SEA LEVEL PRESSURE (hPa) ANOMALY 1987

broad scale SLP are not apparent for the deficient rainfall years. SLP anomalies after October are not well defined and do not suggest any significant fluctuations of the winter circulation after the heavy or deficient rainfall years.

The Sea Level Pressure anomaly patterns (Fig 3.18 – 3.23) during deficient rainfall years of 1951, 1965, 1972, 1979, 1982 and 1987, are quite contrary to good monsoon years. Positive Sea Level Pressure anomalies exist in both Arabian Sea and Bay of Bengal from premonsoon season onwards. This suggests that prior to the occurrence of bad monsoon, zones of high pressure exist over the tropical Indian Ocean. This positive anomaly continues even beyond September. Thus positive Sea level pressure anomalies exist in the Arabian Sea and Bay of Bengal throughout the monsoon season in bad monsoon years. This excess SLP prevails over the region prior to the onset of monsoon. This indicates that the high-pressure zone is characterized by large scale subsidence. The air-sea interaction process is suppressed by subsidence leading to reduction in cloud formation. The distribution of surface pressure can be used as one of the parameters for the study of monsoon activity in the subsequent months.

SLP anomaly during monsoon months suggests that bad monsoon years have positive anomalies over the Indian Seas and some portion of South Indian Ocean. The positive anomalies indicate that SLP is high for bad monsoon months indicating high-pressure zones. The high -pressure zone suppresses the air-sea interaction and convective activity reducing the moisture convergence in the boundary level.

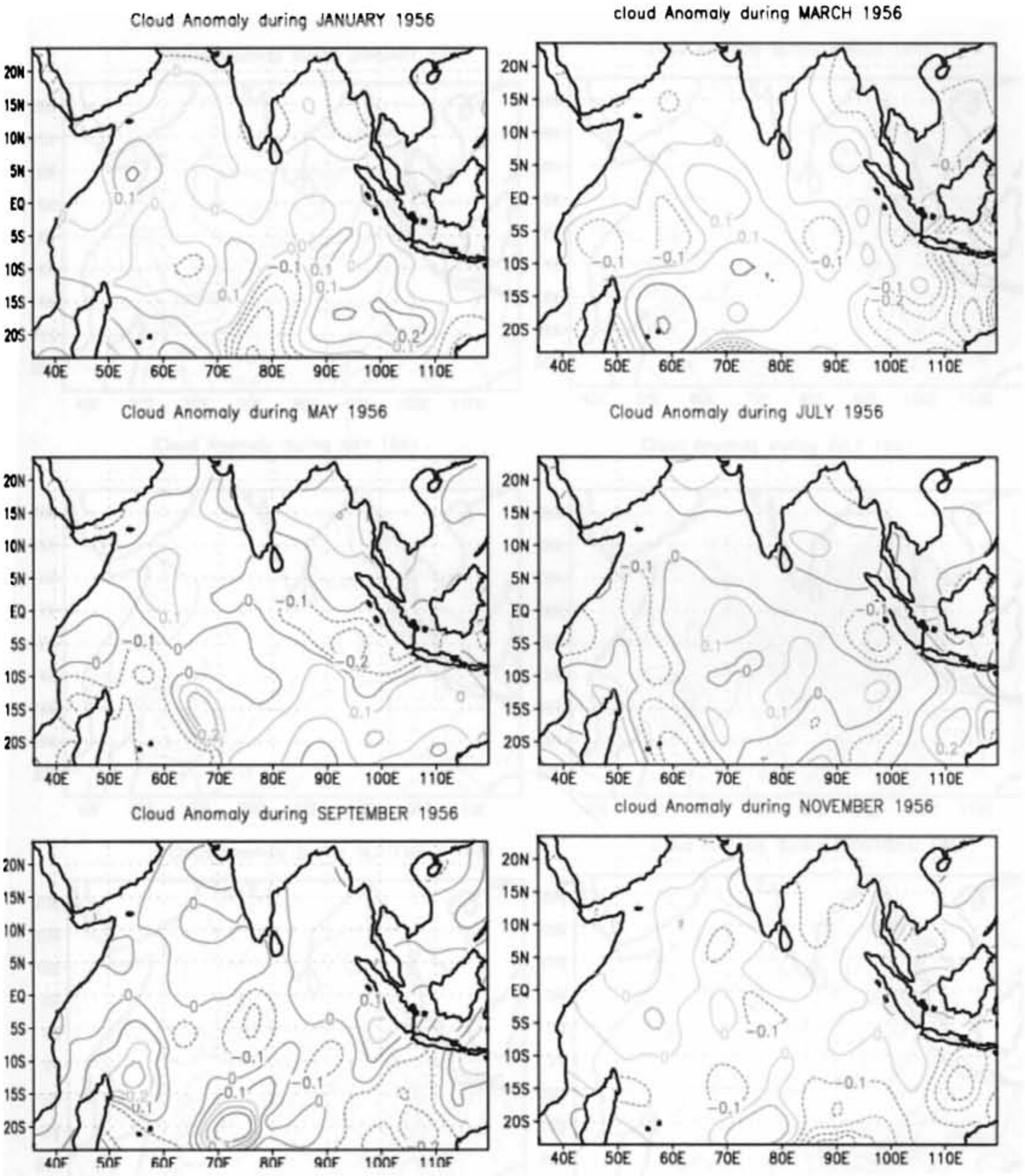


Fig: 3.24 CLOUD COVER (%/100) ANOMALY 1956

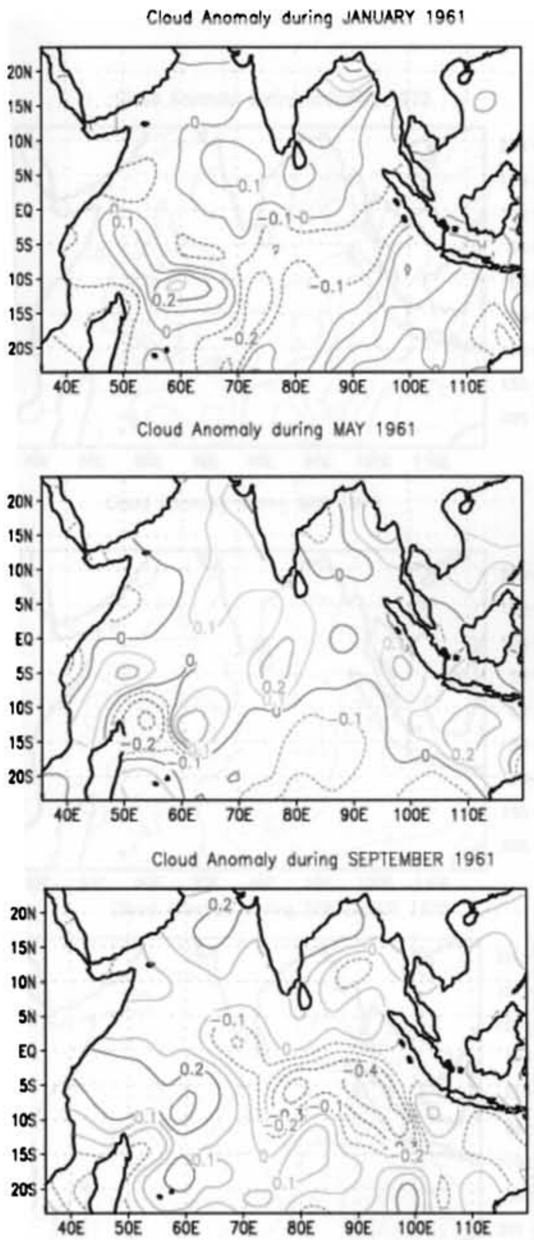


Fig: 3.25 CLOUD COVER (%/100) ANOMALY 1961

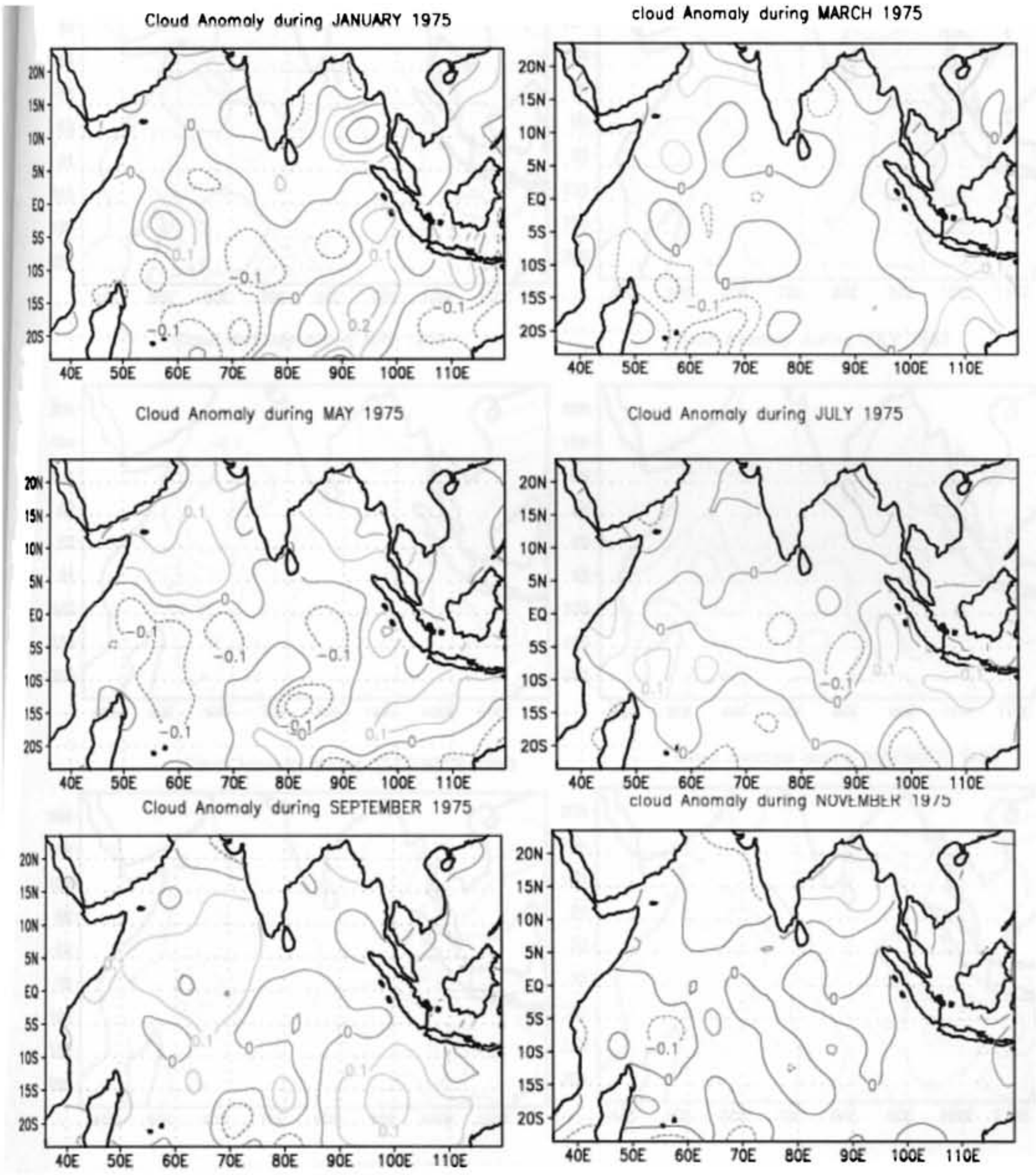


Fig: 3.26 CLOUD COVER (%/100) ANOMALY 1975

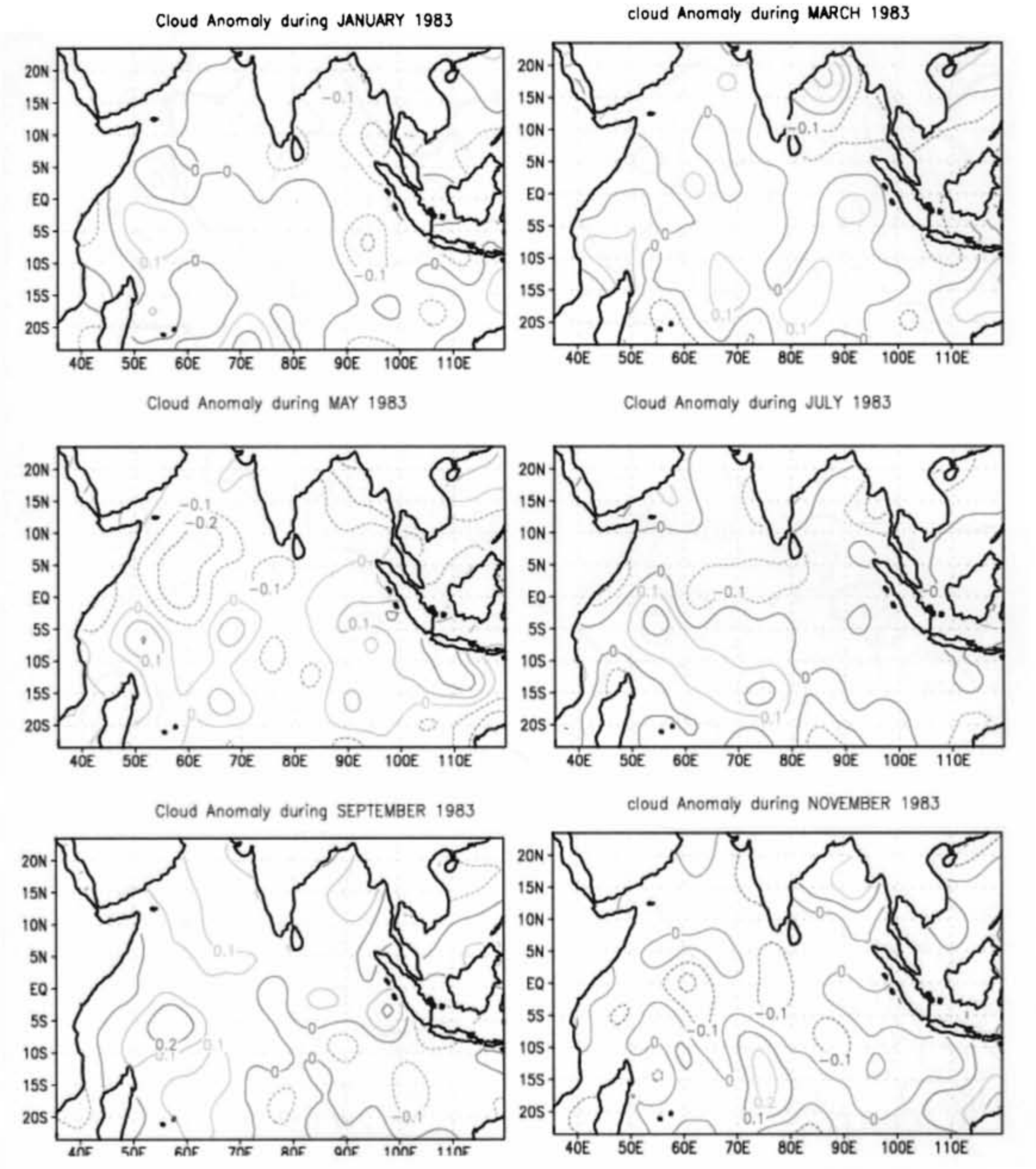


Fig: 3.27 CLOUD COVER (%/100) ANOMALY 1983

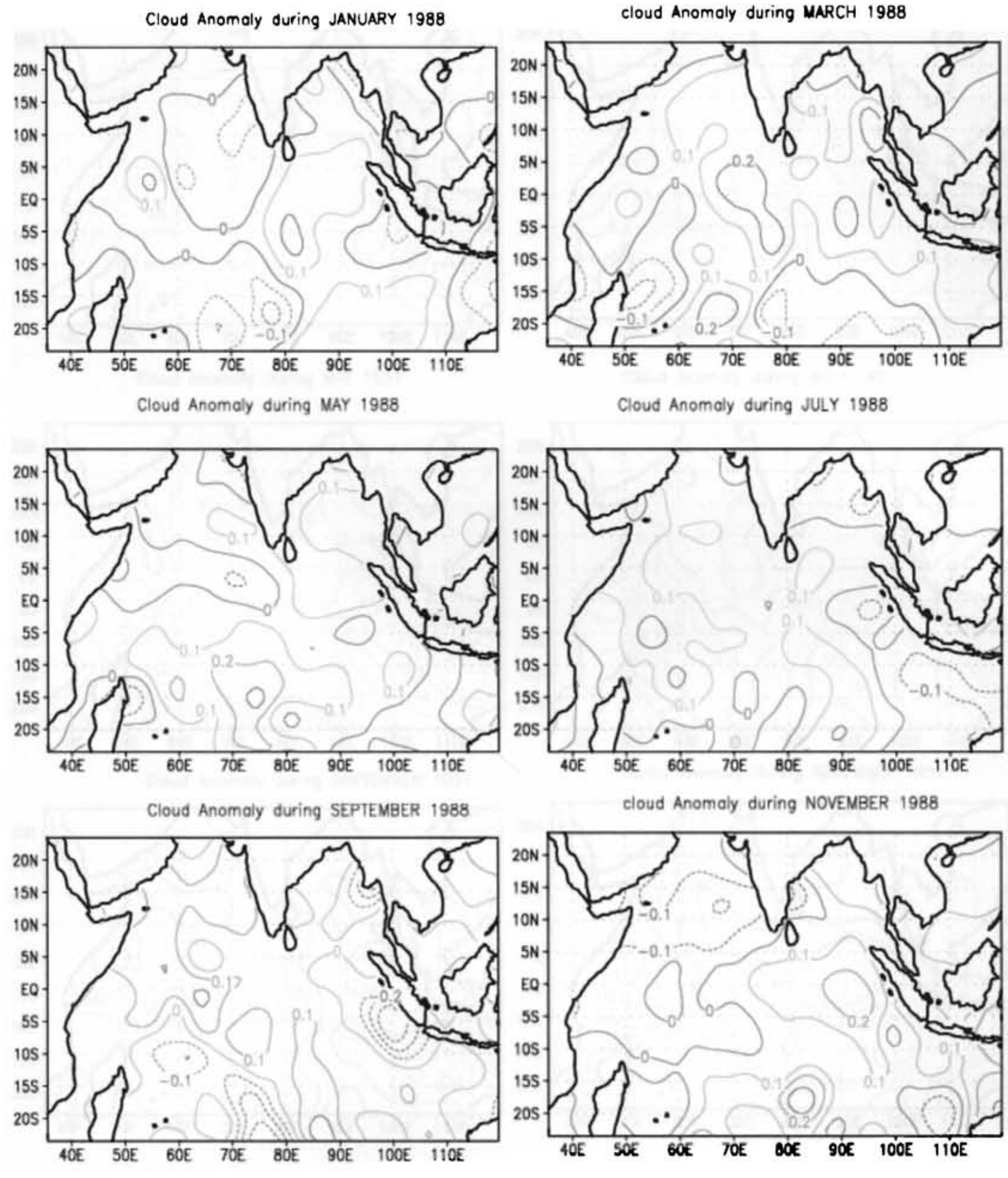


Fig: 3.28 CLOUD COVER (%/100) ANOMALY 1988

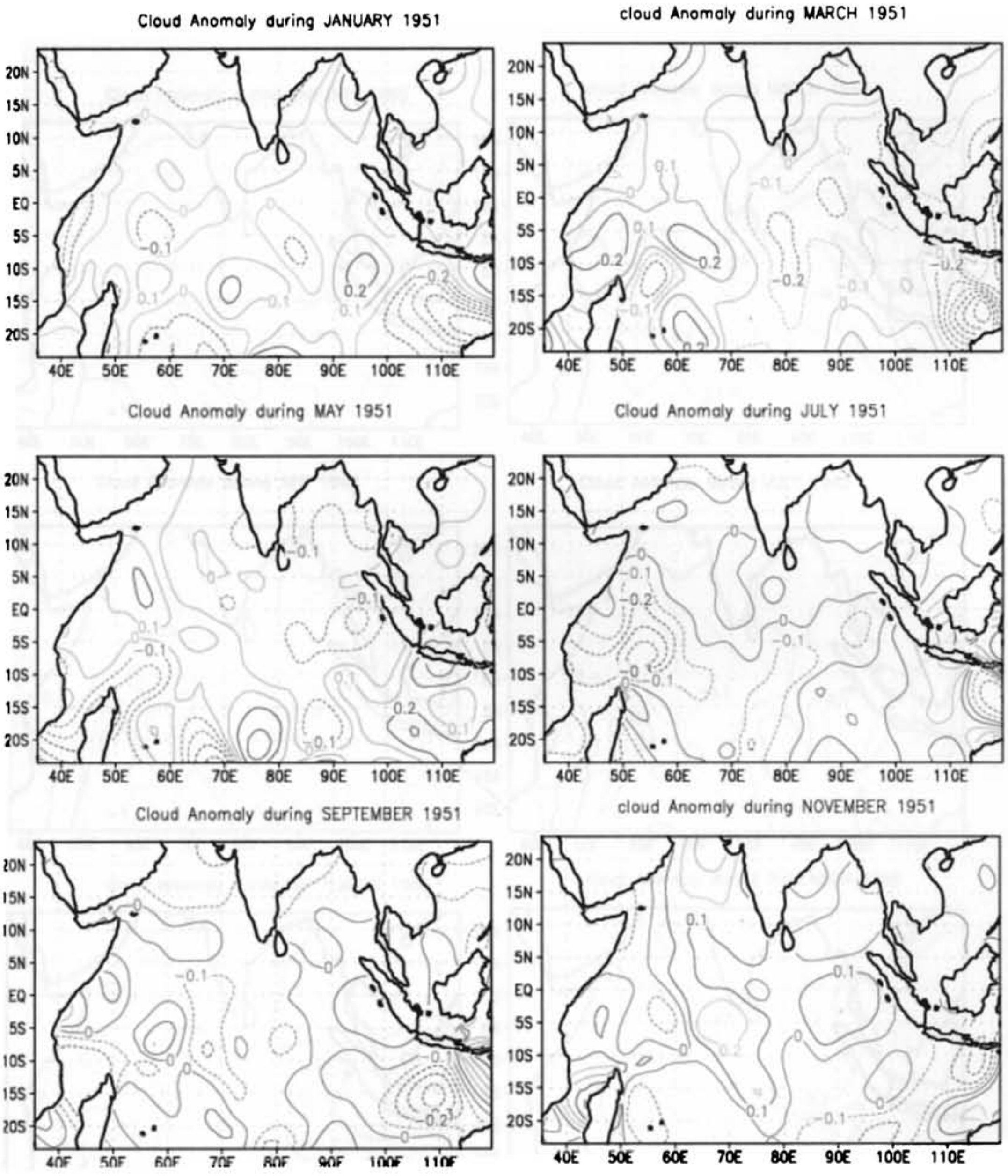


Fig: 3.29 CLOUD COVER (%/100) ANOMALY 1951

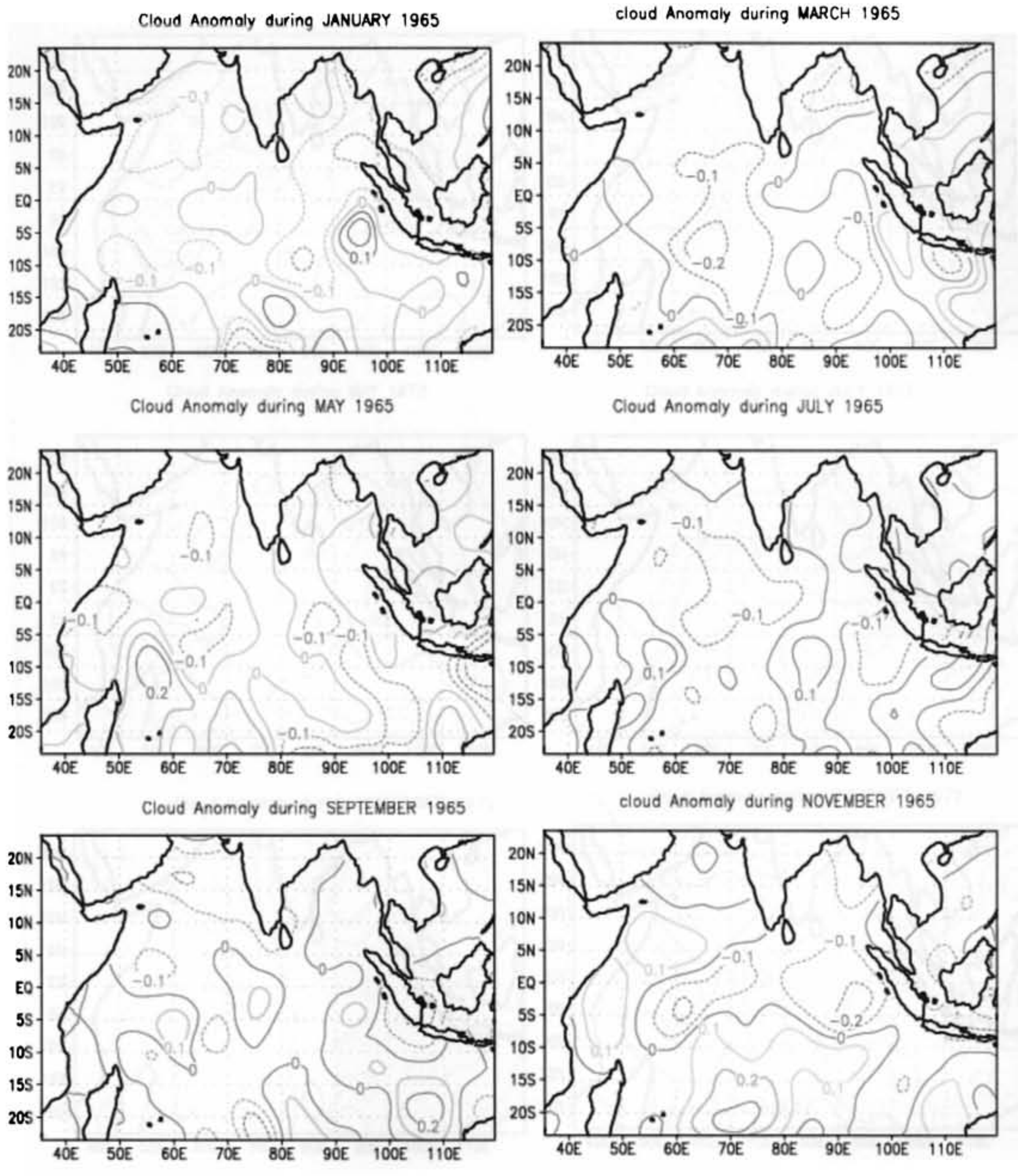


Fig: 3.30 CLOUD COVER (%/100) ANOMALY 1965

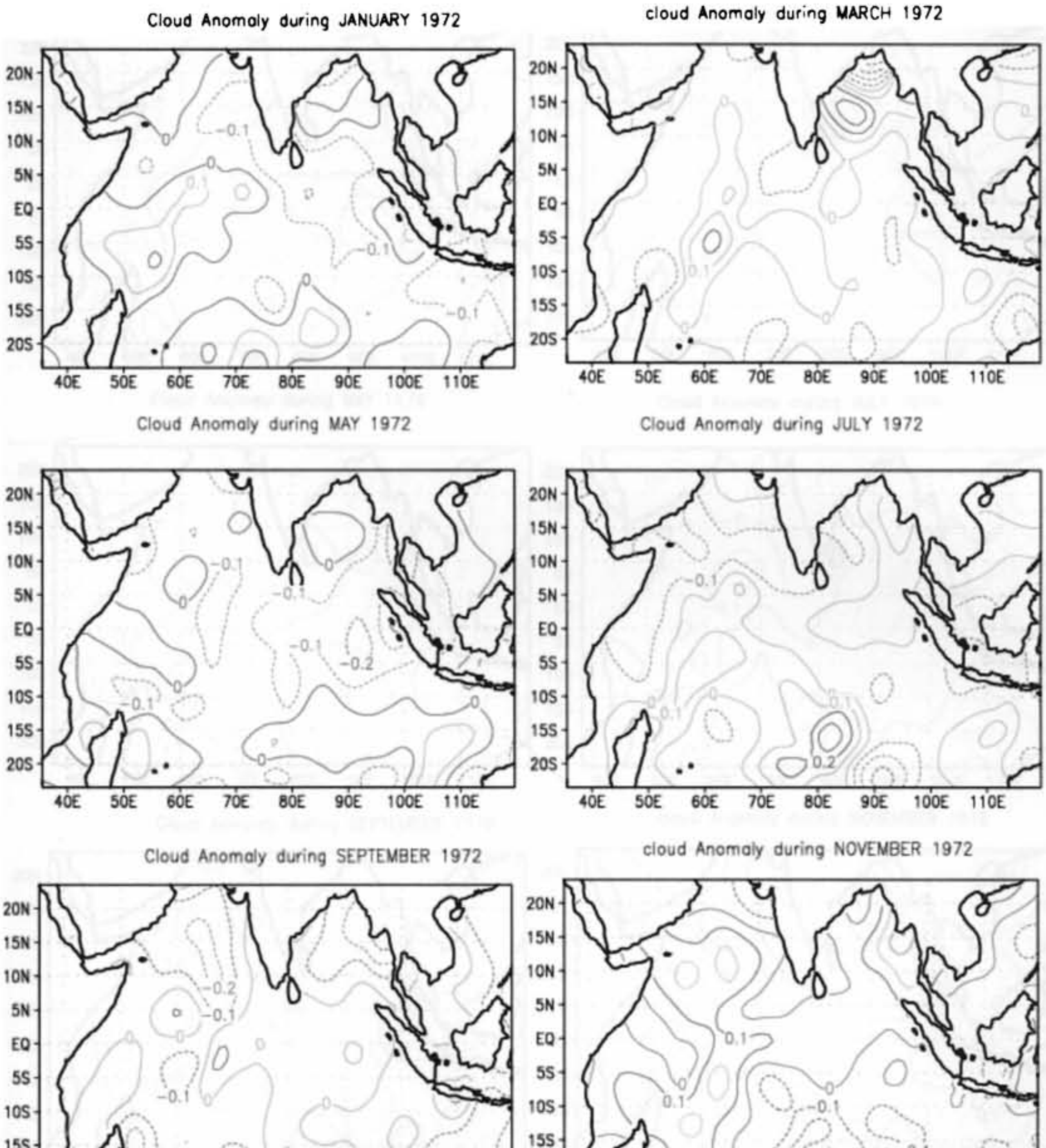


Fig: 3.31 CLOUD COVER (%/100) ANOMALY 1972

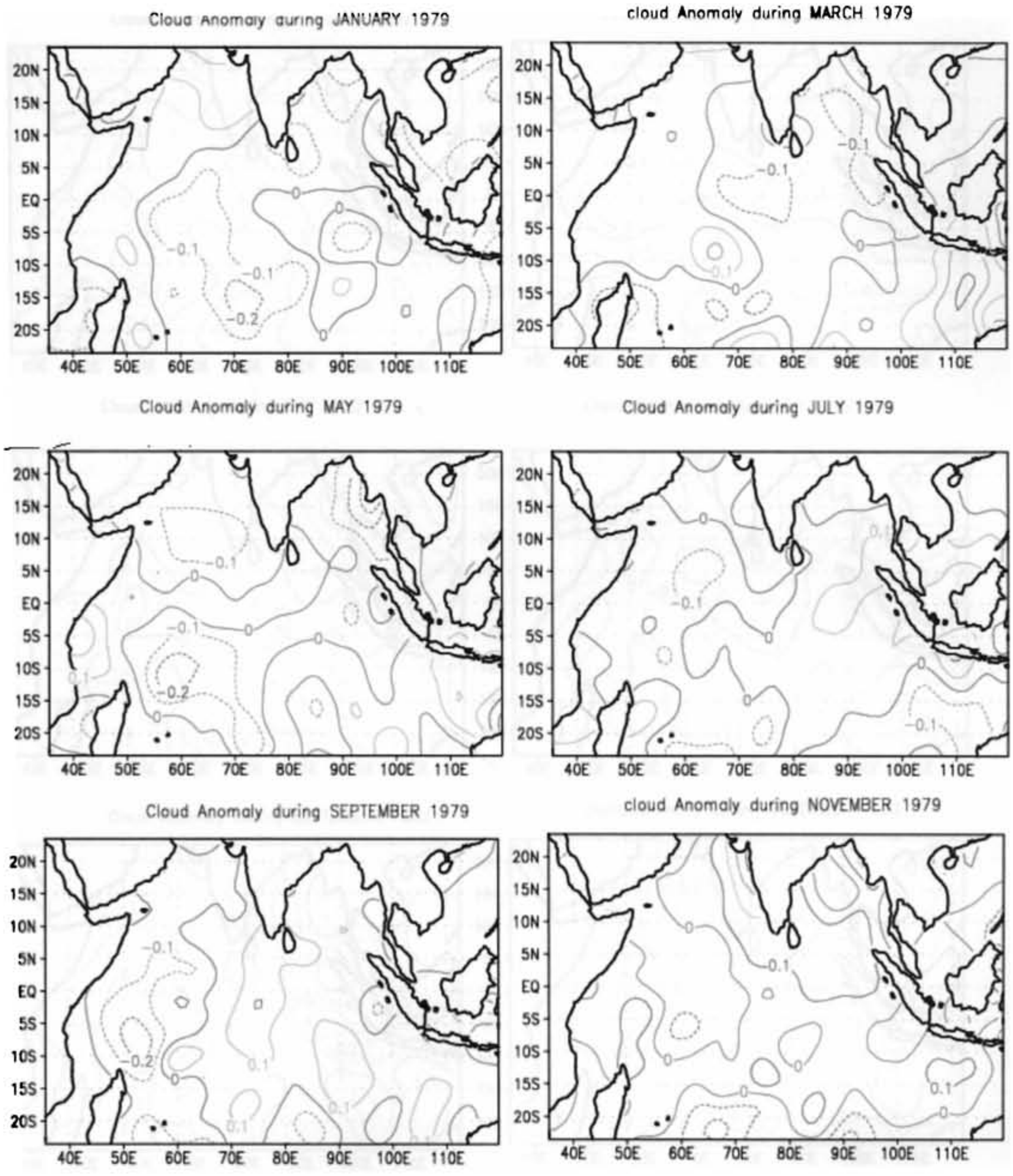


Fig: 3.32 CLOUD COVER (%/100) ANOMALY 1979

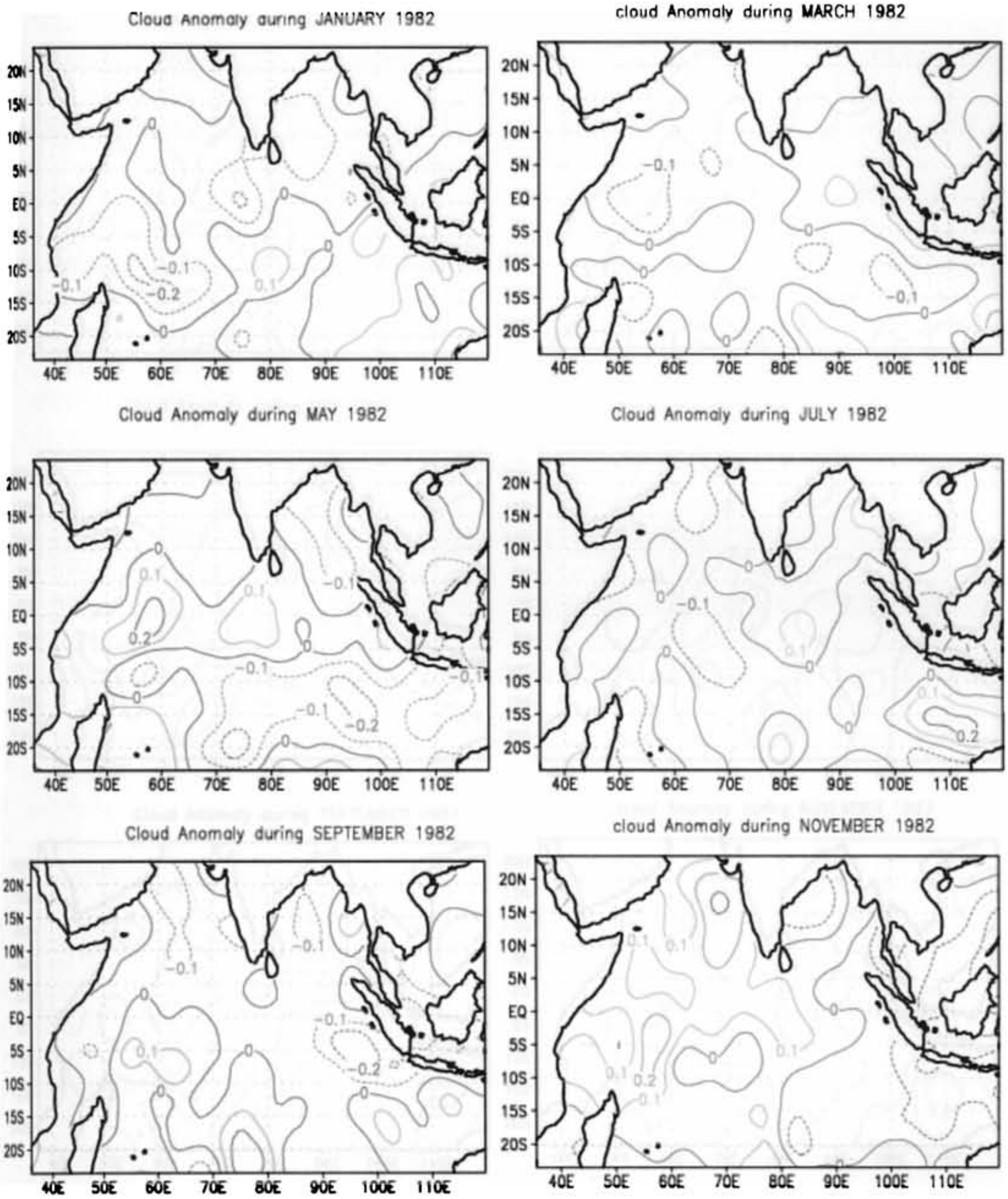


Fig: 3.33 CLOUD COVER (%/100) ANOMALY 1982

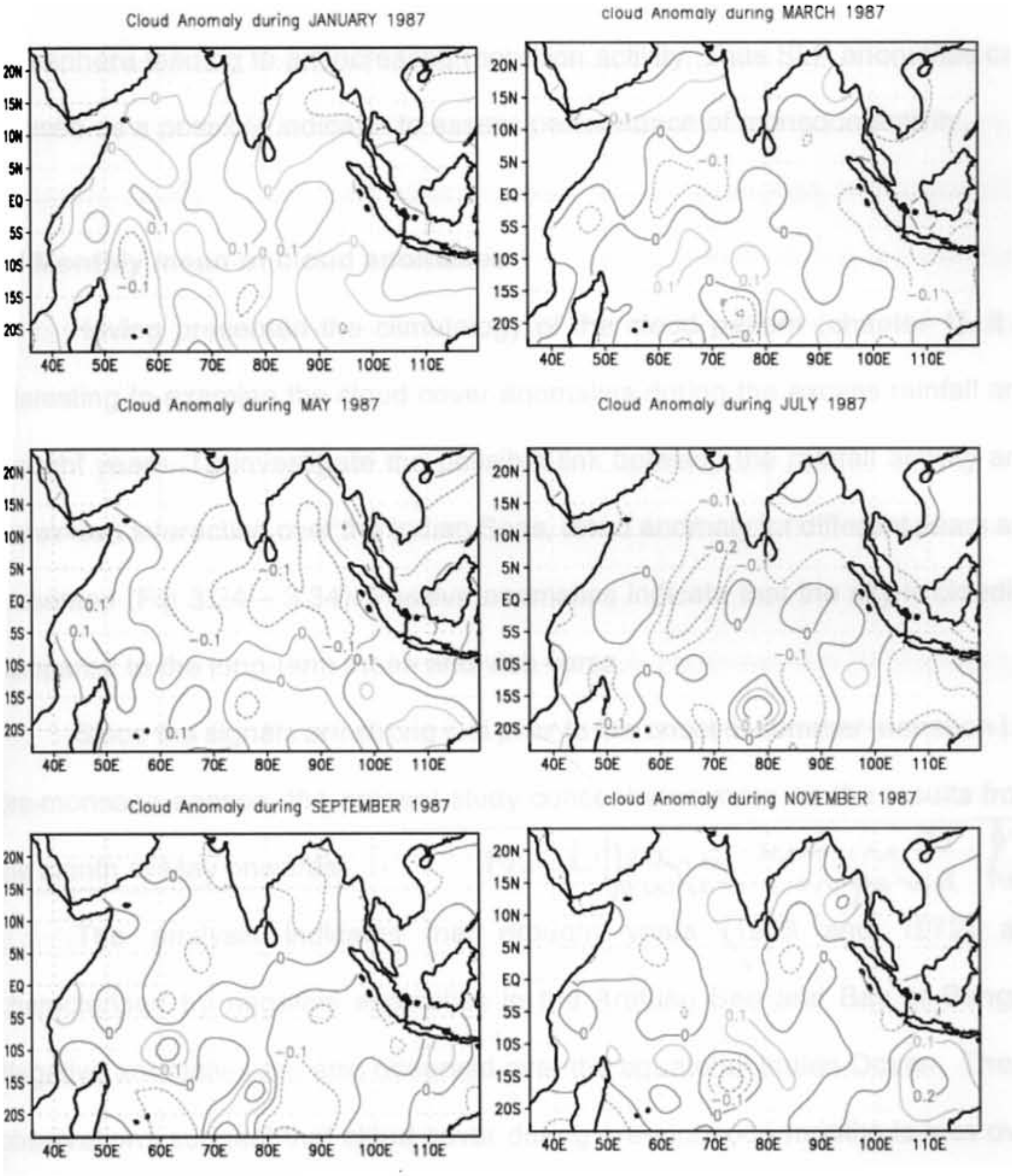


Fig: 3.34 CLOUD COVER (%/100) ANOMALY 1987

However, during good monsoon years, the pressure anomalies are negative indicating a fall in SLP during pre monsoon and monsoon months. The condition is favourable for the updraft and moisture convergence in the atmosphere leading to an increased monsoon activity. Thus SLP anomalies can be used as a possible indicator to assess performance of monsoon activity.

3.7 Monthly mean of cloud anomalies

Having presented the climatology of the cloud pattern (chapter 1), it is interesting to examine the cloud cover anomalies during the excess rainfall and drought years. To investigate the possible link between the rainfall activity and the air-sea interaction over the Indian Seas, cloud anomaly for different years are presented (Fig 3.24 – 3.34). Positive anomalies indicate that the sky is cloudier compared to the long-term mean and vice versa.

Since the signals are strong just prior to the onset of summer monsoon i.e. pre-monsoon season, the present study concentrates more on the results from the month of May onwards.

No difference seen in the prese[illegible] when compared to the [illegible]

The analysis indicates that drought years (1965 and 1972) are characterised by negative anomalies in the Arabian Sea and Bay of Bengal. Negative anomalies are also observed over the equatorial Indian Ocean. These observations suggest that cloud cover during pre-monsoon months is less over the Indian Seas in drought years. The relatively clear sky conditions are due to reduced air-sea interaction and convective activities over the Indian Seas prior to a reduced monsoon activity.

However, the pre-monsoon periods of excess rainfall years (1956, 1961, 1975, 1983 and 1988) indicate a contrasting picture (Fig.3.24 – 3.28) compared to the drought years. The cloud anomalies presented in above figures clearly indicate positive values over the Arabian Sea, Bay of Bengal and equatorial Indian Ocean. This analysis brings out an interesting result that prior to a good monsoon, sky is cloudier indicating vigorous air-sea exchange and convective activity over the oceans as compared to a bad monsoon. Since the moisture starts building up in the atmospheric boundary layer during pre-monsoon season increased cloud cover is an indication of this process.

For instance, the drought monsoon years (Fig.3.29 – 3.34) are characterised by the negative or near zero cloud anomalies in the Arabian Sea and Bay of Bengal during the monsoon season (June-September). The outcome is the reduced rainfall activity over the Indian sub continent due to reduction in rain bearing clouds. This result tells us that during the monsoon season (June to September) in drought years convective activity is reduced. The cloud anomaly picture for the monsoon season during excess rainfall years indicates a different picture. Most of the Arabian Sea, Bay of Bengal and tropical Indian Ocean are characterised by positive cloud anomalies indicating enhanced convective activity of these regions. This process in turn will enhance the monsoon activity over the Indian sub continent. The study revealed that in pre-monsoon season and monsoon season during excessive rainfall years the cloud cover is more in the equatorial and North Indian Ocean compared to same period of drought years.

3.8 Indian Ocean Dipole and Zonal Wind Anomaly

The Indian Ocean Dipole was first mapped using on Dipole Mode Index (DMI) based on monthly SST anomalies in the western and eastern tropical Indian Ocean, associated with the SST anomalies in the sea level pressure, cloud cover and wind anomalies. In fact, the monthly SST anomalies in the Indian Ocean are too small and often within the uncertainties in the data involved. Hence only three major positive dipole years are recorded in the last four decades i.e. 1961, 1994 and 1997 though there are indications of this in several other years (1972, 1982, 1987 etc.) associated with IOD, anomalies occur in surface wind sea level pressure and cloud cover. The anomalies of various surface meteorological parameters used in this study have shown that the amplitude of anomaly is maximum in the wind field especially in the zonal wind field in the region just south of the equator in the eastern Indian Ocean. Hence based on this study a slightly different index of Indian Ocean Dipole is proposed here. For this purpose the zonal wind anomaly in the region 0- 10° S, 85- 95°E is averaged for each month and taken as a single value for that month where the anomaly is strongest. The zonal wind anomaly for each month for the period 1945-1993 is computed. Since our interest is limited to IOD events, the time series of zonal wind anomaly is subjected to Buttenworth Filter to remove periodicities upto the annual cycle. The resultant values are presented in figure 3.35.

The filtered zonal wind anomaly (Fig. 3.35) shows large amplitude fluctuations, which were maximum prior to 1970, and thereafter the amplitude diminished. The large negative zonal wind anomalies are observed towards the later half of 1951, 1957, 1961, 1963, 1965, 1967, 1972, 1982 and 1987 while significant positive anomalies are noticed in 1947, 1950, 1955, 1958, 1962, 1964, 1970, 1975 and 1981. It is not clear, why the zonal wind anomaly was particularly strong in 1960's and weak in 1980's.

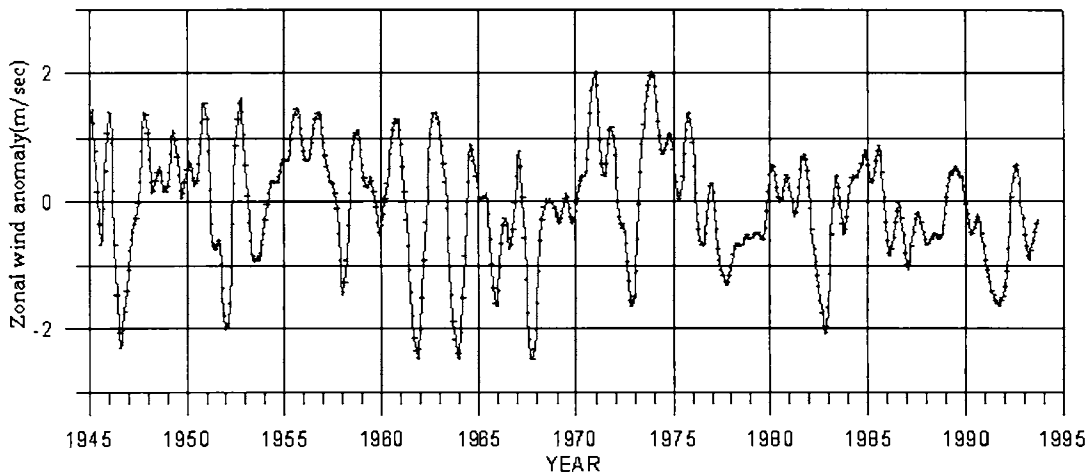


Fig. 3.35 Zonal wind anomaly averaged over 0 - 10°S, 85 - 95°E between 1945 –1993

The zonal wind anomaly became much larger in the 1990's again especially in 1994 and 1997 (Yamagat et al 2001). FFT analysis suggested a periodicity of about 2.5 years in the zonal wind anomaly, which is close to the quasi biennial oscillation (QBO). However, it is not clear whether IOD events are manifestation of QBO itself or not and also the reasons for higher amplitude in some years. It is also not clear why it has maximum amplitude in the east tropical Indian Ocean compared to other regions.

CHAPTER IV

ANOMALIES OF LATENT HEAT FLUX, SHORT WAVE RADIATION, NET HEAT FLUX AND EVAPORATION OVER THE INDIAN OCEAN

4.1 Introduction

There has been an increased awareness of air-sea flux studies owing to their application in climate diagnostic studies. A better understanding of air-sea interaction processes over the tropical Indian Seas on a synoptic and climatic time scale is required to study the variability of Indian summer monsoon. The tropical oceans act as reservoir of heat and supply the necessary energy for the large-scale monsoon circulation. The tropical Indian Ocean is characterised by intense air-sea exchange especially during the summer monsoon season. Studies carried out by a number of authors (Pisharoty, 1965; Mohanty et al, 1983) have established the linkage between the monsoon activity over the Indian subcontinent and the air-sea fluxes of heat and moisture. These fluxes are used as the lower boundary conditions to drive large-scale ocean-atmosphere coupled models (Simonot and Le Treut, 1990). Many studies have established a possible link between sea surface temperature variations over the Indian seas and summer monsoon activity (Shukla, 1987, Rao and Goswamy, 1988), though they are inconclusive.

The studies on variability of fluxes of heat and moisture over the tropical Indian Seas have assumed considerable importance in recent years. However, most of these studies are restricted to specific regions due to the constraints in the availability of data over remote oceanic regions. The oceanic heat budget studies over the tropical Indian Ocean are limited due to this fact. With the advent of satellites there has been a spurt in the surface data collection over the vast oceanic regions.

4.2 Heat budget estimates over the Indian Ocean

Studies of heat budget for the tropical Indian Ocean are few due to the inadequacy of the data. Wyrtki (1971) published an oceanographic atlas exclusively covering equatorial Indian Ocean. Hastenrath and Lamb (1979 a, b) carried out a comprehensive study of the variability of surface marine meteorological parameters and oceanic heat budget components. Their study made use of monthly averages for 60 years (1911-1970) of surface marine met observations over the tropical Indian Ocean with a spatial resolution of 1^{o} X 1^{o}. Bunker's unpublished data sets also deal with the variability of surface marine met and oceanic heat budget estimates for the tropical Indian Ocean for the period 1948- 1972. Though few studies dealing the oceanic heat budget variability over the tropical Indian Ocean are available, most of them do not examine the linkage between air-sea fluxes and summer monsoon activity.

The present study is undertaken with the objective of investigating the linkage between air-sea fluxes and monsoon forcing. Since the monsoon activity is linked to fluxes, the variability of surface marine met fields under the variable monsoon conditions is also studied. To study the variability, individual years are categorised into two groups viz. deficient rainfall years (drought years) and excess rainfall years (flood years) based on the rainfall activity over the Indian subcontinent in these years. A number of researchers have made attempts to correlate the surface marine met field especially sea surface temperature (SST) over the tropical Indian Ocean with the monsoon activity (Pisharoty, 1965; Shukla, 1987; Rao and Goswamy, 1988). It is seen that none of these studies could concretely establish the link between SST and the monsoon activity as conflicting results were obtained.

In this section, the variability of monthly anomalies of surface marine meteorological parameters and the components of oceanic heat budget are studied for typical excess rainfall years (flood years) and deficient rainfall years (drought years). The parameters analysed are anomalies of net heat flux, short wave radiation, latent heat flux and evaporation. The period chosen essentially covers the pre-monsoon, summer monsoon and post monsoon regimes when the air-sea interaction processes are expected to be different. Our aim is to investigate the influence of oceanic heat budget components on the inter-annual variability of the summer monsoon.

Before we discuss the variability of the anomaly field, the climatologic fields are presented to get an overall picture and to understand the monthly evolution of different parameters.

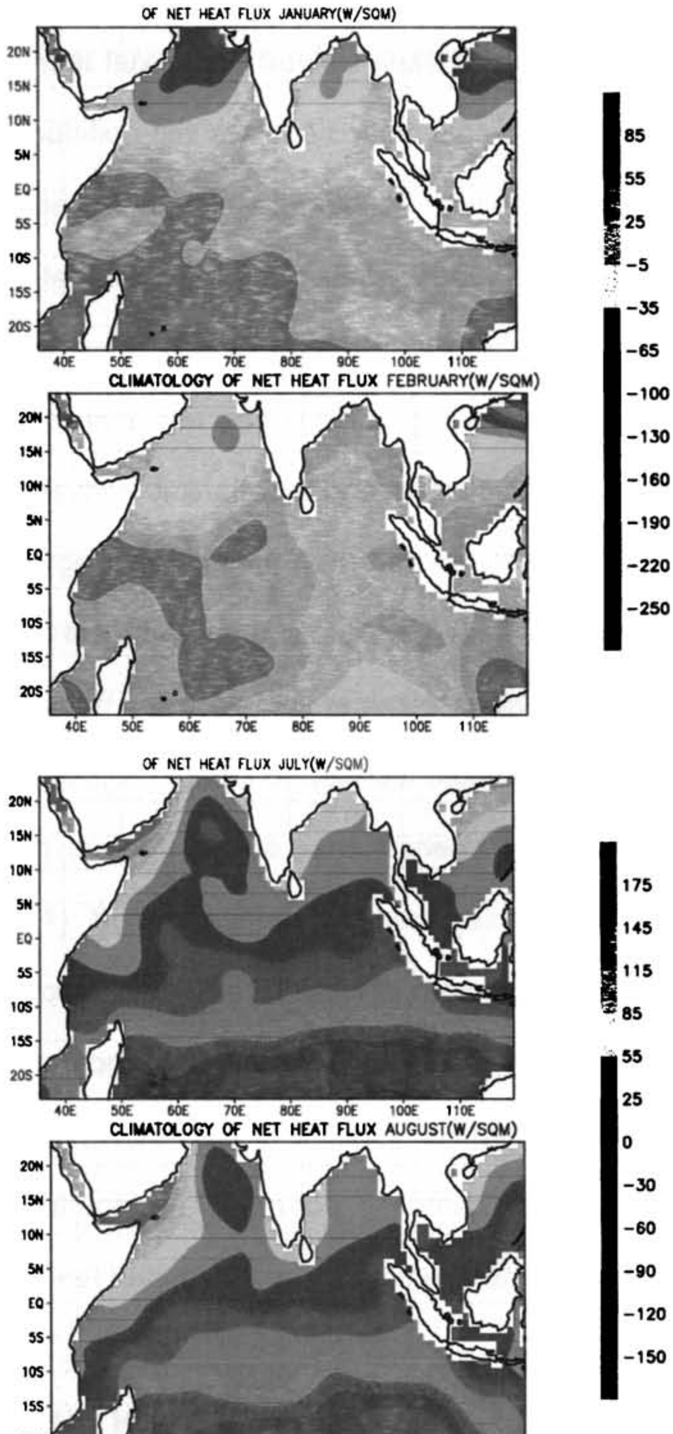


Fig. 4.1 Climatology of Net Heat flux (w/ m^2)

4.3 Climatology of Net Heat Flux

The net heat flux of the tropical seas is mostly controlled by latent heat flux and solar radiation flux and the contribution of long wave flux and sensible heat flux are small. It is mainly a balance between these two fluxes. Over the tropical seas, latent heat flux represents a heat loss from the ocean while the incoming short wave flux represents a heat gain. To a large extent, these two quantities compensate one another during summer monsoon. A negative net heat flux indicates the dominance of latent heat flux over solar radiative flux. This suggests that large evaporation from the oceans takes place due to moisture convergence. On the other hand, a positive heat flux represents less cloud cover with increase in short wave flux and reduction in evaporation from the ocean.

The climatologic Figures of net heat flux show that generally it is positive in Indian Ocean throughout the year, especially in Arabian Sea and Bay of Bengal (Fig. 4.1). So an absolute net heat gain occurs in northern as well as central Indian Ocean. Only the subtropical region of Indian Ocean experiences a heat loss in pre-monsoon season and during the course of monsoon. Post-monsoon season again experiences a net heat gain throughout the Indian Ocean including the subtropics. It can be assumed that the Indian Ocean gains momentum from net heat flux throughout the year.

4.4 Anomaly of Net Heat Flux

The Net heat flux gives an estimation of the amount of heat gained/lost, at the air-sea interface. Similar to other fields, Net heat flux anomaly also indicates

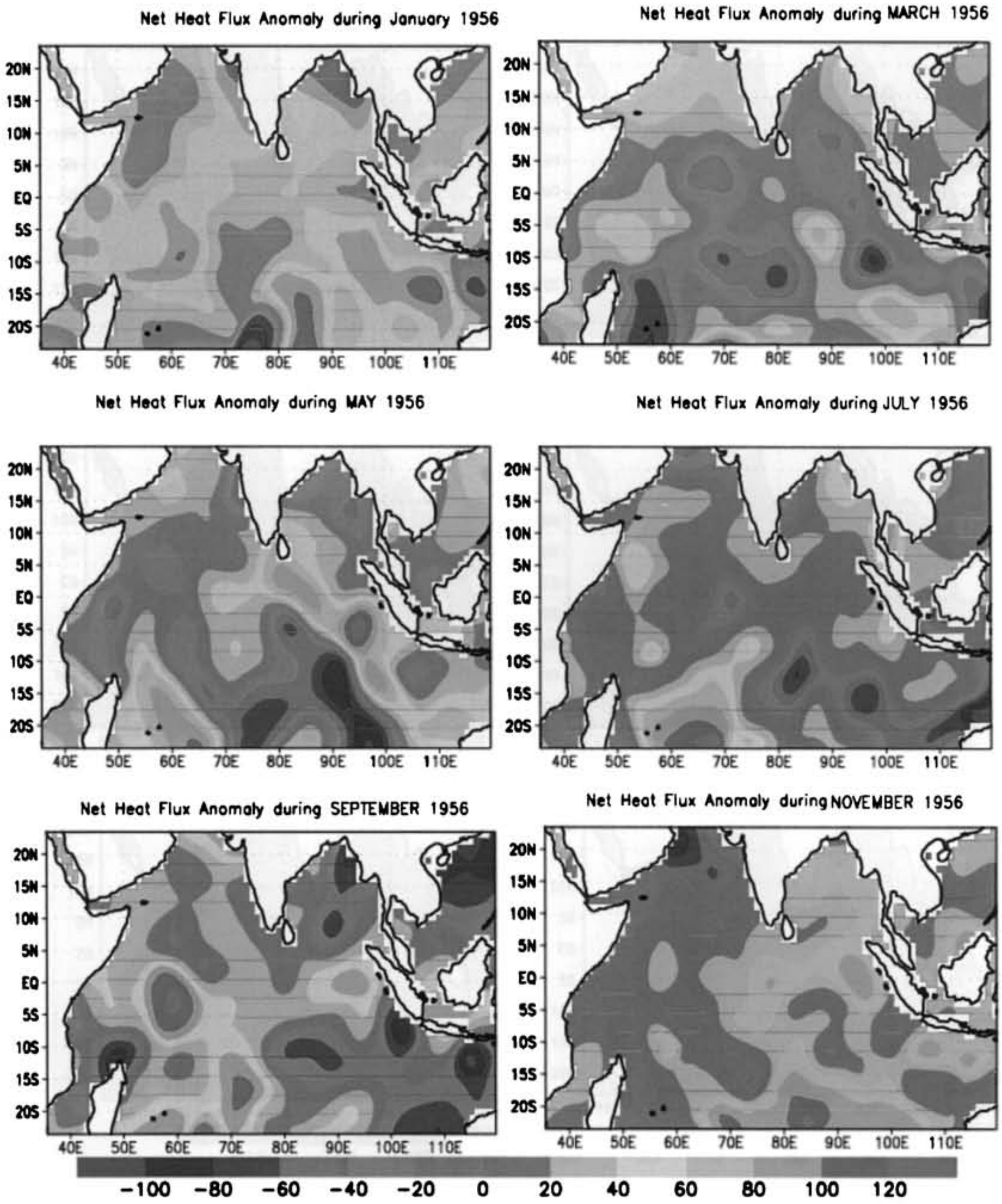


Fig: 4.2 Monthly Mean of Net Heat Flux Anomaly (w/m^2) in the Indian Ocean during 1956

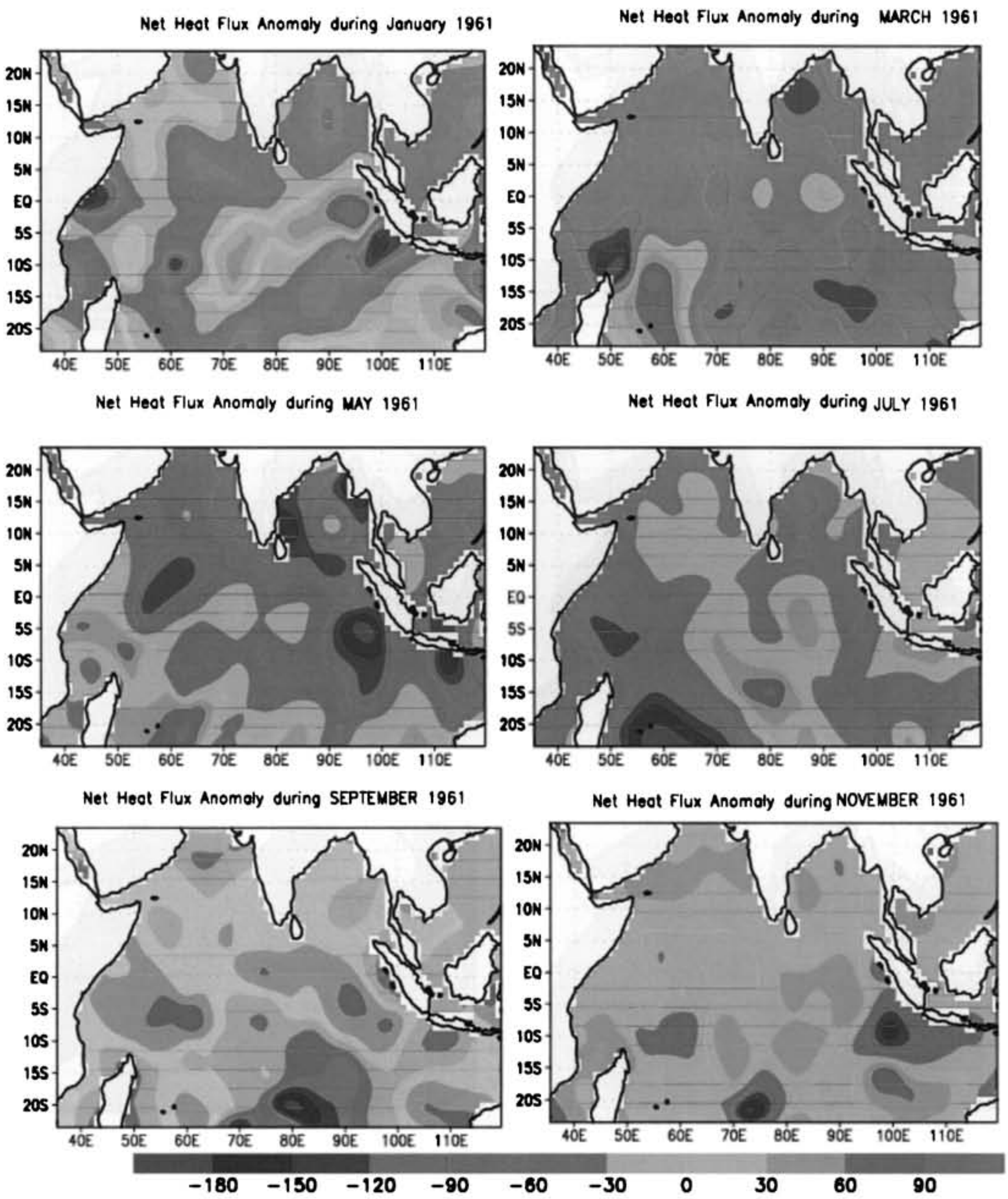


Fig: 4.3 Monthly Mean of Net Heat Flux Anomaly (w/m^2) in the Indian Ocean during 1961

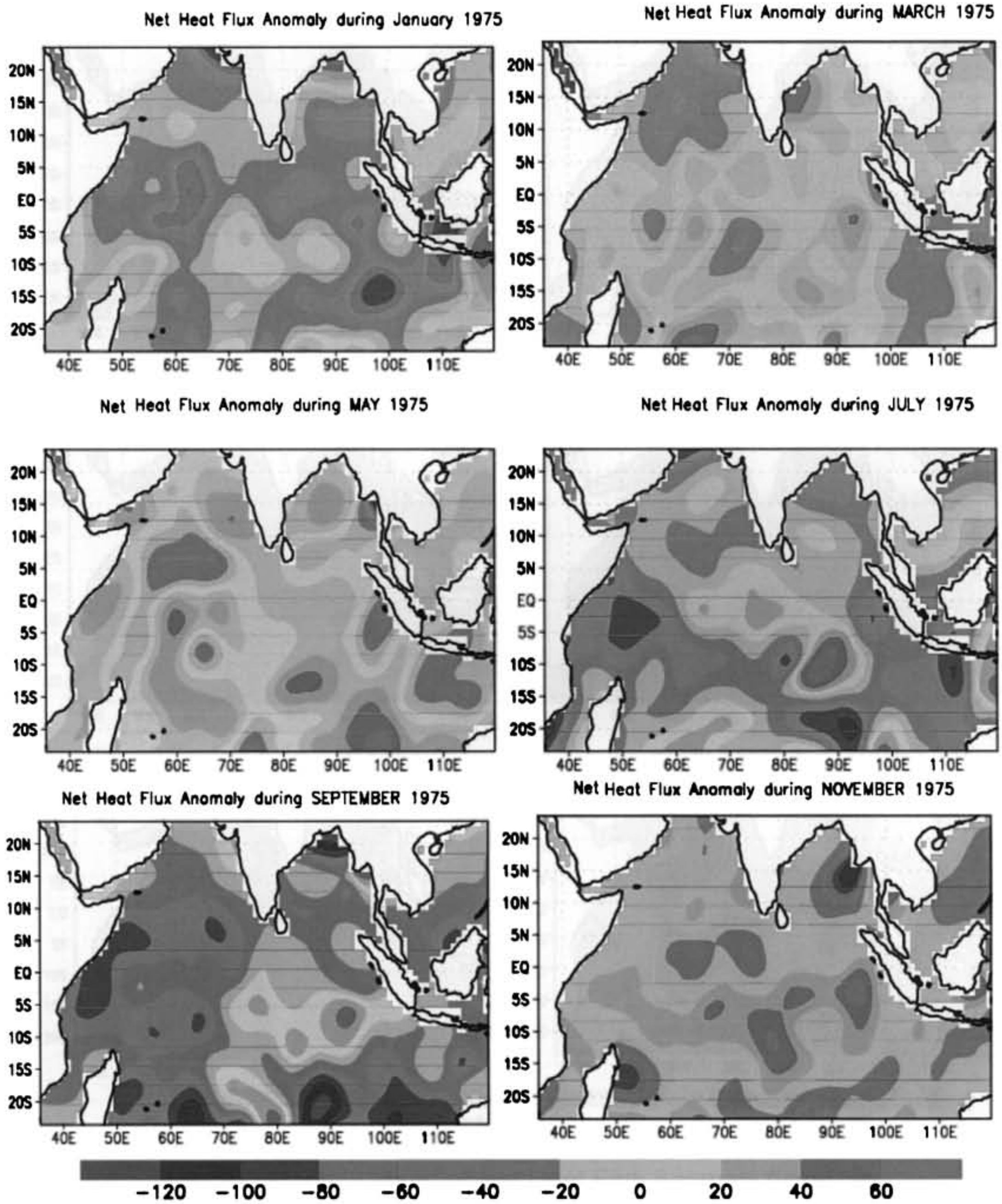


Fig: 4.4 Monthly Mean of Net Heat Flux Anomaly (w/m^2) in the Indian Ocean during 1975

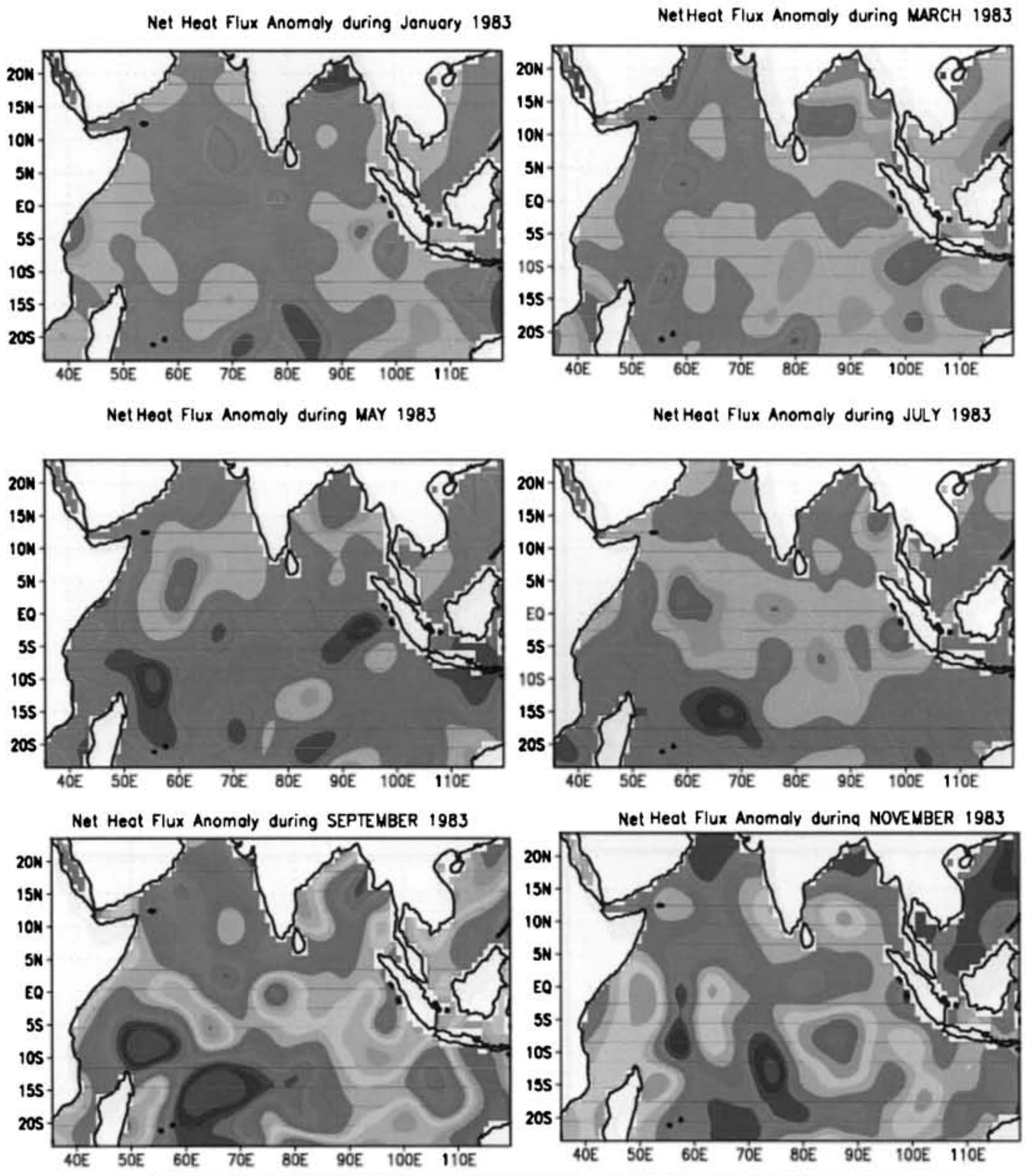


Fig: 4.5 Monthly Mean of Net Heat Flux Anomaly (w/m^2) in the Indian Ocean during 1983

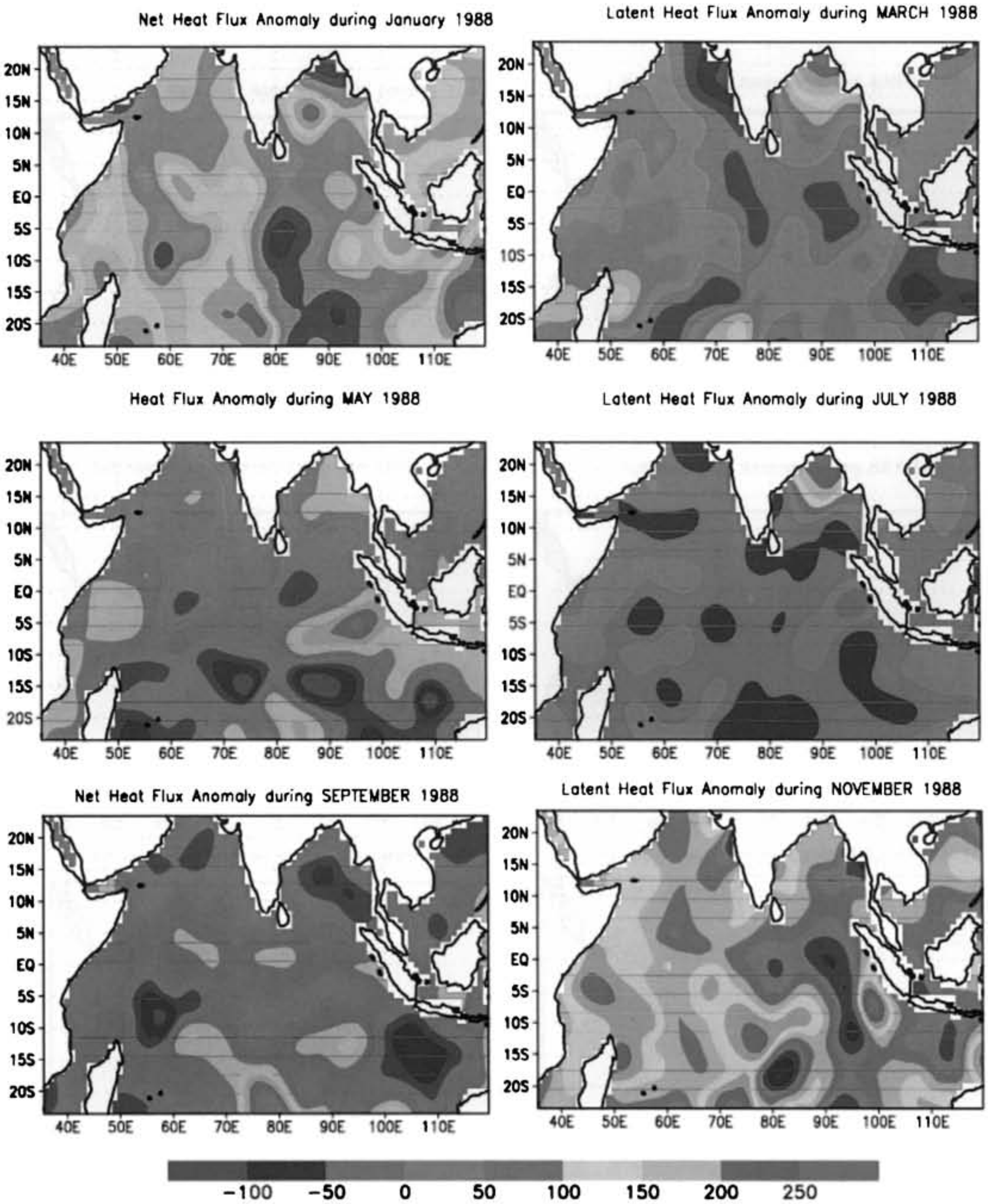


Fig: 4.6 Monthly Mean of Net Heat Flux Anomaly (w/m^2) in the Indian Ocean during 1988

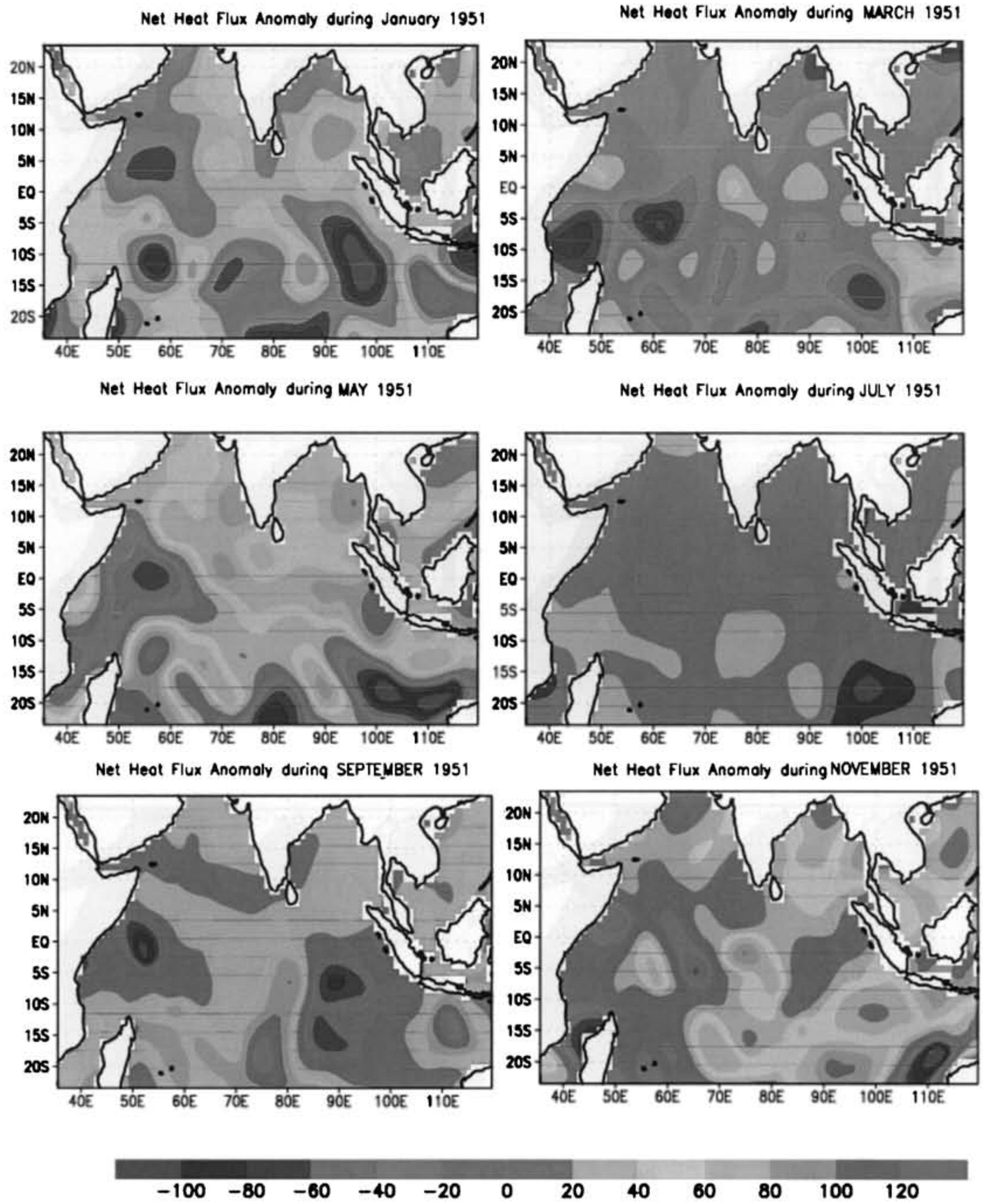


Fig: 4.7 Monthly Mean of Net Heat Flux Anomaly (w/m^2) in the Indian Ocean during 1951

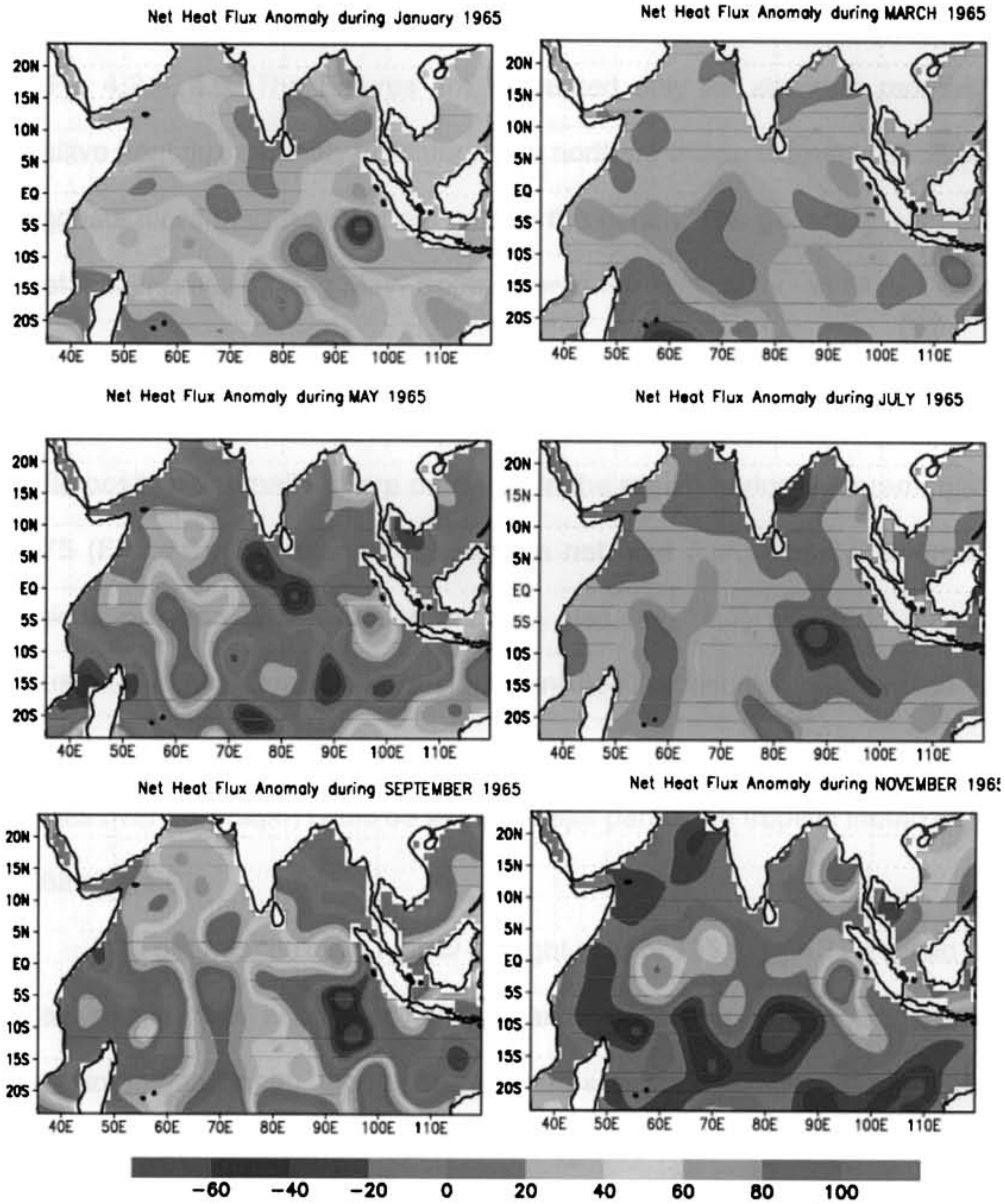


Fig: 4.8 Monthly Mean of Net Heat Flux Anomaly (w/m^2) in the Indian Ocean during 1965

that negative flux anomaly is associated with increased air-sea interaction process.

The net heat flux anomaly for typical good monsoon years are presented in Fig 4.2 – 4.6. The Figures are presented only for alternate months. The positive heat flux anomaly in winter in the northern Indian Ocean in 1956 (Fig. 4.2) suggests more than normal heat gain to the ocean. The generally negative net heat flux anomaly during May- September in the tropical Indian Ocean could be due to decrease in solar radiation as well as increased evaporation. The zone of negative heat flux anomaly increased in 1961 except in September (Fig. 4.3) while positive anomalies were dominant in the region during the premonsoon of 1975 (Fig. 4.4). The domain of negative net heat flux anomaly covered most regions of north Indian Ocean in 1983 also (Fig. 4.5) except in the central equatorial Indian Ocean where significant positive anomalies prevailed during monsoon. Exceptionally large negative anomalies of net heat flux (sometimes values over 50 w/sqm) could be seen in major part of the tropical Indian Ocean in 1988 (Fig.4.6).

The net heat flux in a typical drought year of 1951(Fig.4.7) showed mostly positive heat flux anomalies in May and September over major portion of northern Indian Ocean, though in July there was considerable negative heat flux anomalies over the same region. The negative heat flux anomalies over the same region were noticed only towards the end of monsoon of 1965 (Fig 4.8), though the geographic extent of the zone of negative heat flux anomalies was limited to the seas around India.

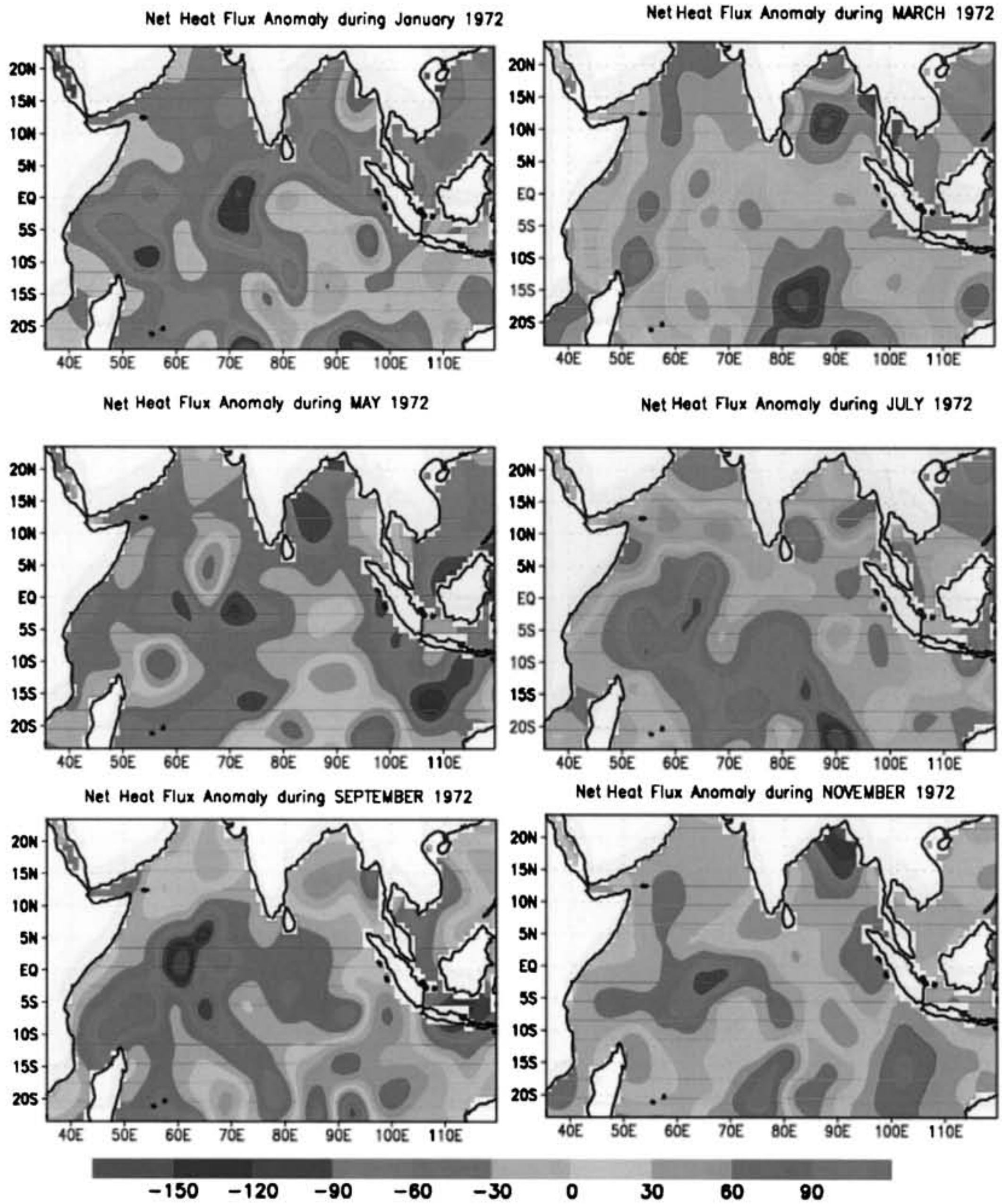


Fig: 4.9 Monthly Mean of Net Heat Flux Anomaly (w/m^2) in the Indian Ocean during 1972

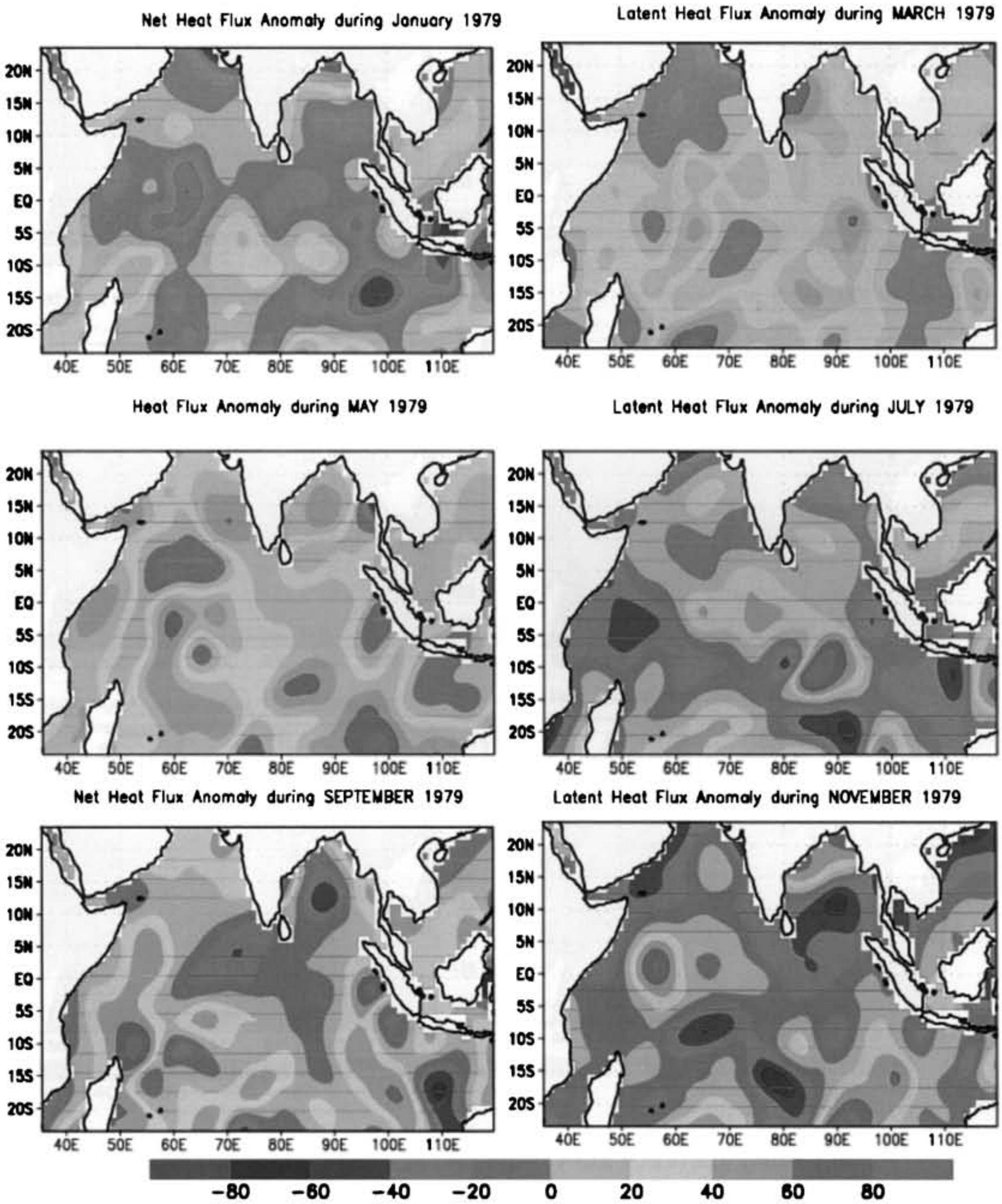


Fig: 4.10 Monthly Mean of Net Heat Flux Anomaly (w/m^2) in the Indian Ocean during 1979

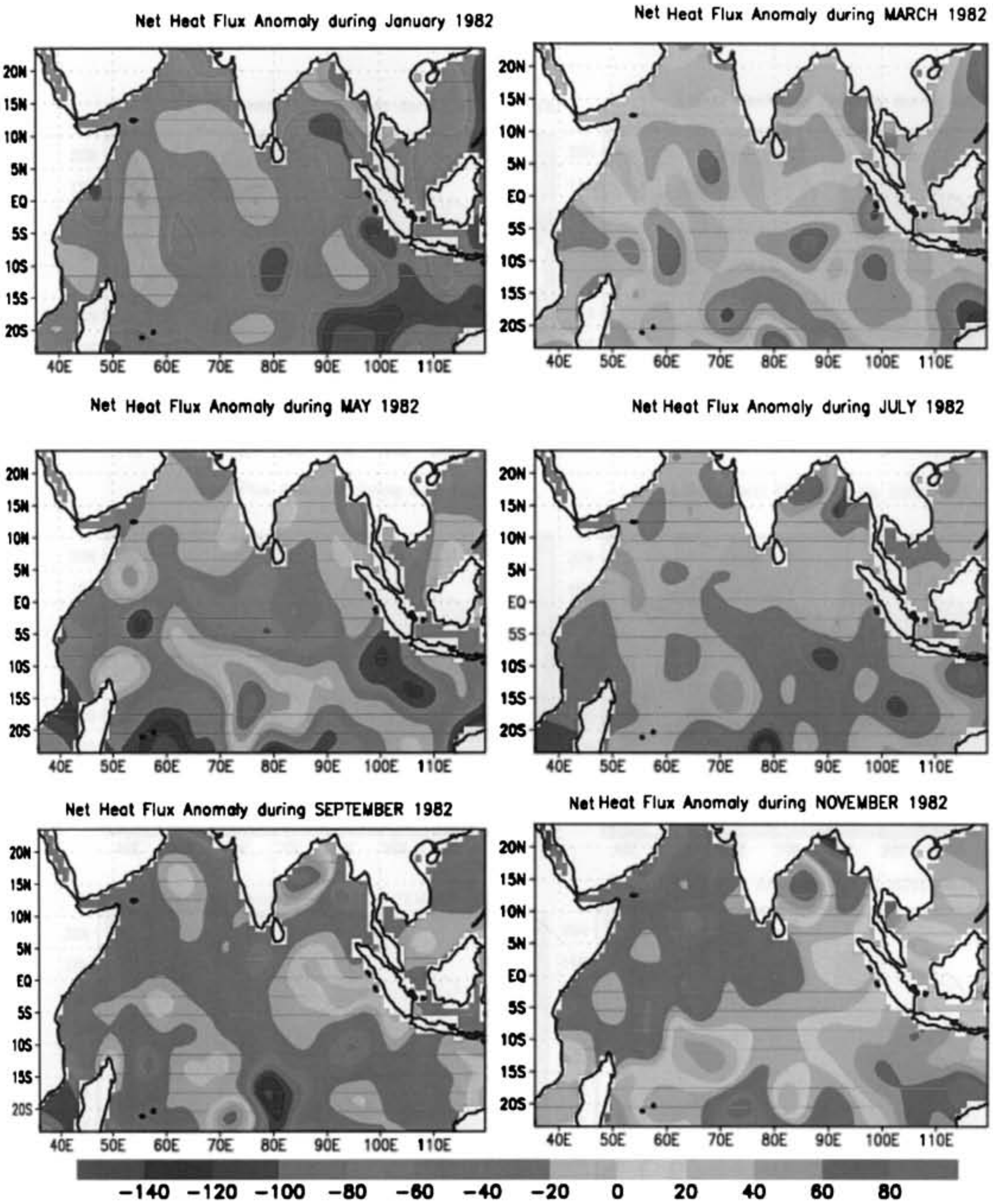


Fig: 4.11 Monthly Mean of Net Heat Flux Anomaly (w/m^2) in the Indian Ocean during 1982

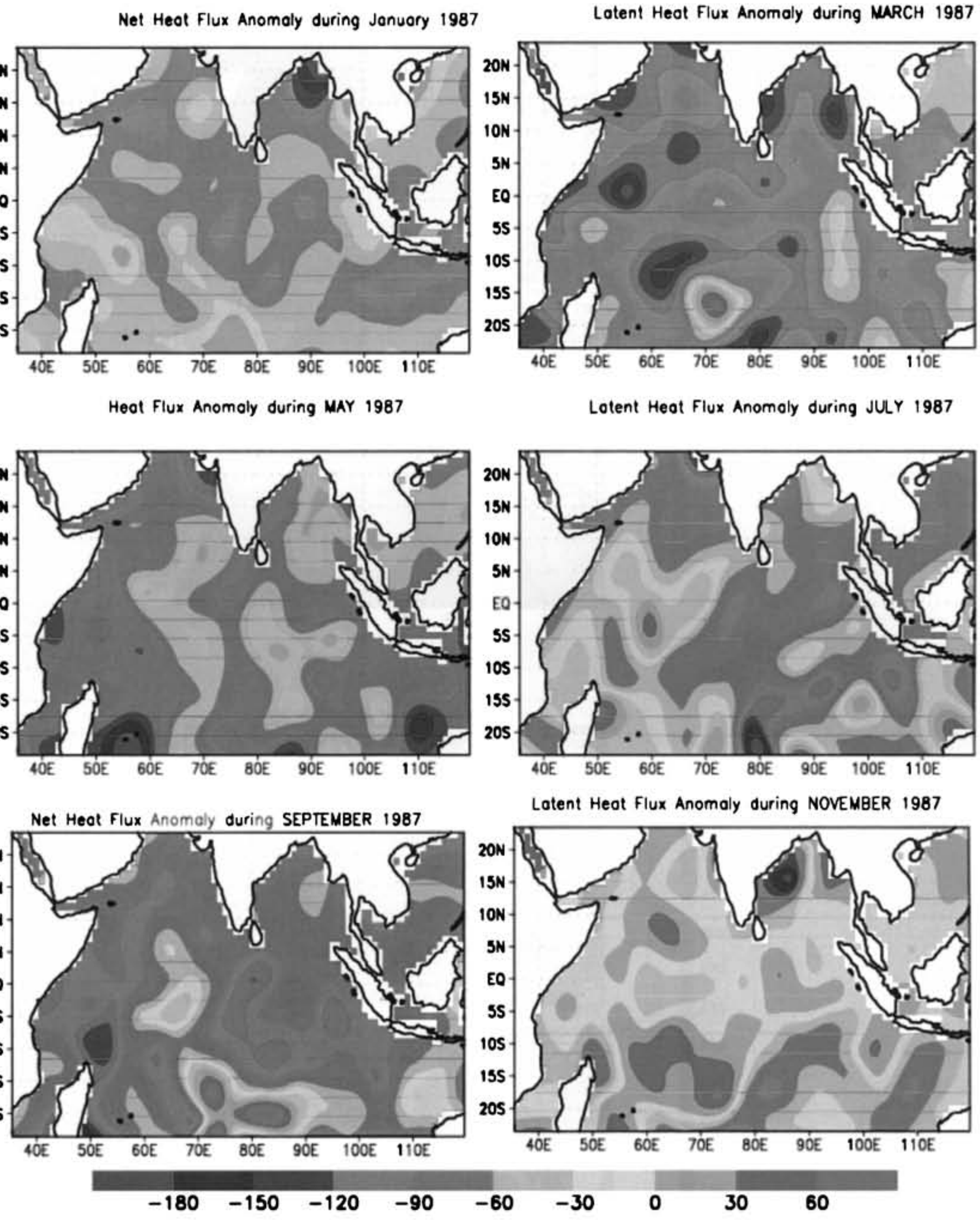


Fig: 4.12 Monthly Mean of Net Heat Flux Anomaly (w/m^2) in the Indian Ocean during 1987

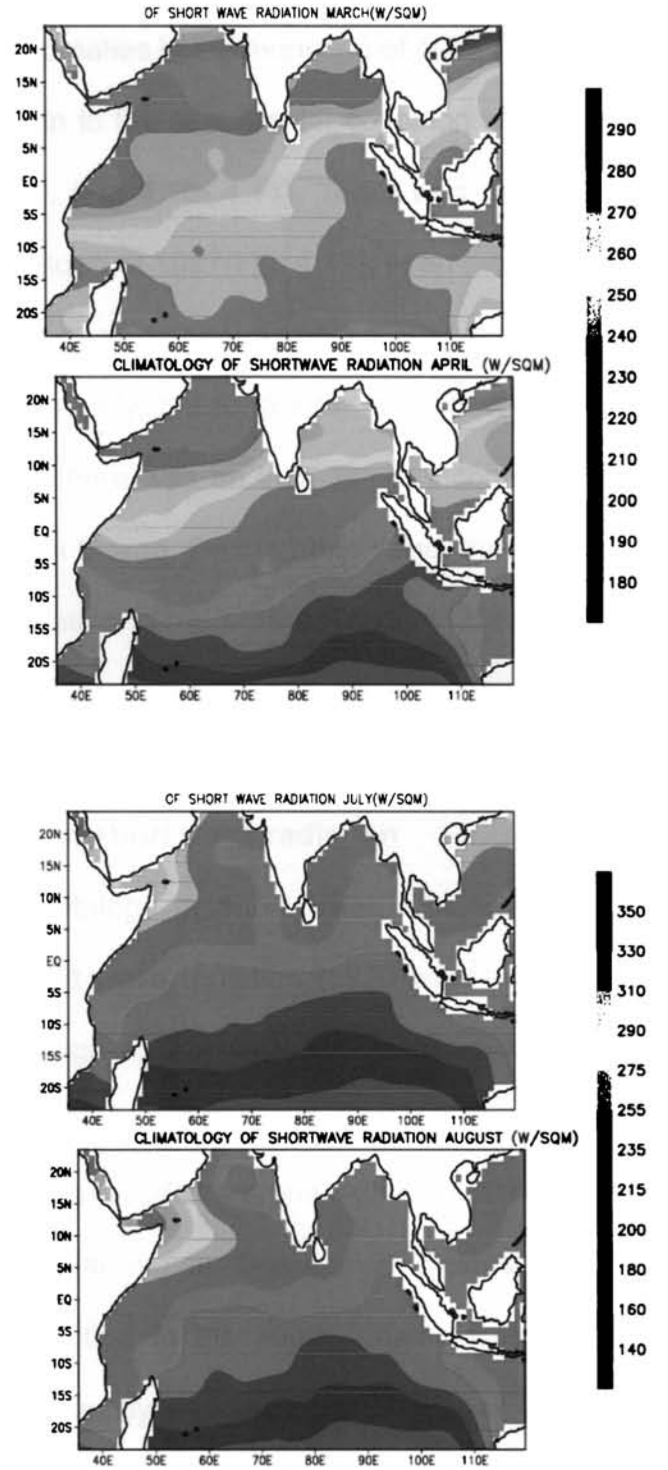


Fig. 4.13 Climatology of Short Wave Radiation (w/m^2)

Substantially large area of the northern Indian Ocean experienced positive net heat flux anomalies in the monsoon of 1972 (Fig. 4.9) resulting in higher than normal heat gain to the sea. Similar condition with minor difference occurred in 1979 and 1982. (Fig 4.10- 4.11). There was marginal heat loss by the ocean during the monsoon of 1987(Fig. 4.12), even though it was a drought year.

Hence in good monsoon years there was considerable decrease in short wave radiation mainly due to greater cloud cover in the north Indian Ocean. At the same time, there was an increase in incoming short wave radiation in the equatorial Indian Ocean. On the other hand, there was considerable increase in short wave radiation in the northern Indian Ocean associated with reduction in the equatorial Indian Ocean.

4.5 Anomalies of short wave radiation

The climatology of short wave radiation (Fig. 4.13) shows maximum incidents of short wave radiation (>250 w/m^2) in the sub-tropical south Indian Ocean which progressively decreases northwards with less than 200 w/m^2 in the the northern Arabian sea and Bay of Bengal. By March – April the zone of maximum incoming short wave radiations shifts to northern Arabian Sea, Bay of Bengal and western equatorial Indian Ocean with values of 270 w/m^2 while there is substantial reduction in the south Indian Ocean. The values of short wave radiation decreases by May and during monsoon in the entire north Indian Ocean the incoming short wave radiation decreases to value below 240 w/m^2 while north with the advent of austral winter the short wave flux remains very low. During post

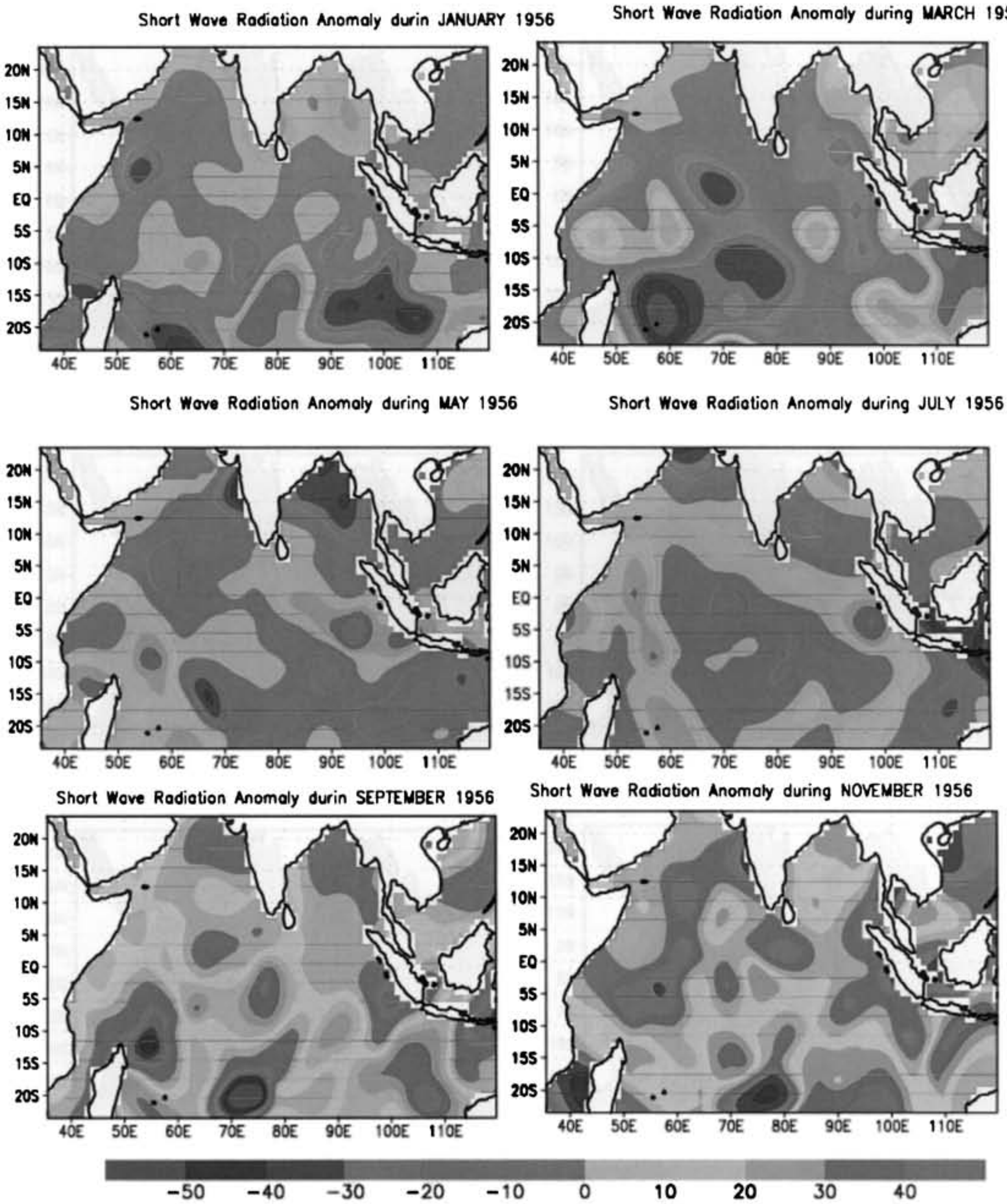


Fig: 4.14 Monthly Mean of Short Wave Radiation Anomaly (w/m^2) in the Indian Ocean during 1956

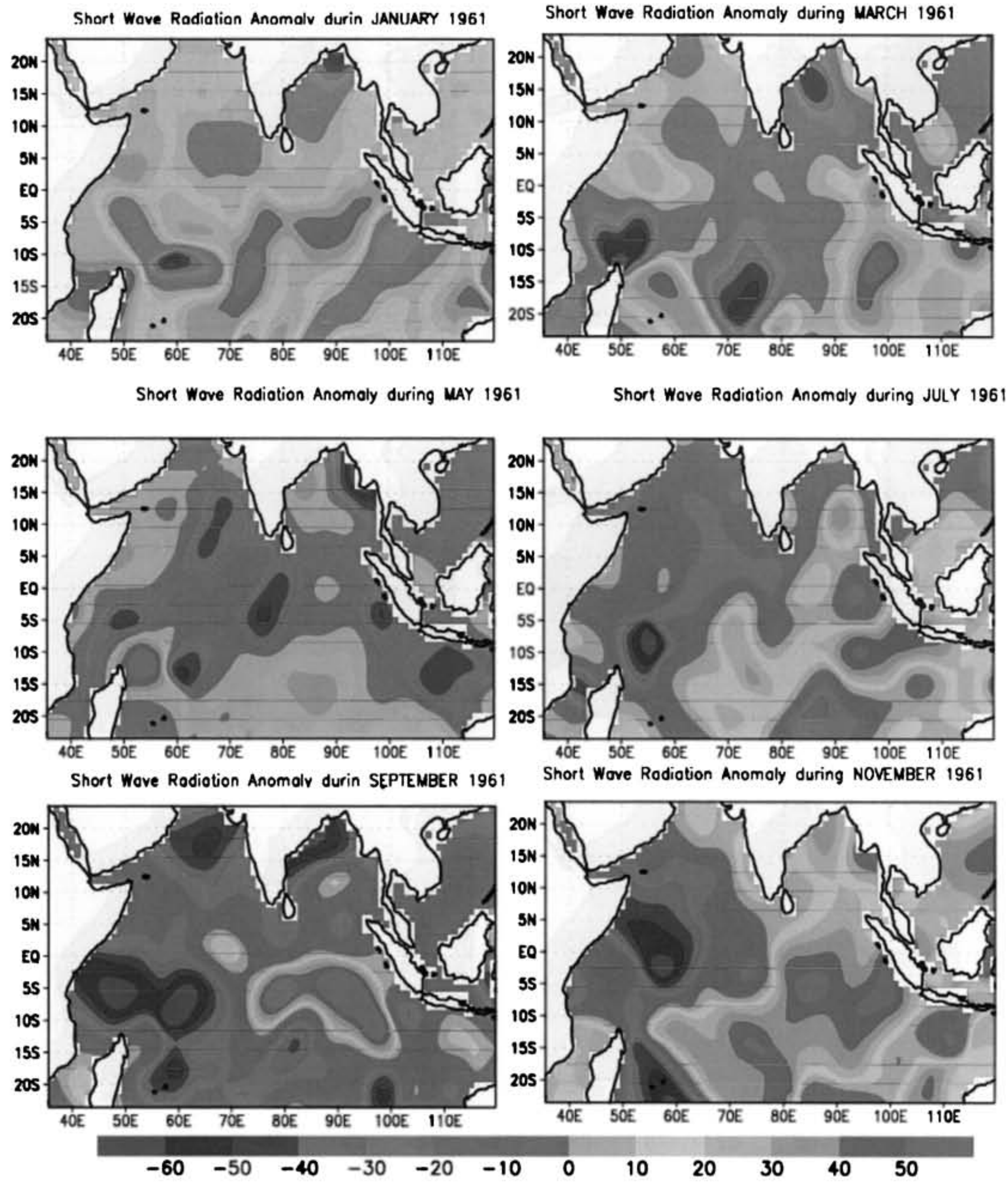


Fig: 4.15 Monthly Mean of Short Wave Radiation Anomaly (w/m^2) in the Indian Ocean during 1961

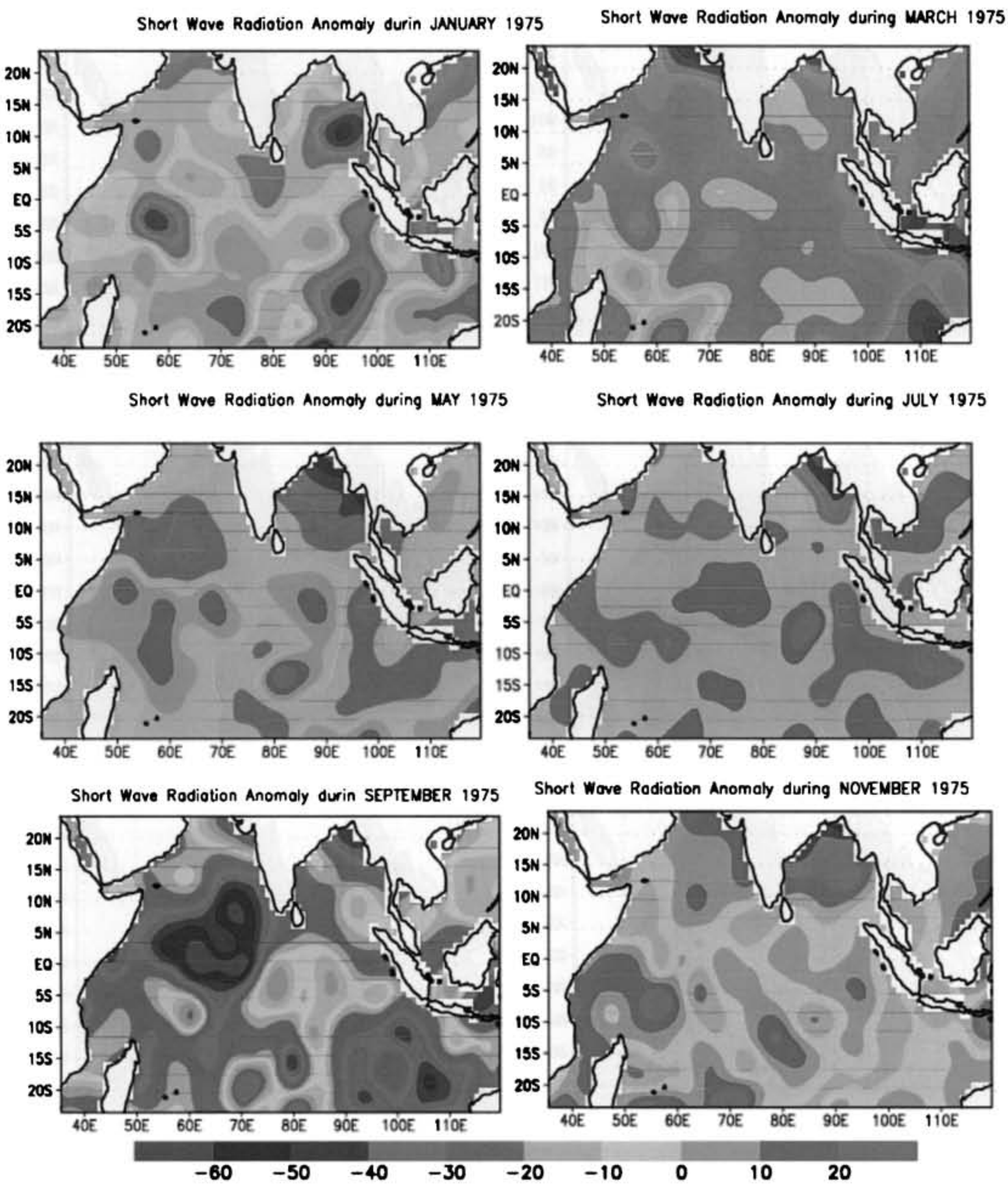


Fig: 4.16 Monthly Mean of Short Wave Radiation Anomaly (w/m^2) in the Indian Ocean during 1975

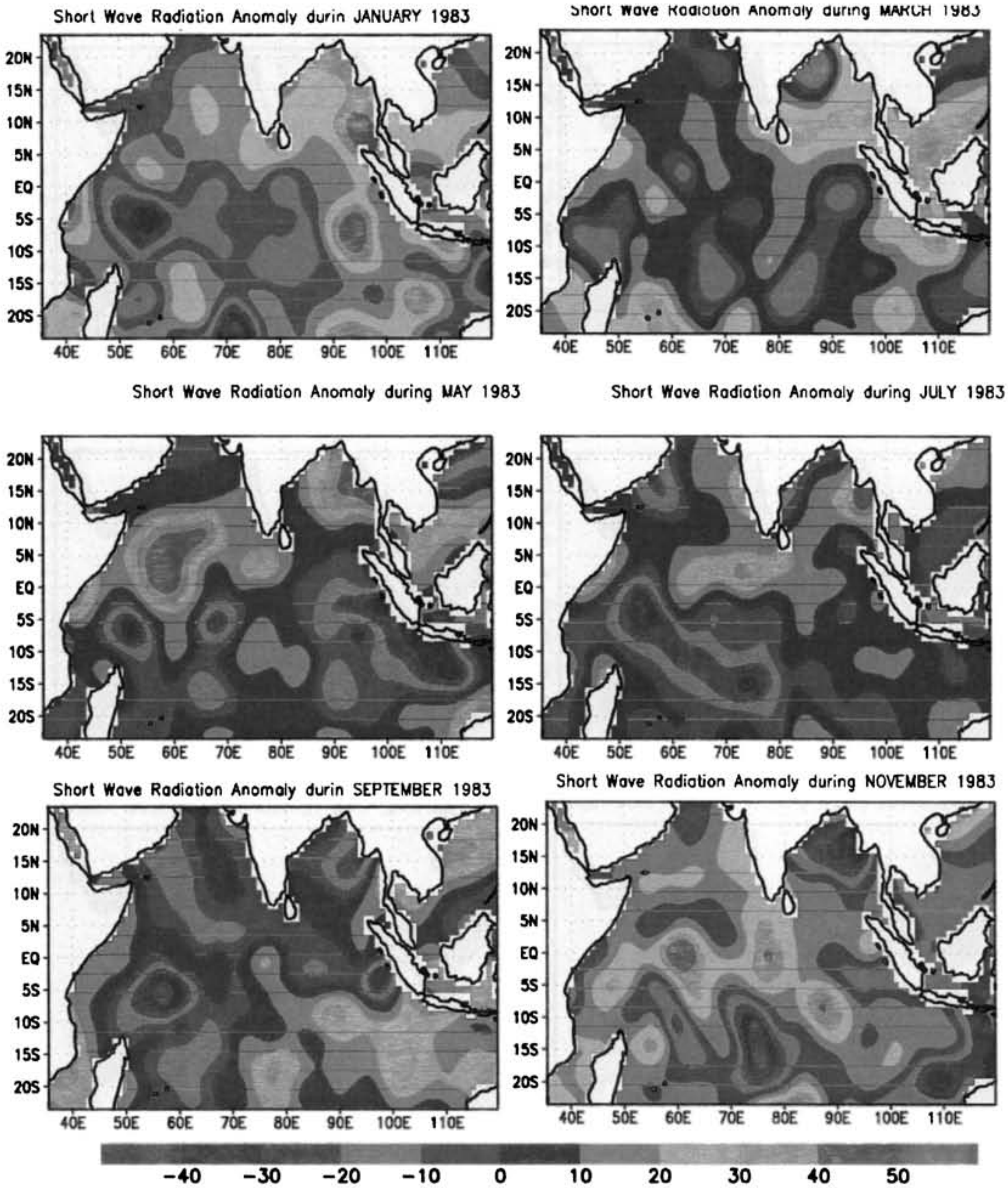


Fig: 4.17 Monthly Mean of Short Wave Radiation Anomaly (w/m^2) in the Indian Ocean during 1983

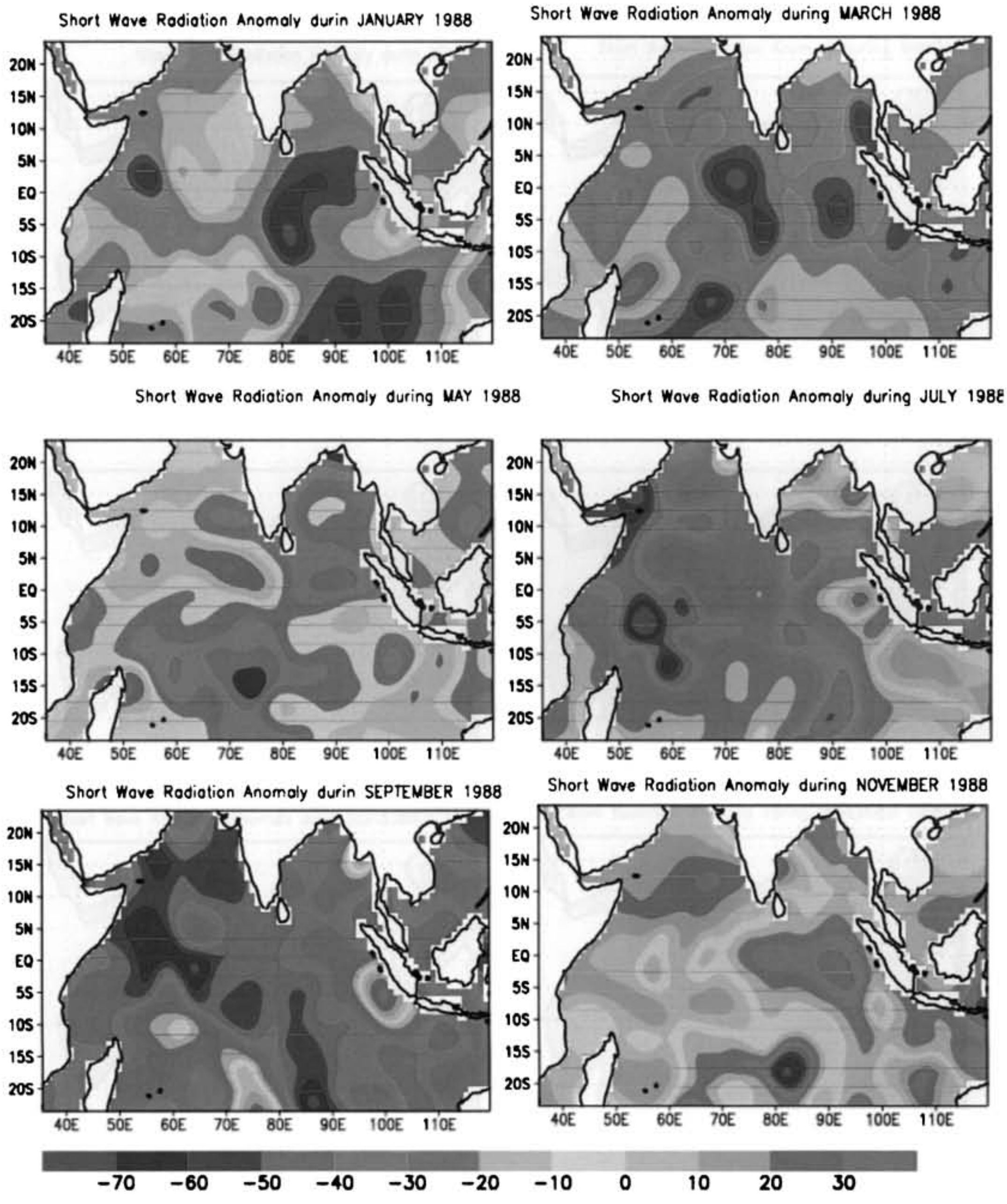


Fig: 4.18 Monthly Mean of Short Wave Radiation Anomaly (w/m^2) in the Indian Ocean during 1988

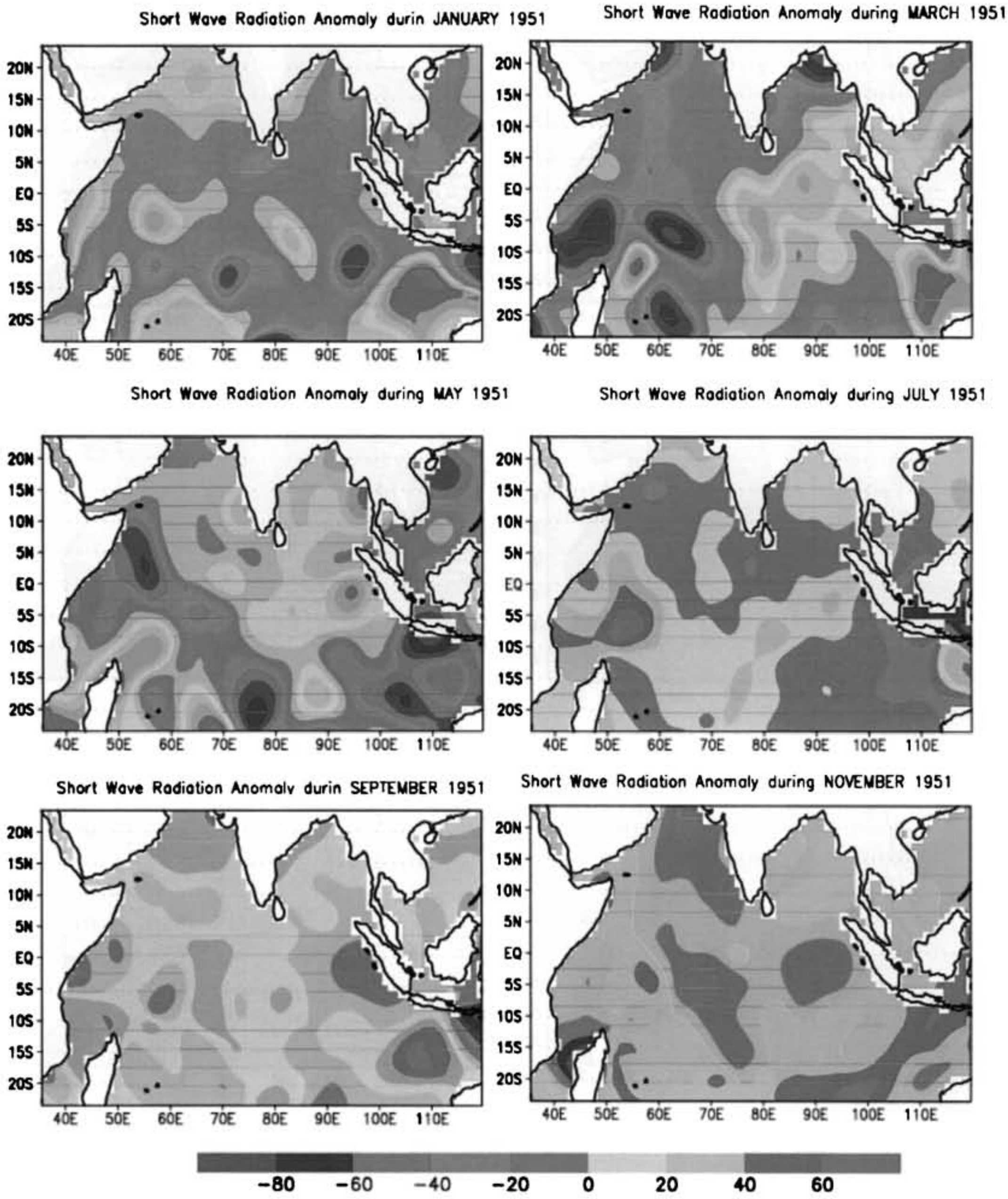


Fig: 4.19 Monthly Mean of Short Wave Radiation Anomaly (w/m^2) in the Indian Ocean during 1951

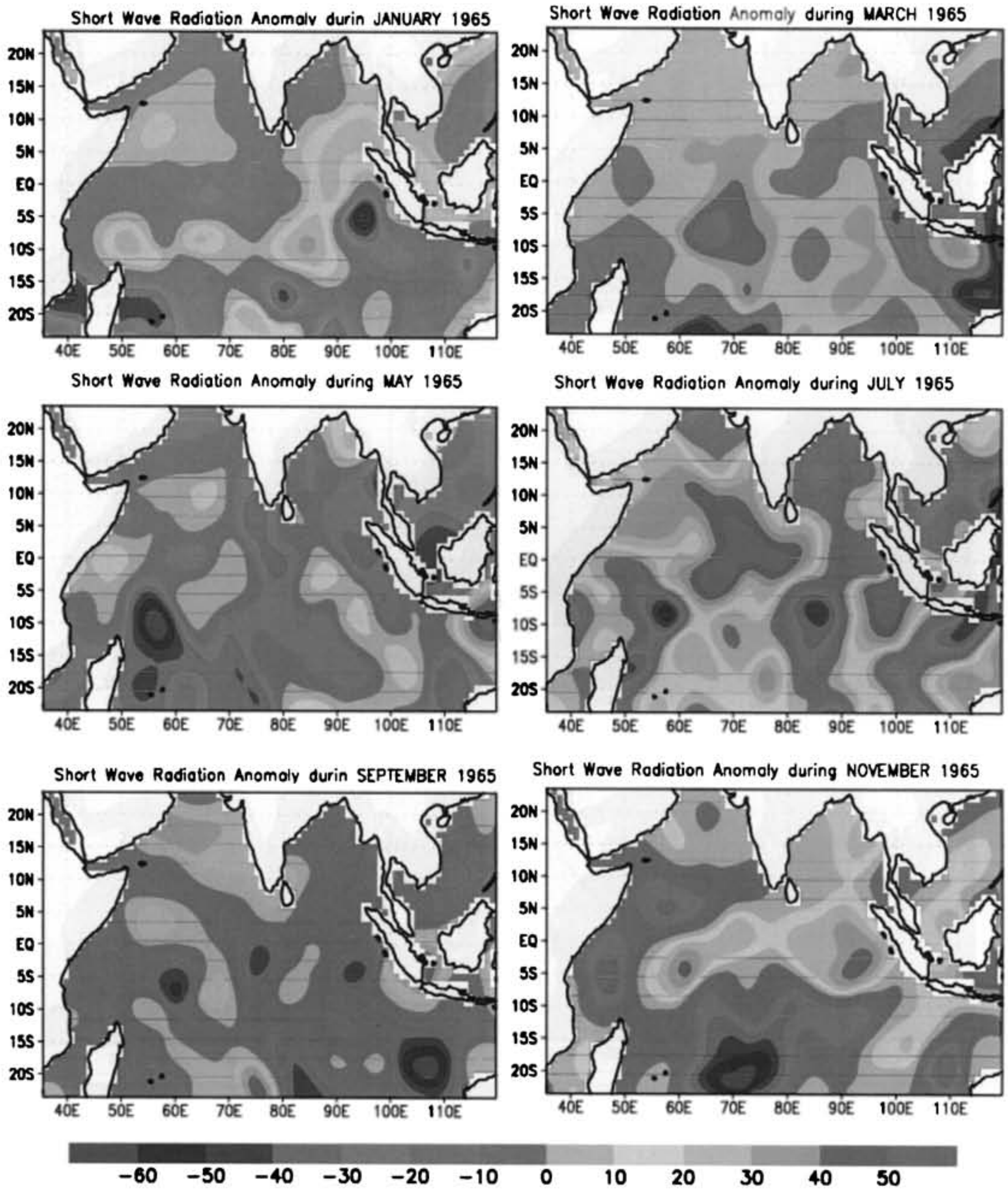


Fig: 4.20 Monthly Mean of Short Wave Radiation Anomaly (w/m^2) in the Indian Ocean during 1965

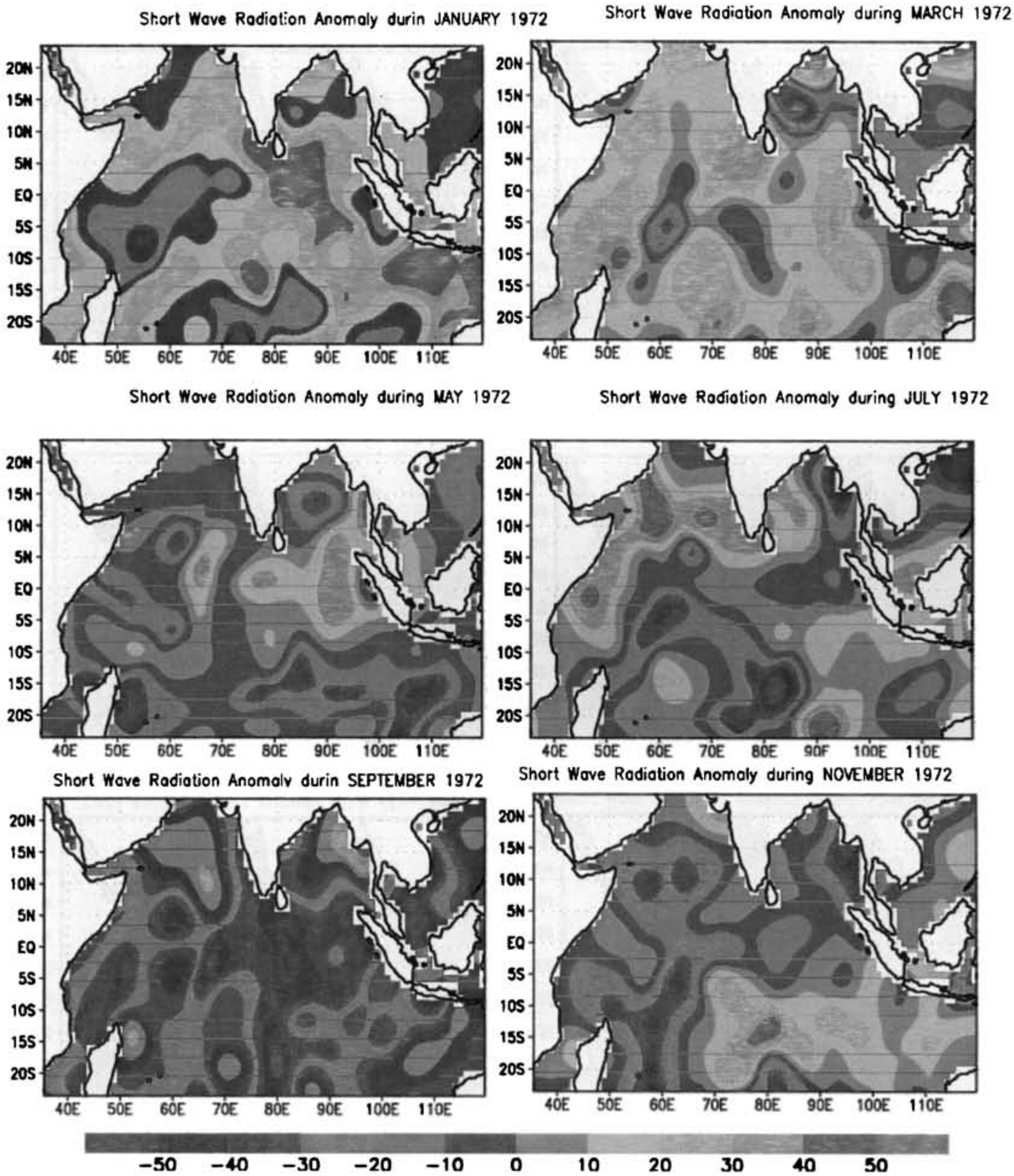


Fig: 4.21 Monthly Mean of Short Wave Radiation Anomaly (w/m^2) in the Indian Ocean during 1972

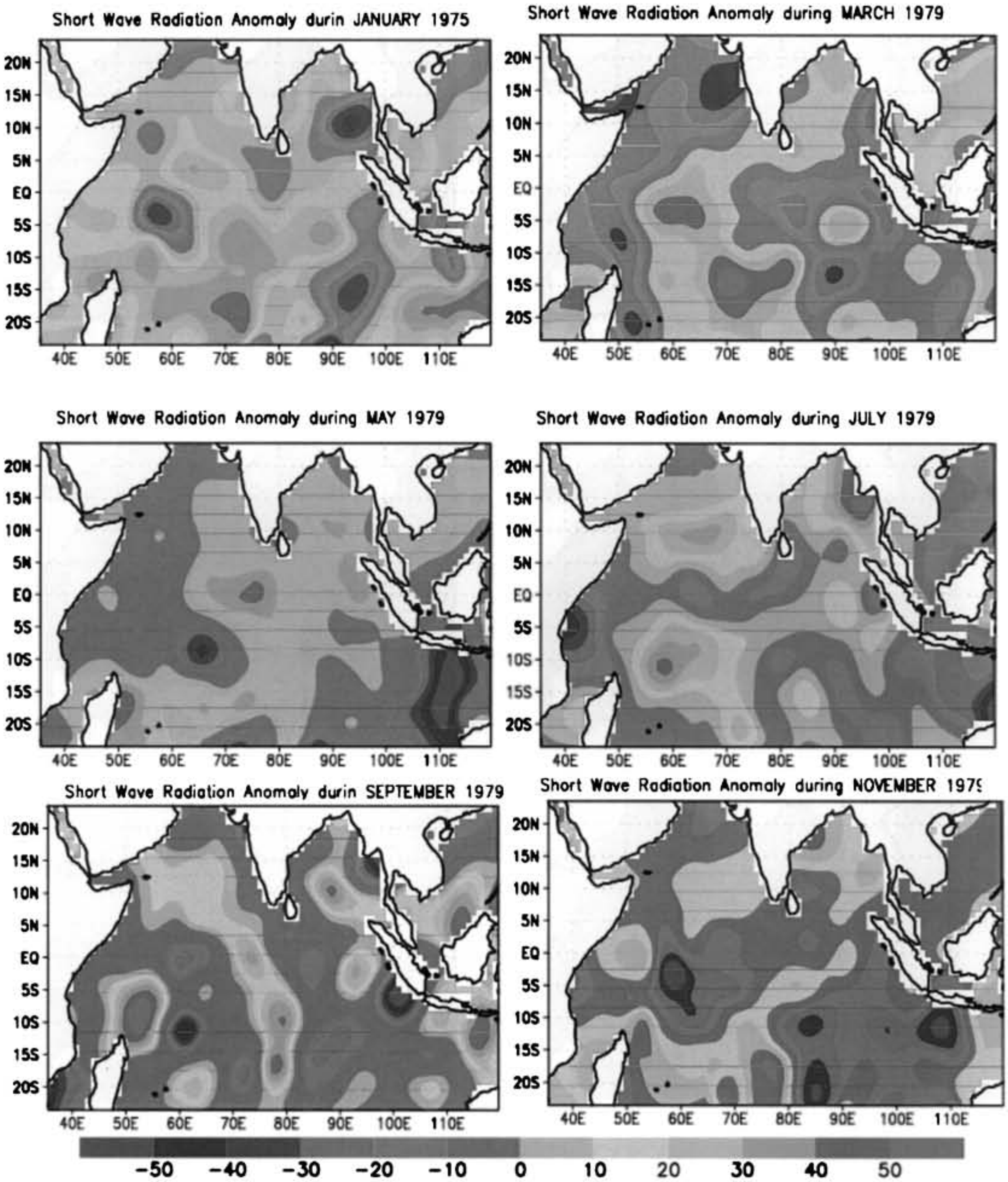


Fig: 4.22 Monthly Mean of Short Wave Radiation Anomaly (w/m^2) in the Indian Ocean during 1979

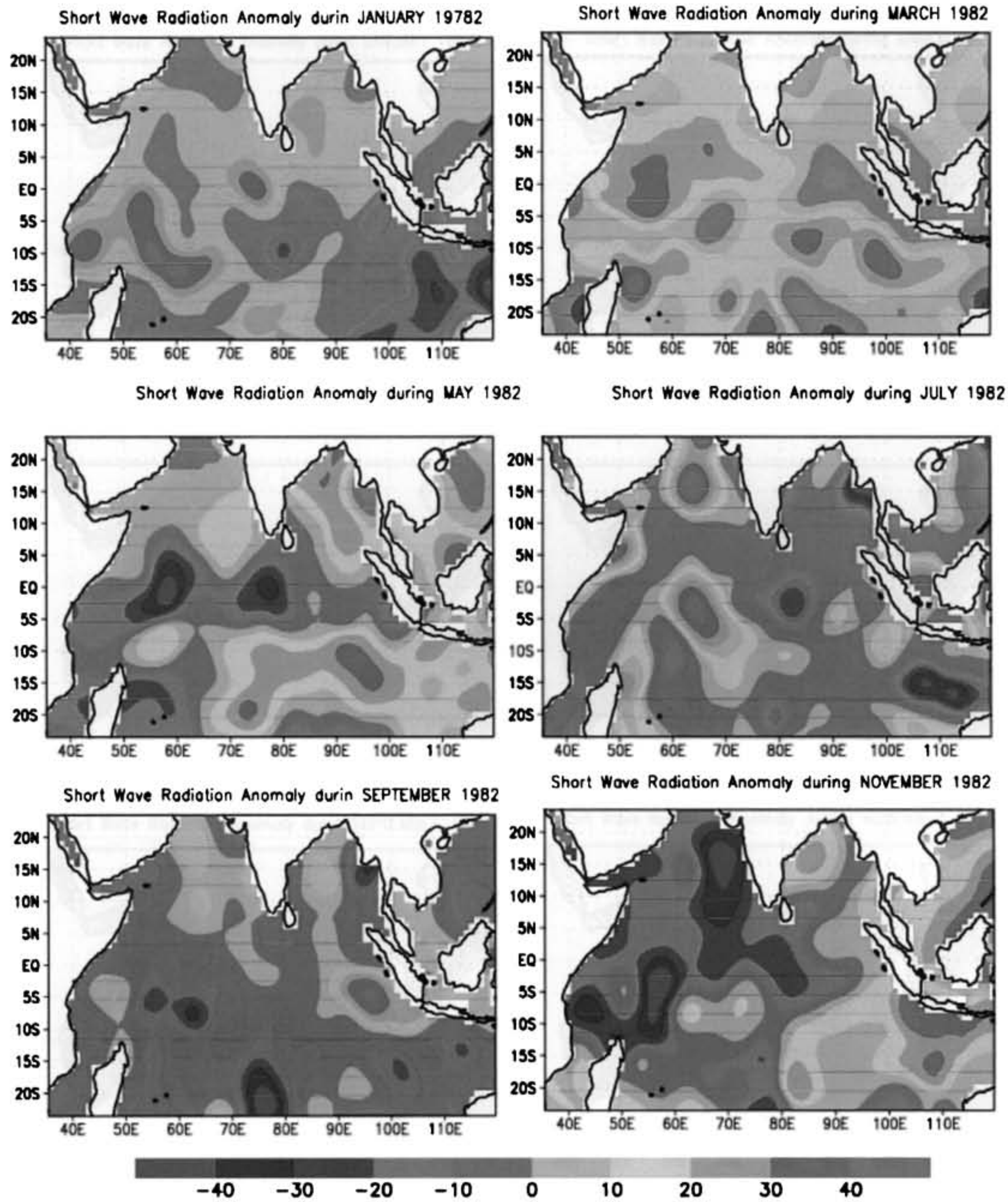


Fig: 4.23 Monthly Mean of Short Wave Radiation Anomaly (w/m^2) in the Indian Ocean during 1982

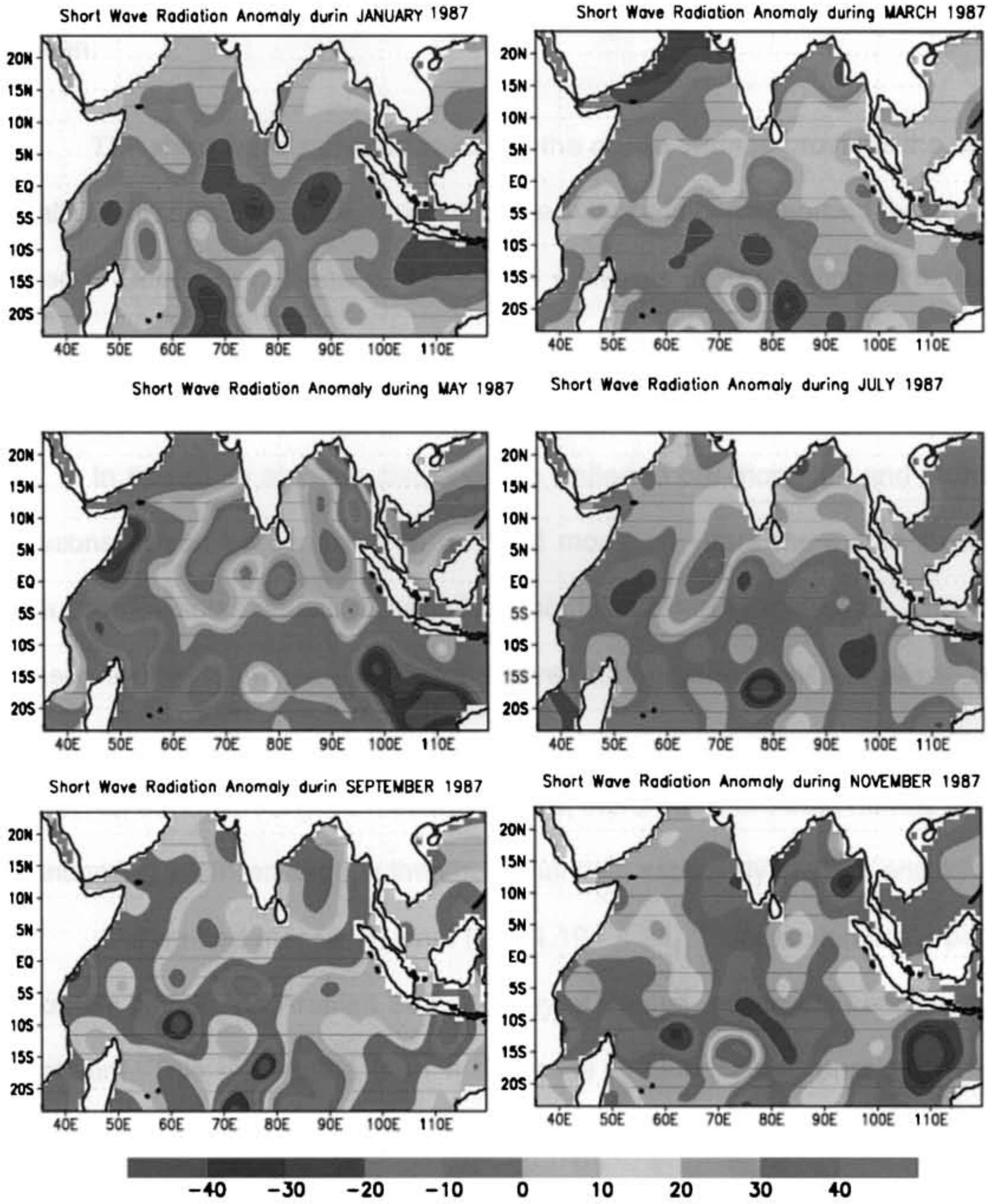


Fig: 4.24 Monthly Mean of Short Wave Radiation Anomaly (w/m^2) in the Indian Ocean during 1987

monsoon, there is large east west asymmetry with values over 260 w/m^2 in the western Indian Ocean and with values less than 220 w/m^2 in the eastern Indian Ocean.

The short wave radiation reaching the ocean surface provides the source of all weather patterns. The variation in the solar radiation absorbed by the upper layer of Ocean is not one of the crucial elements in forecasting climate. The anomaly of incoming solar radiation during severe drought and flood year of radial Indian monsoon are presented in figures 4.14 - 4.24.

In this study also the discussion is limited to pre-monsoon and monsoon seasons. It can be seen that in all good monsoon years there was sustained as reduction in solar radiation (negative anomalies) during monsoon. The major reason for this is the increased cloud cover during good monsoon years, which inhibit the incoming solar radiation especially in the northern Indian Ocean. However, during most good monsoon years, there was more than normal amount of incoming solar radiation in the equatorial belt, especially in 1973 and 1983.

During bad monsoon years (Fig. 4.19 - 4.24), there is significant positive anomaly both in the Arabian Sea and Bay of Bengal, especially near the Indian continent. This is resulted from a decrease in cloud cover during the deficit monsoon years. However, there were more negative anomalies of solar radiation in the equatorial and south Indian Ocean during these years with some year-to-year variations.

The monthly evolution of the spatial variability of evaporation indicated

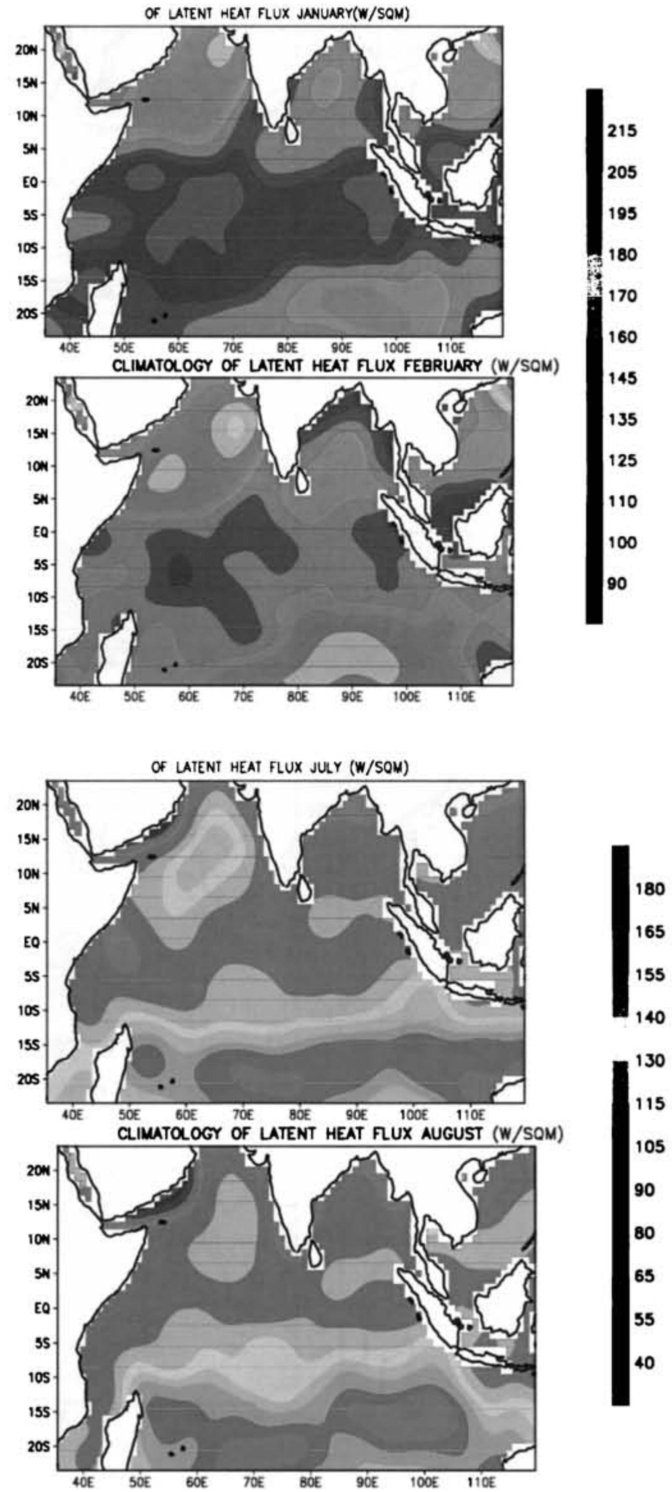


Fig. 4.25 Climatology of Latent Heat Flux (w/m^2)

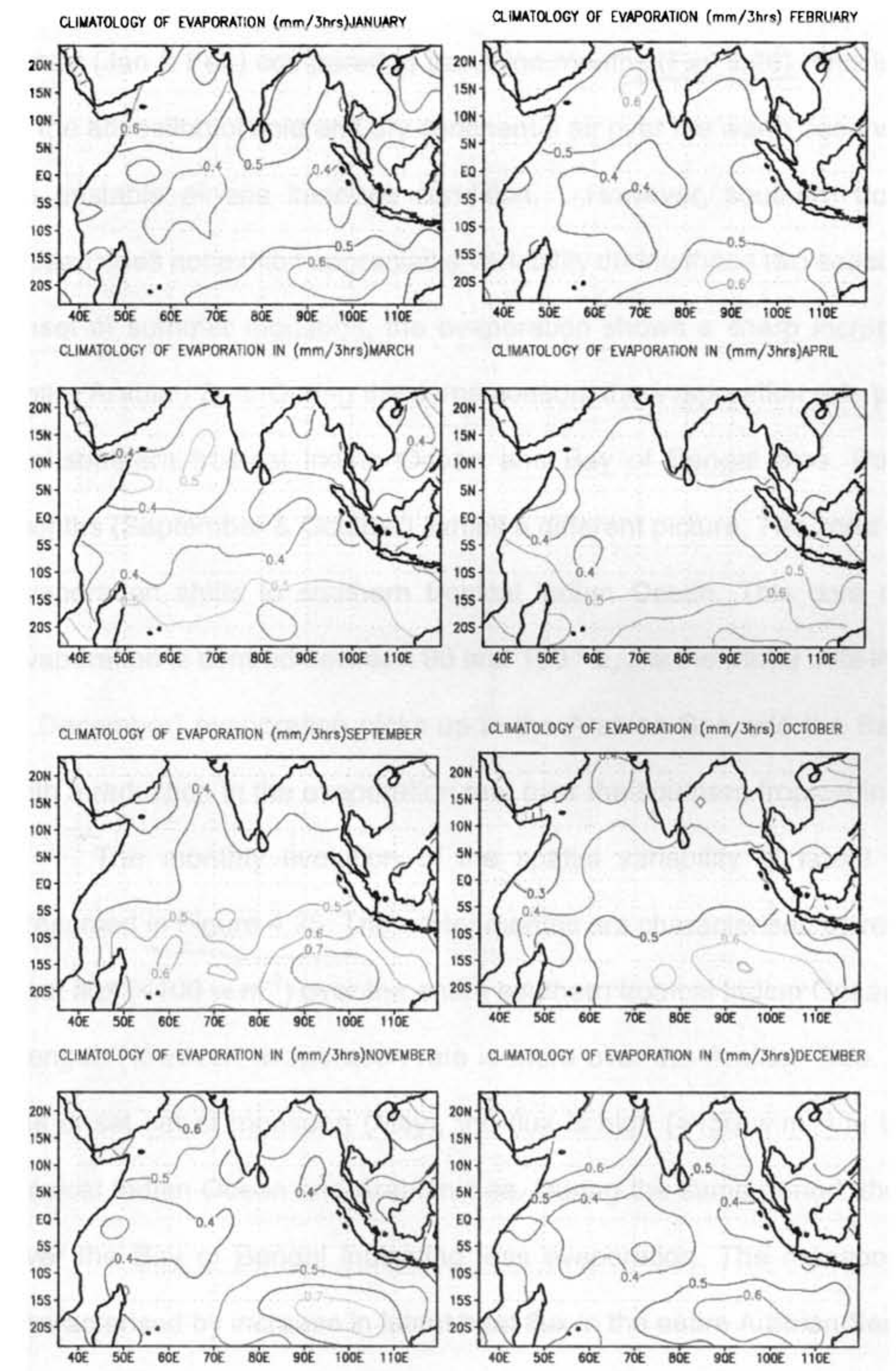


Fig. 4.26 Climatology of Evaporation (mm/3hrs.)

higher evaporation rates over the Arabian Sea and the Bay of Bengal during winter (Jan & Feb) compared to transition months (Fig. 4.26). This is mainly due to the advection of cold and dry continental air over the warm oceans resulting in an unstable air-sea interface condition. However, southern tropical Indian Ocean does not exhibit appreciable variability during these two seasons. With the onset of summer monsoon, the evaporation shows a sharp increase over the entire Arabian Sea. During the same season, the evaporation rate picks up over the southern tropical Indian Ocean and Bay of Bengal also. Post monsoon months (September & October) exhibit a different picture. The zone of maximum evaporation shifts to southern tropical Indian Ocean. This core of maximum evaporation is centred between 90 and 100 $^{\circ}$ E. As the winter sets in (November & December) evaporation picks up in the Arabian Sea and the Bay of Bengal with a reduction in the evaporation rate over the southern tropical Indian Ocean.

The monthly evolution of the spatial variability of latent heat flux is presented in Figure 4.25. The winter months are characterised by reduced latent heat flux (<100 w.m^{-2}) over the entire southern tropical Indian Ocean and after Bay of Bengal. However, evaporation rate is more over the Arabian Sea. Just prior to the onset set of monsoon (May), the flux is high (>150 w.m^{-2}) in the southern tropical Indian Ocean and Arabian Sea. During the same period, the flux is less over the Bay of Bengal indicating less evaporation. The monsoon season is characterised by increase in latent heat flux in the entire Arabian Sea and Bay of Bengal. During July – September, a zone of large latent flux in the southern tropical Indian Ocean is observed.

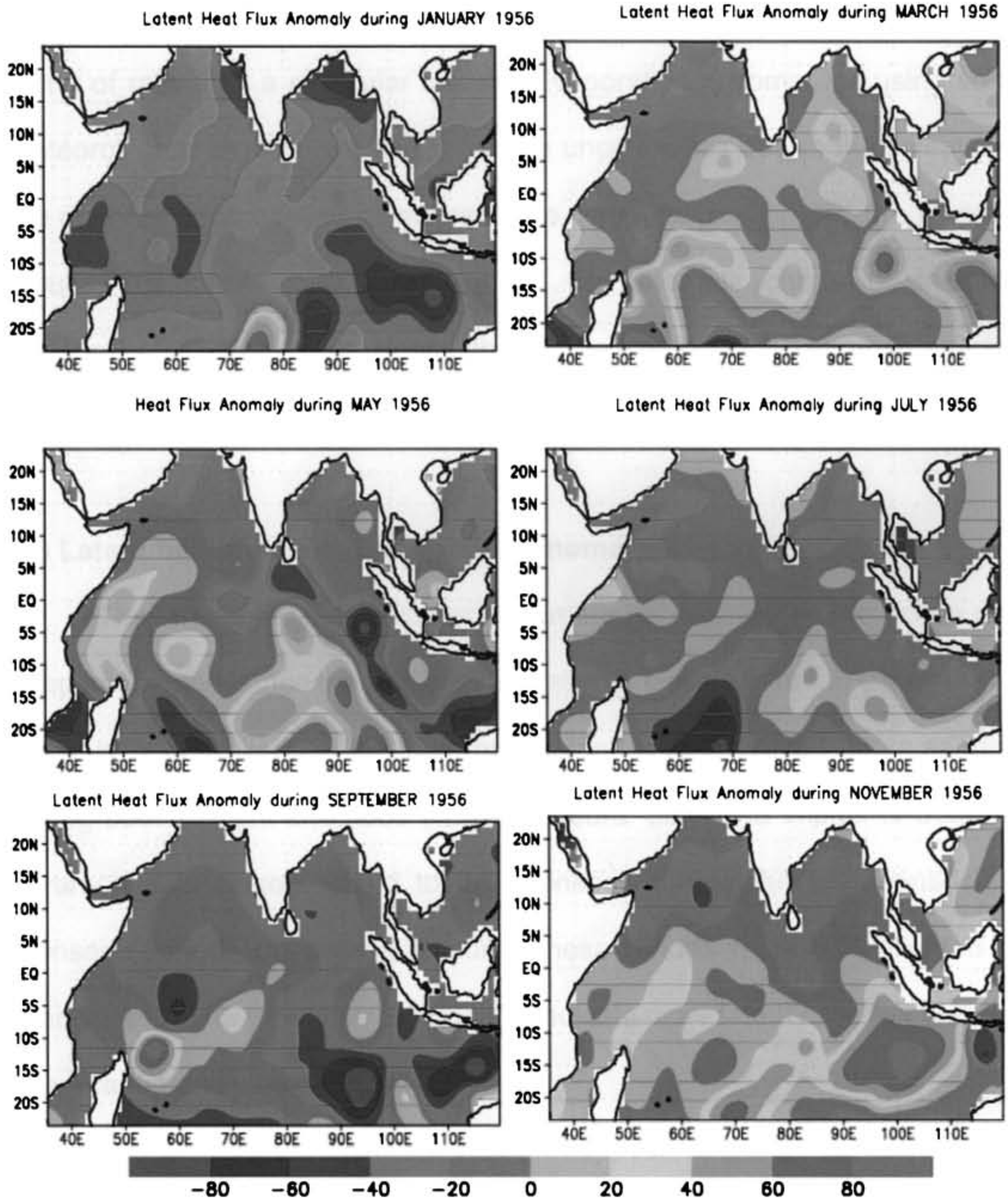


Fig: 4.27 Monthly Mean of Latent Heat Anomaly (w/m^2) in the Indian Ocean during 1956

The amount of evaporation occurring in the ocean ultimately accounts for the rainfall over different parts of the world. The rainfall over the land areas depends on the moisture transported by the wind system and the favourable factor of rainfall in a particular region. Evaporation is computed using surface meteorological parameters and there are uncertainties due to the limitations of the data set used as well as errors in the computational procedure. Hence the results are more noisy compared to the anomaly charts of thesurface meteorological parameters. Hence, though the monthly mean anomaly charts are presented, the results are mostly discussed based on seasonal variations.

4.6 Latent heat flux and Evaporation Anomalies

The latent heat flux indicates the amount of heat lost by the ocean due to evaporation and hence the rate of evaporation is closely linked to latent heat flux. There is lot of similarity in the patterns obtained in their monthly mean anomalies during both drought and flood monsoon years. Since the impact of both these parameters is closely linked to the monsoon during the pre monsoon and monsoon periods, the anomalies during these periods are only highlighted here. A positive anomaly in these parameters indicate more latent heat flux and hence evaporation and vice versa.

In a typical good monsoon year (1956) positive latent heat flux and hence evaporation is maximum in the western tropical Indian Ocean and south Indian Ocean in July (Fig. 4.27 and 4.28) whereas the area of positive anomalies of these features covered a very large area covering the entire north Indian and

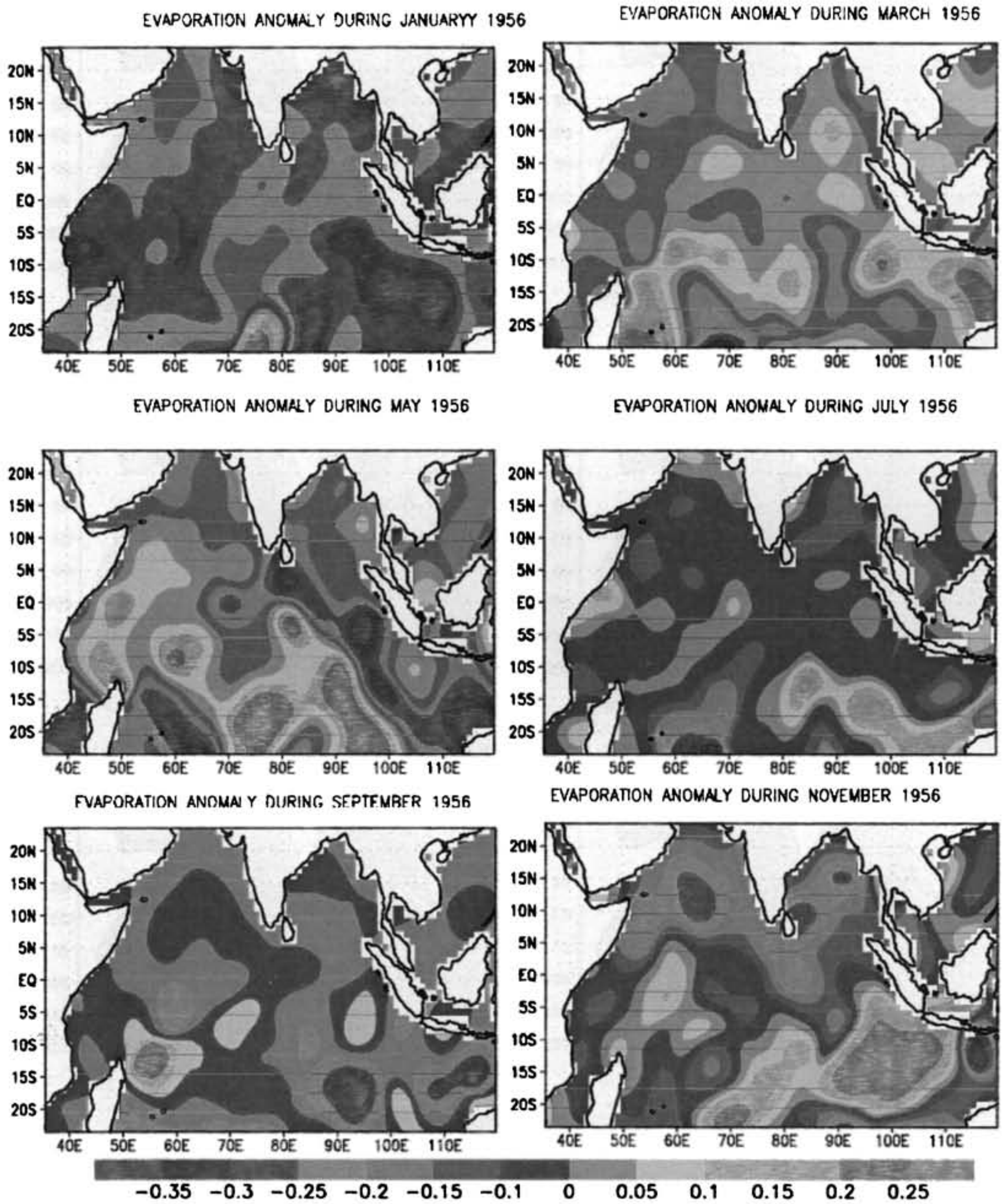


Fig: 4.28 Monthly Mean of Evaporation Anomaly (mm/ 3hrs.) in the Indian Ocean during 1956

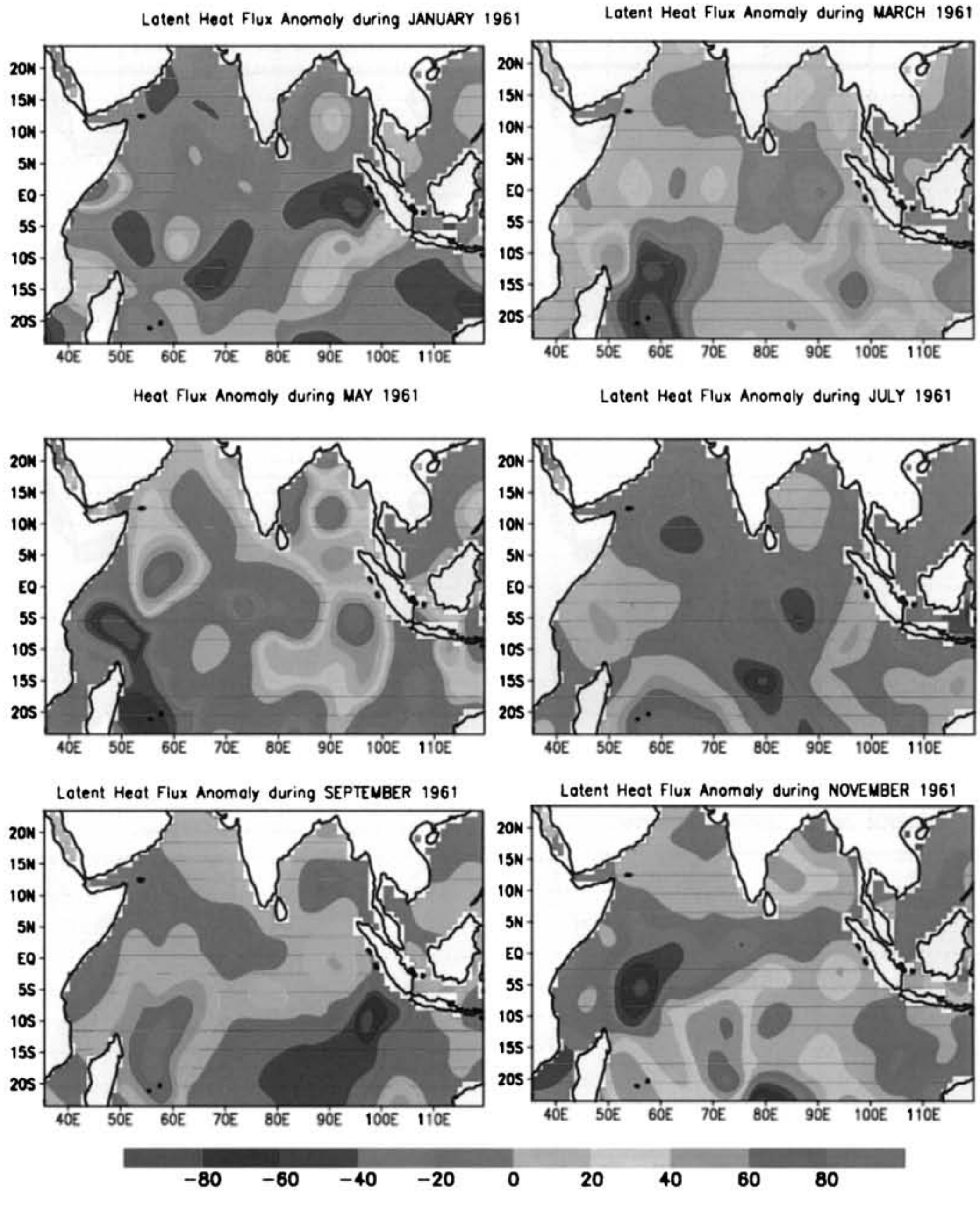


Fig: 4.29 Monthly Mean of Latent Heat Anomaly (w/m^2) in the Indian Ocean during 1961

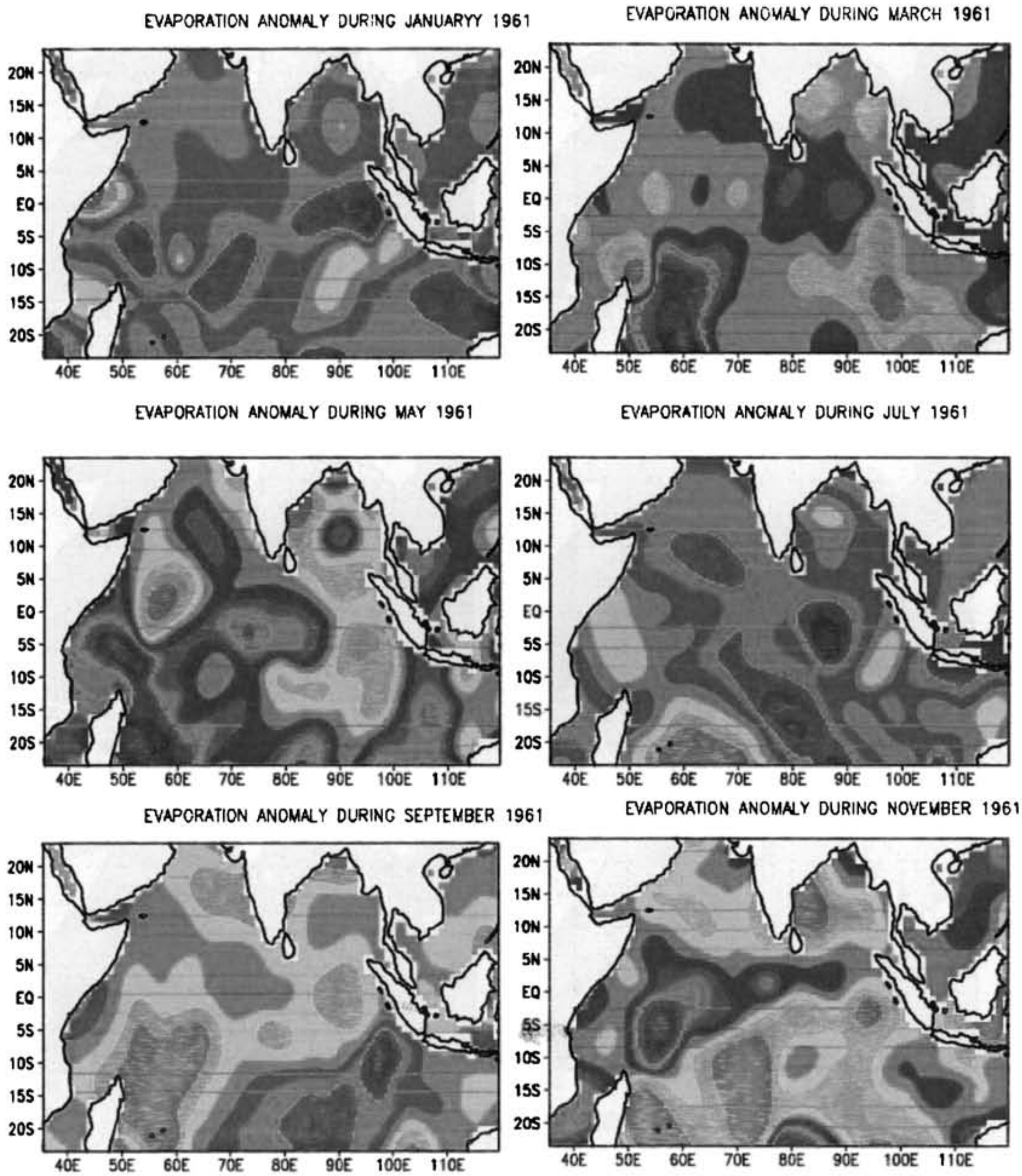


Fig: 4.30 Monthly Mean of Evaporation Anomaly (mm/ 3hrs.) in the Indian Ocean during 1961

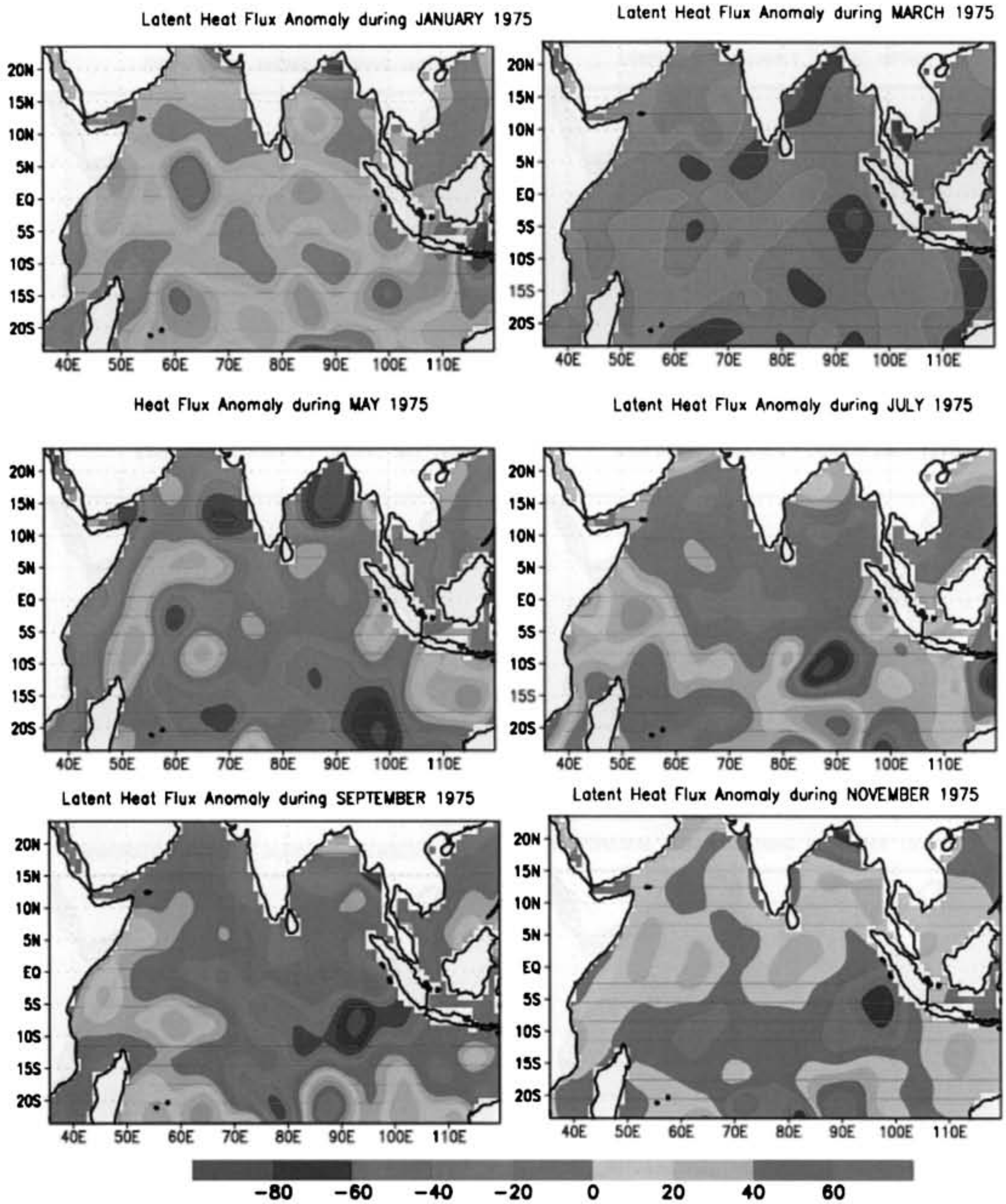


Fig: 4.31 Monthly Mean of Latent Heat Anomaly (w/m^2) in the Indian Ocean during 1975

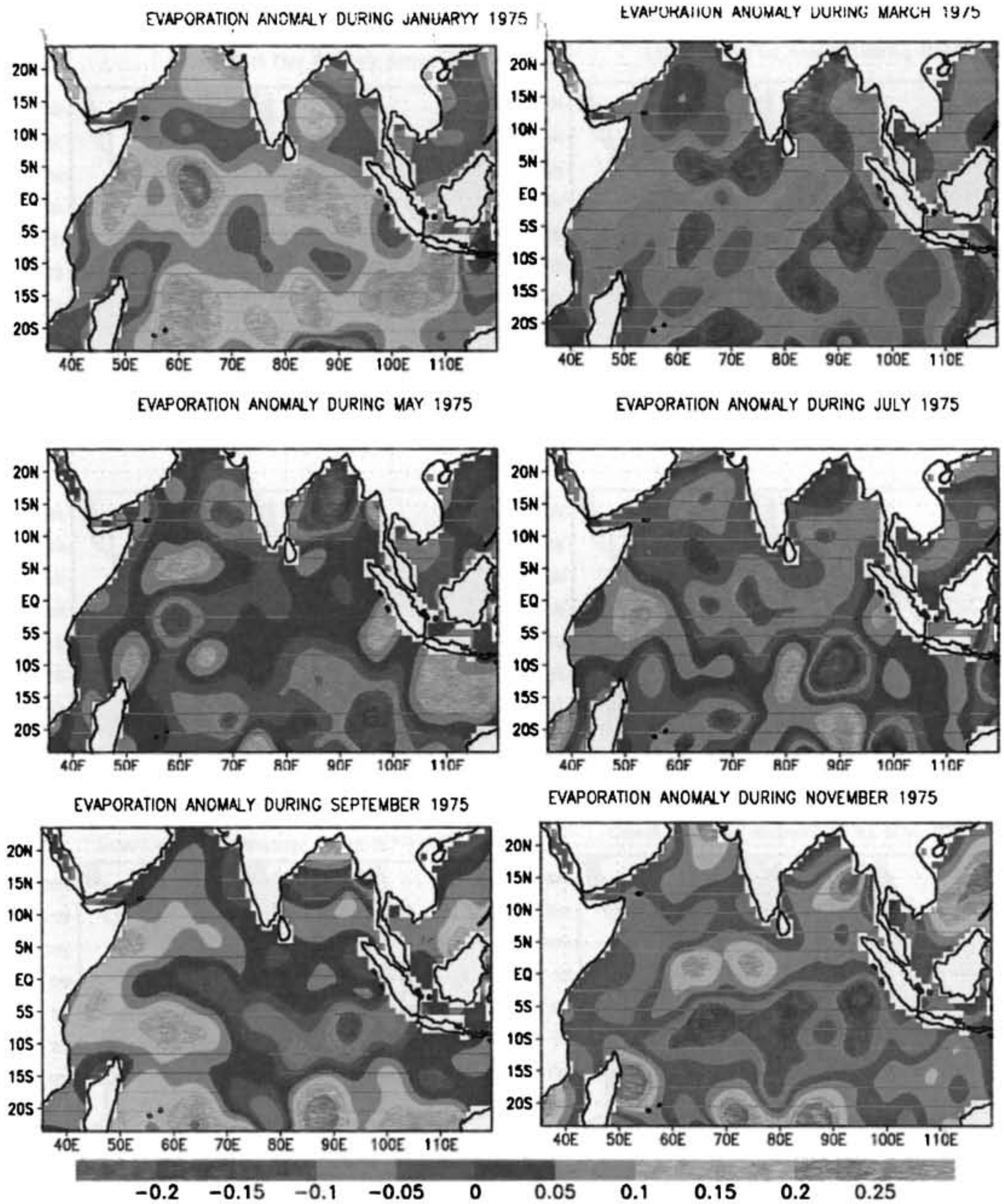


Fig: 4.32 Monthly Mean of Evaporation Anomaly (mm/ 3hrs.) in the Indian Ocean during 1975

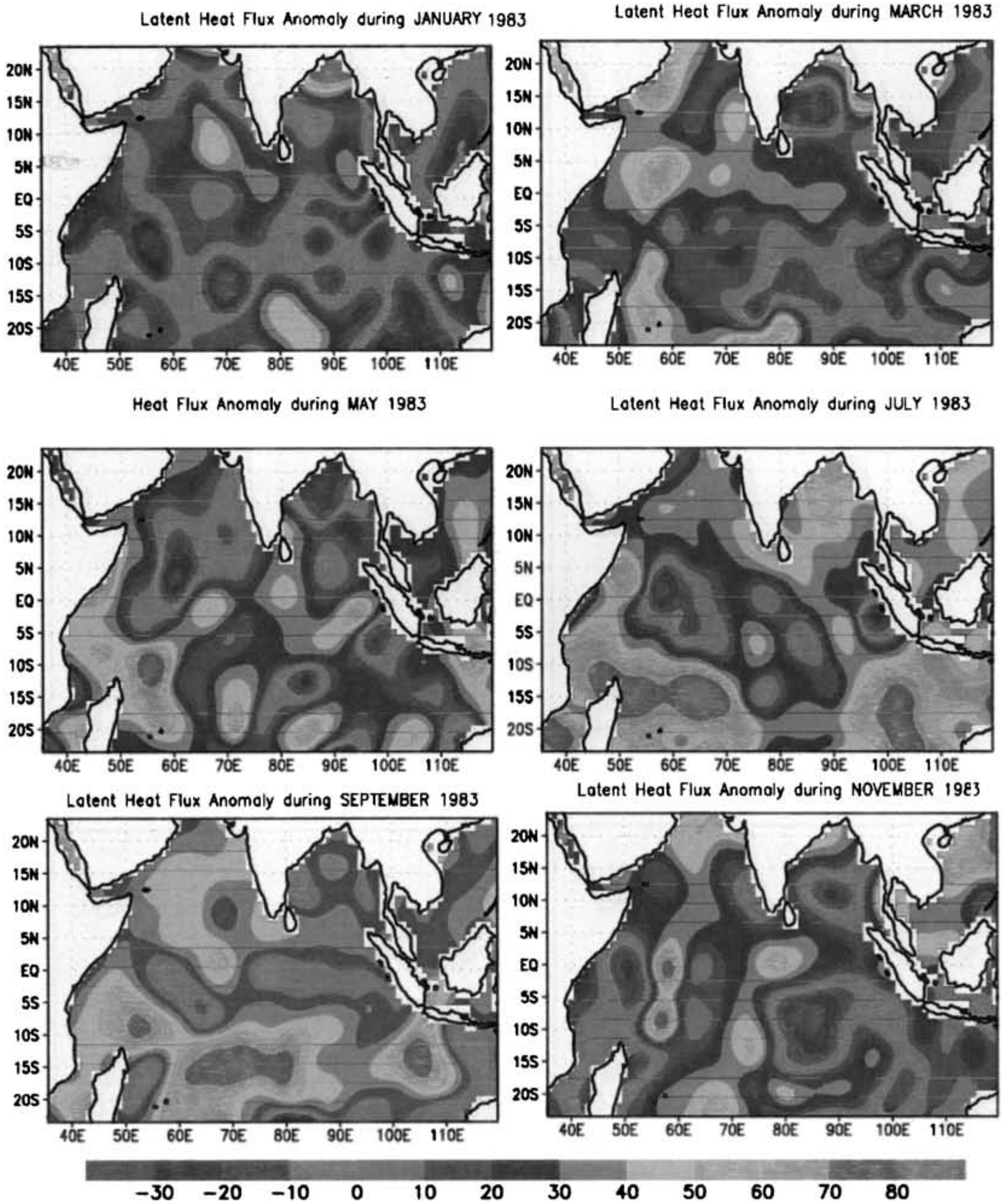


Fig: 4.33 Monthly Mean of Latent Heat Anomaly (w/m^2) in the Indian Ocean during 1983

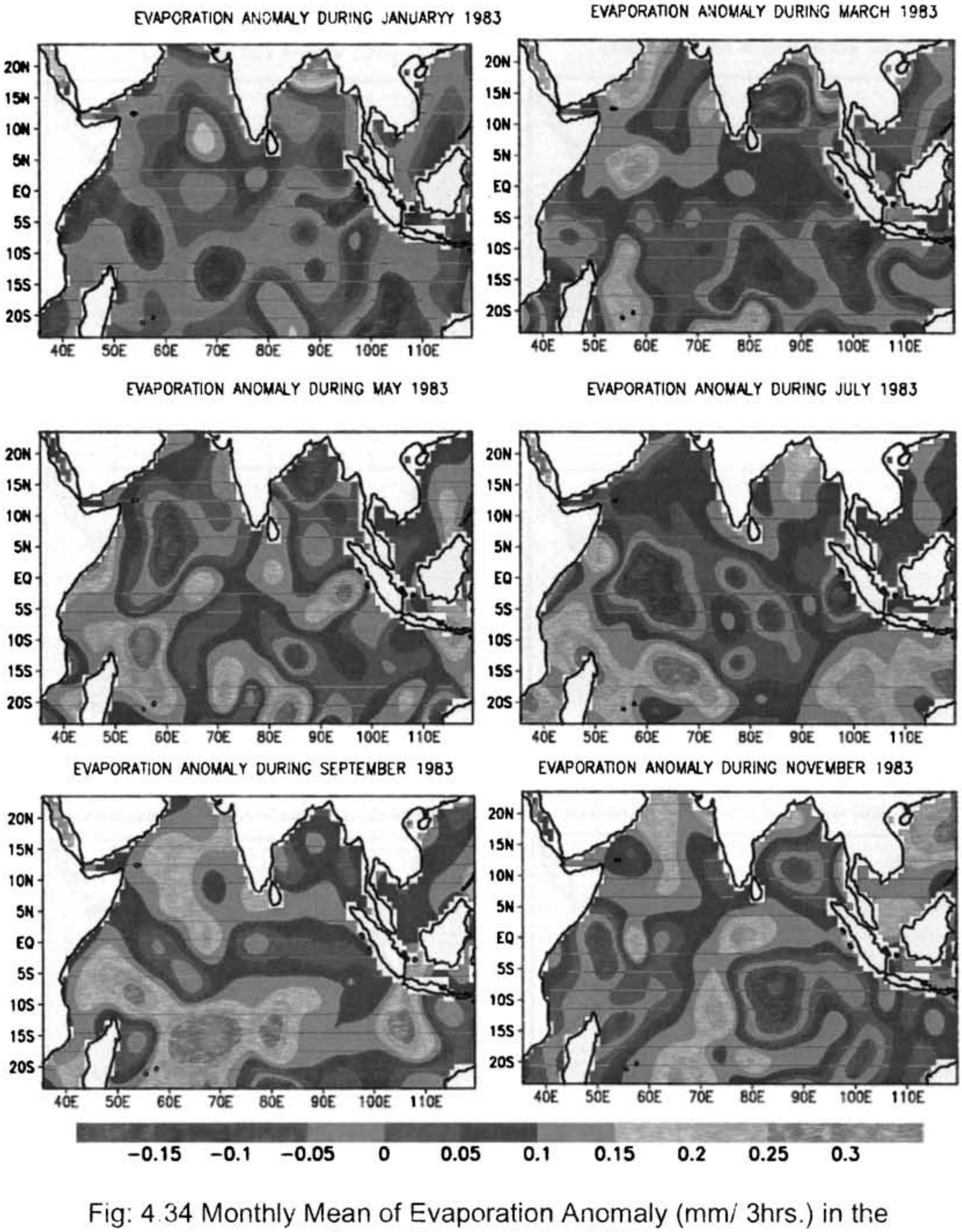


Fig: 4.34 Monthly Mean of Evaporation Anomaly (mm/ 3hrs.) in the indian Ocean during 1983

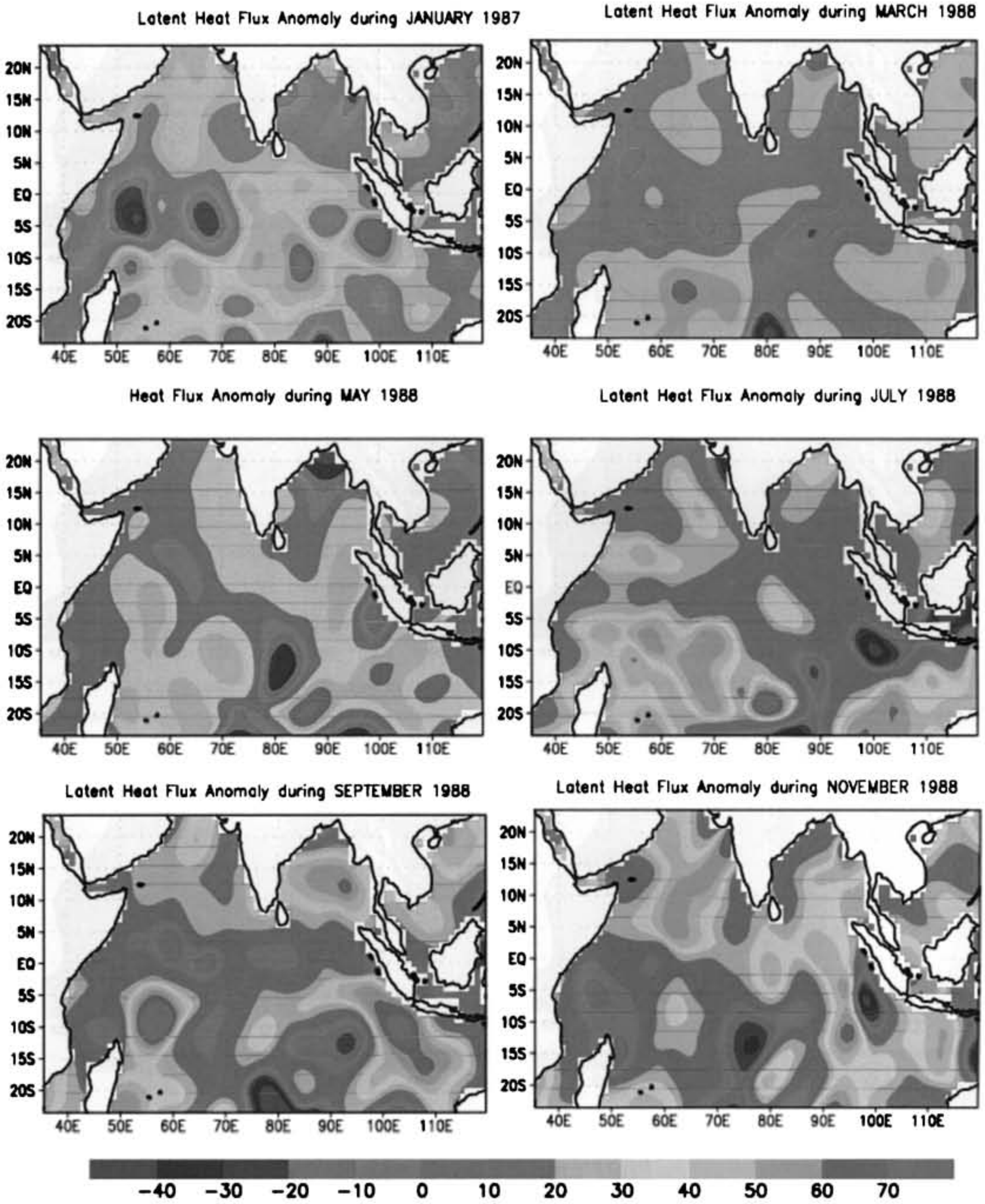


Fig: 4.35 Monthly Mean of Latent Heat Anomaly (w/m^2) in the Indian Ocean during 1988

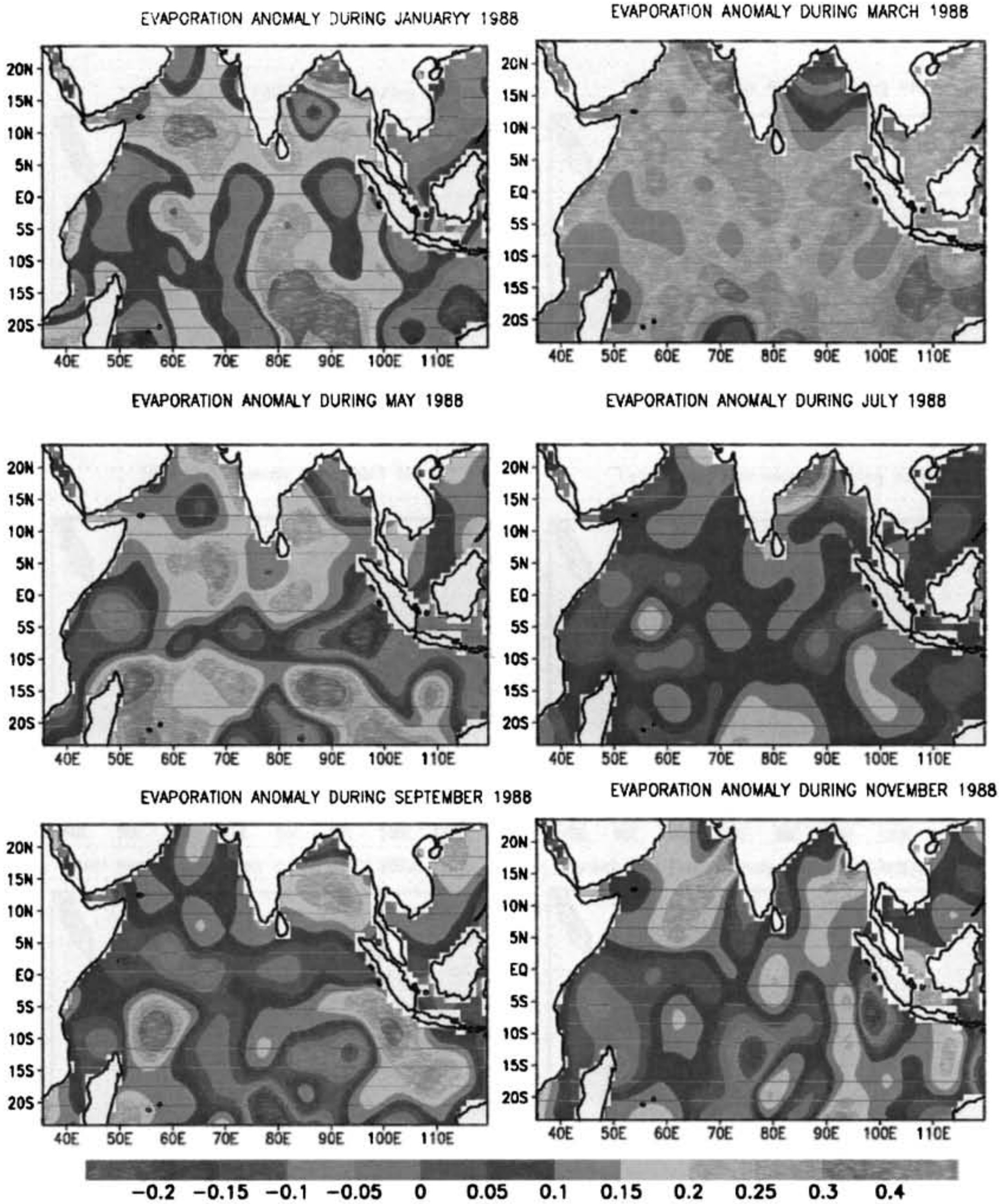


Fig: 4.36 Monthly Mean of Evaporation Anomaly (mm/ 3hrs.) in the Indian Ocean during 1988

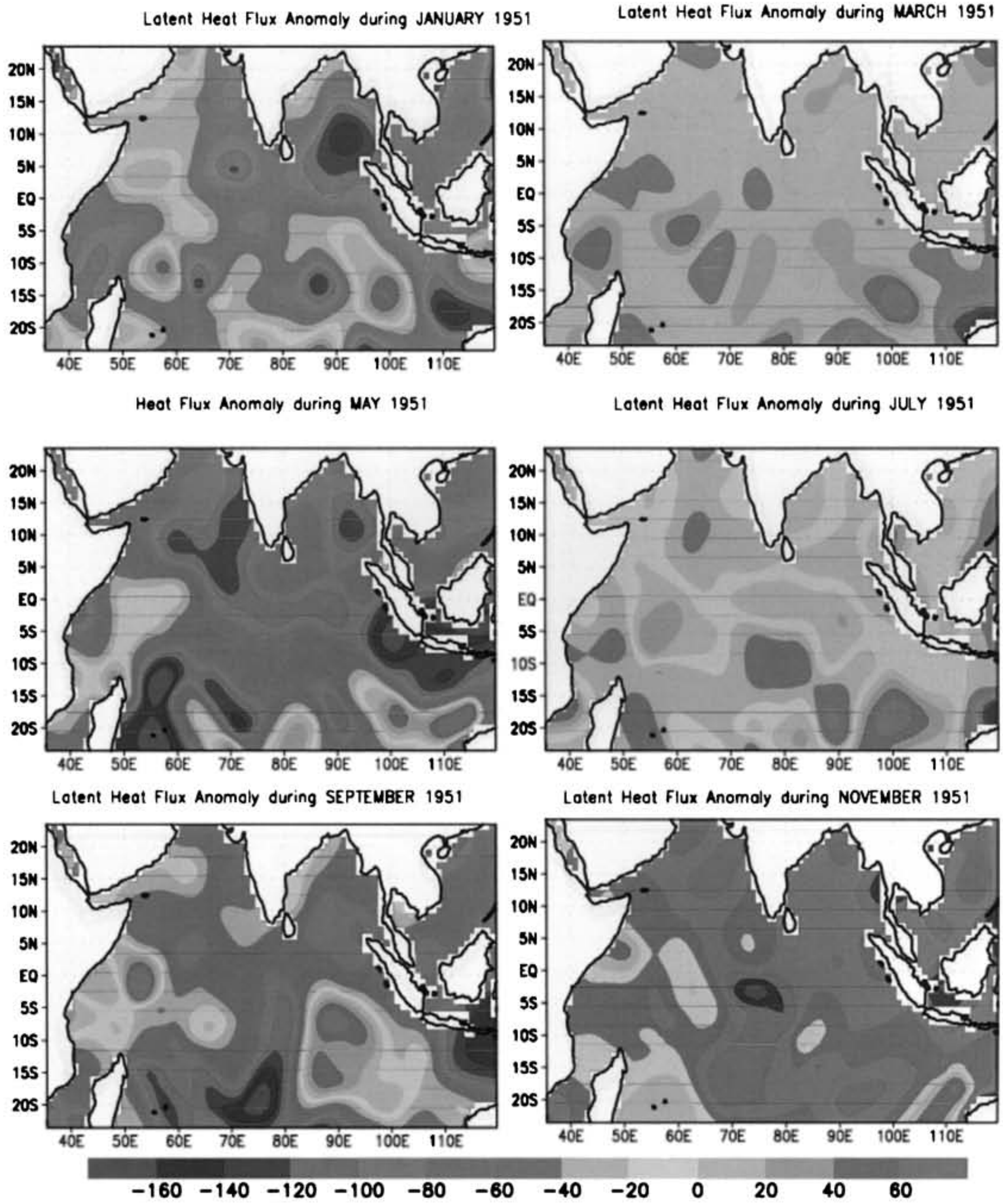


Fig: 4.37 Monthly Mean of Latent Heat Anomaly (w/m^2) in the Indian Ocean during 1951

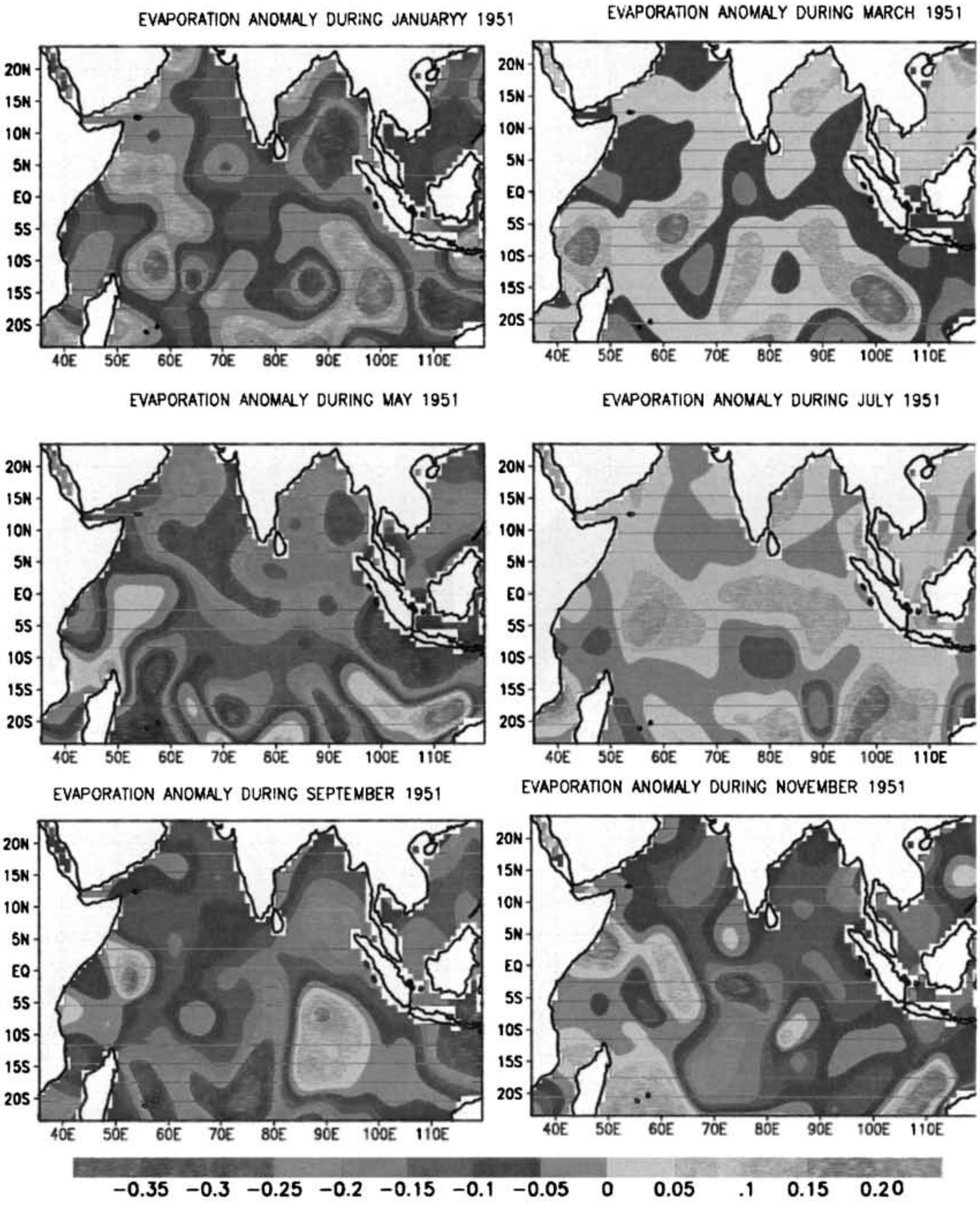


Fig: 4.38 Monthly Mean of Evaporation Anomaly (mm/ 3hrs.) in the Indian Ocean during 1951

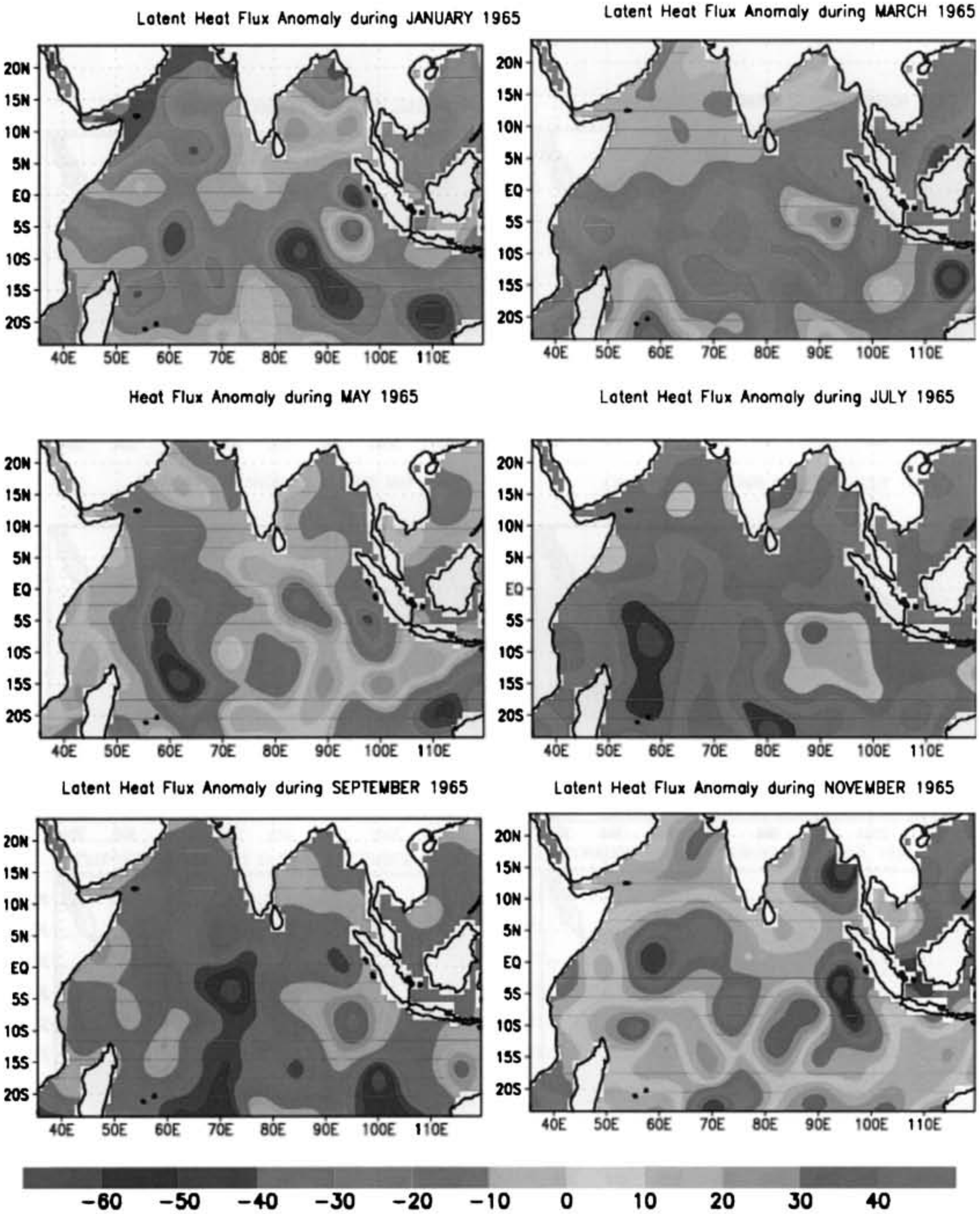


Fig: 4.39 Monthly Mean of Latent Heat Anomaly (w/m^2) in the Indian Ocean during 1965

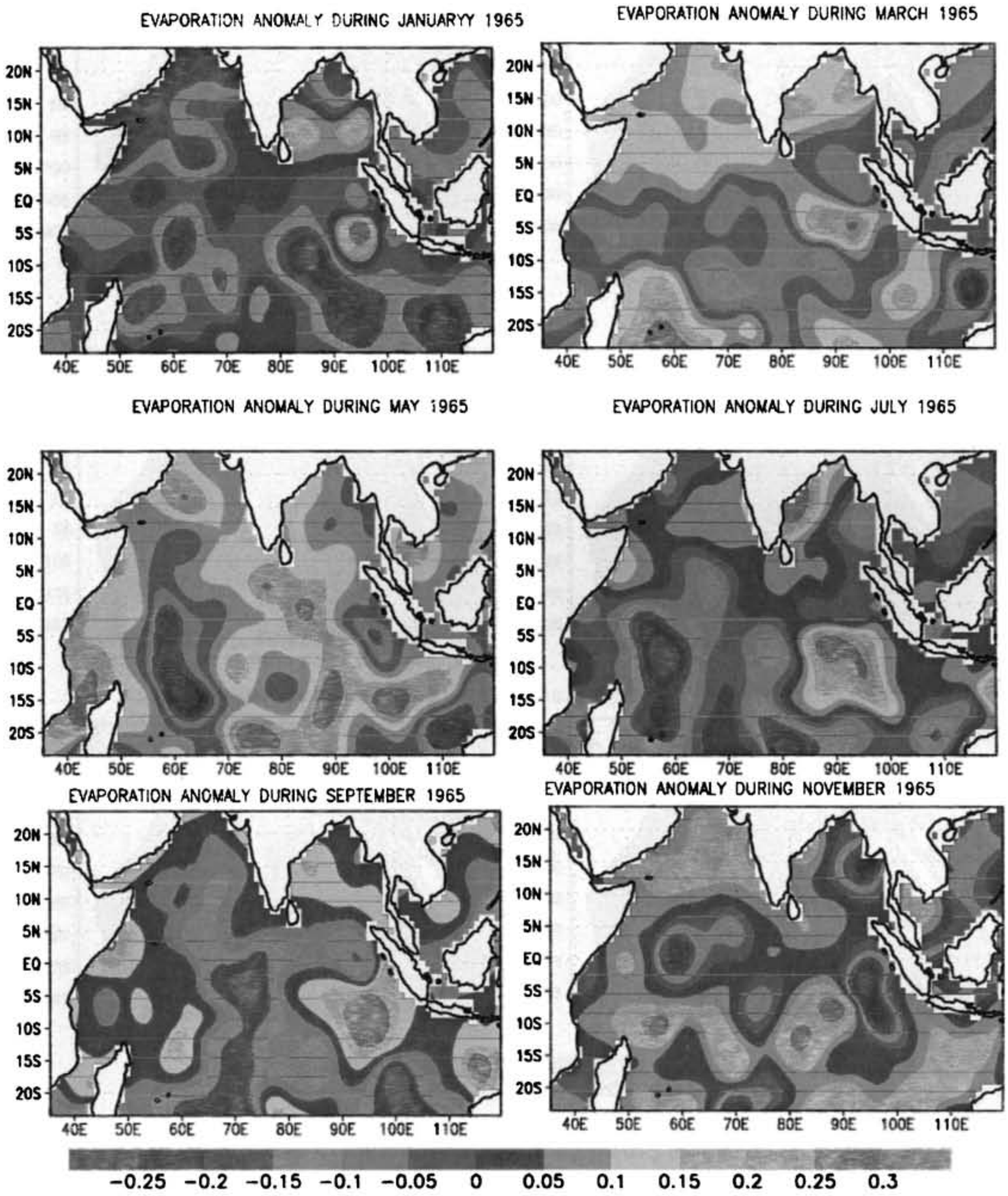


Fig: 4.40 Monthly Mean of Evaporation Anomaly (mm/ 3hrs.) in the Indian Ocean during 1965

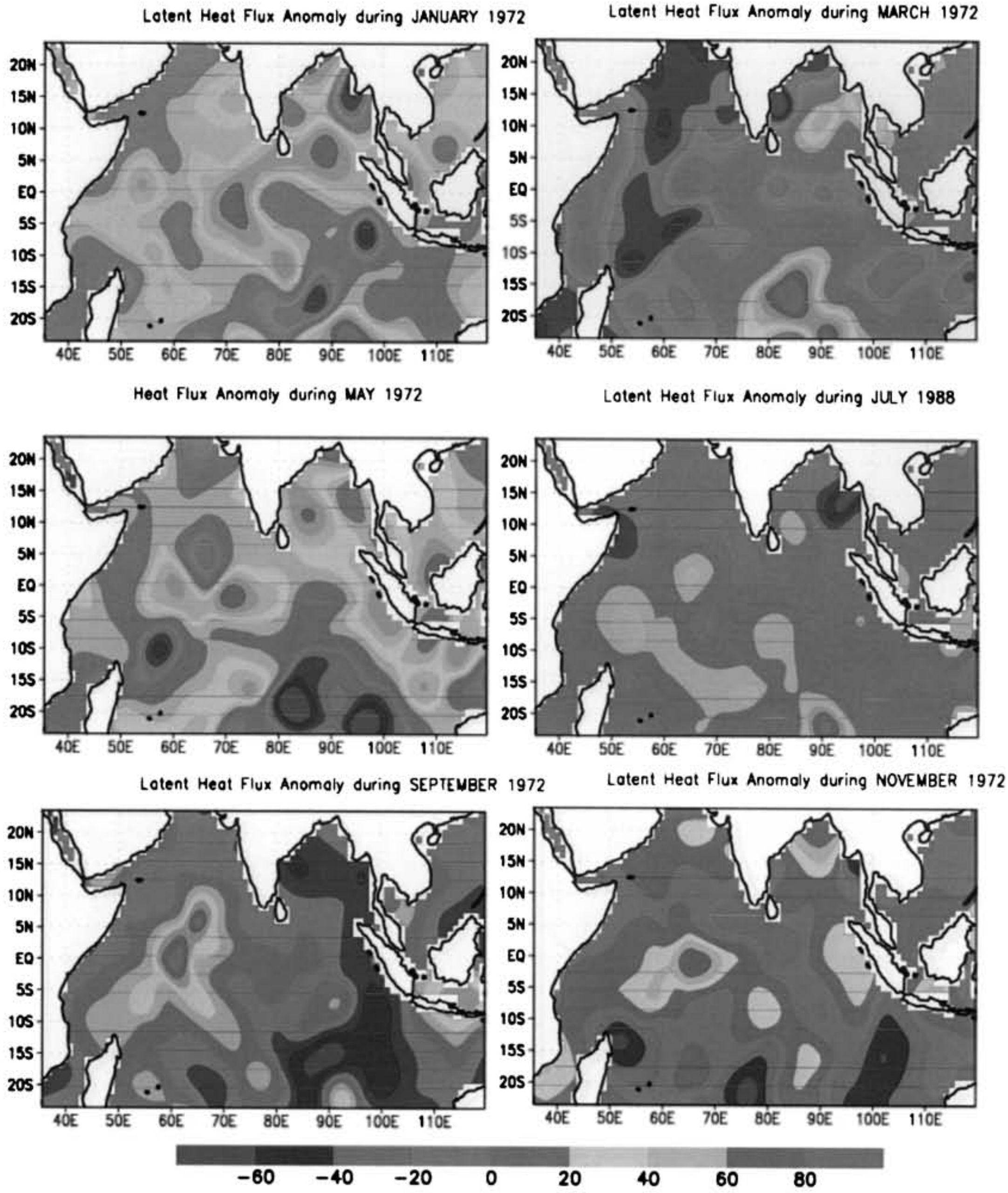


Fig: 4.41 Monthly Mean of Latent Heat Anomaly (w/m^2) in the Indian Ocean during 1972

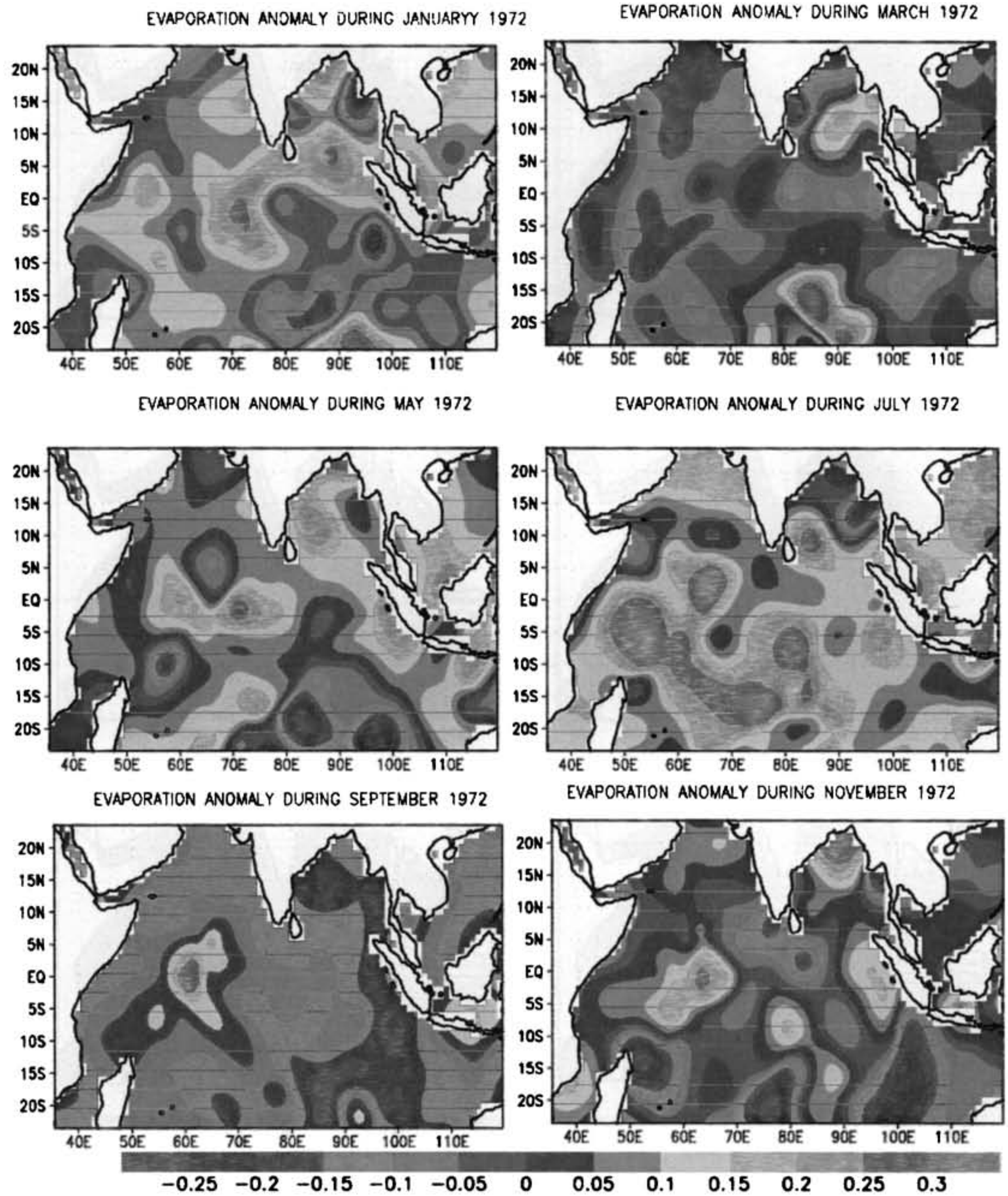


Fig: 4.42 Monthly Mean of Evaporation Anomaly (mm/ 3hrs.) in the Indian Ocean during 1972

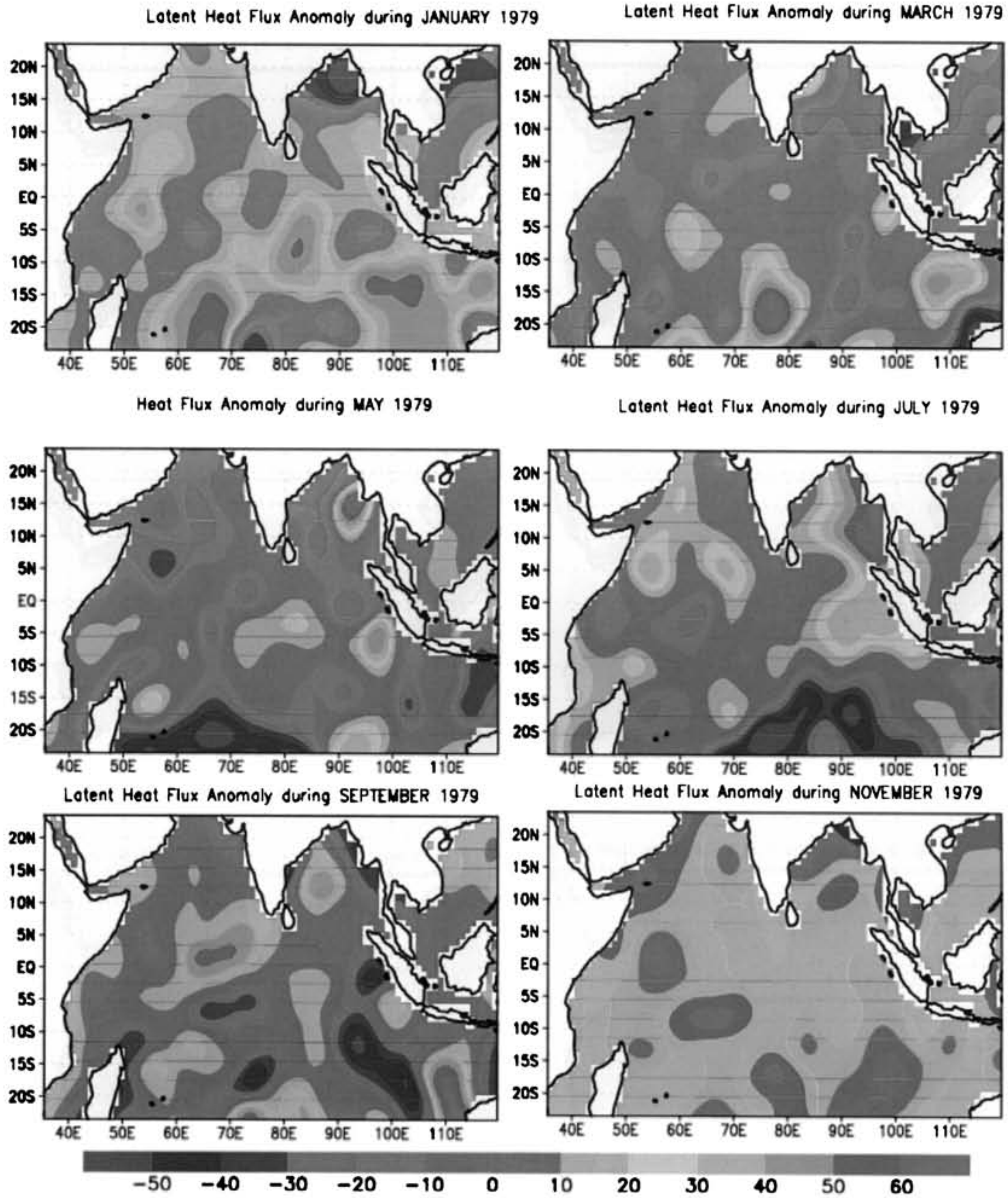


Fig: 4.43 Monthly Mean of Latent Heat Anomaly (w/m^2) in the Indian Ocean during 1979

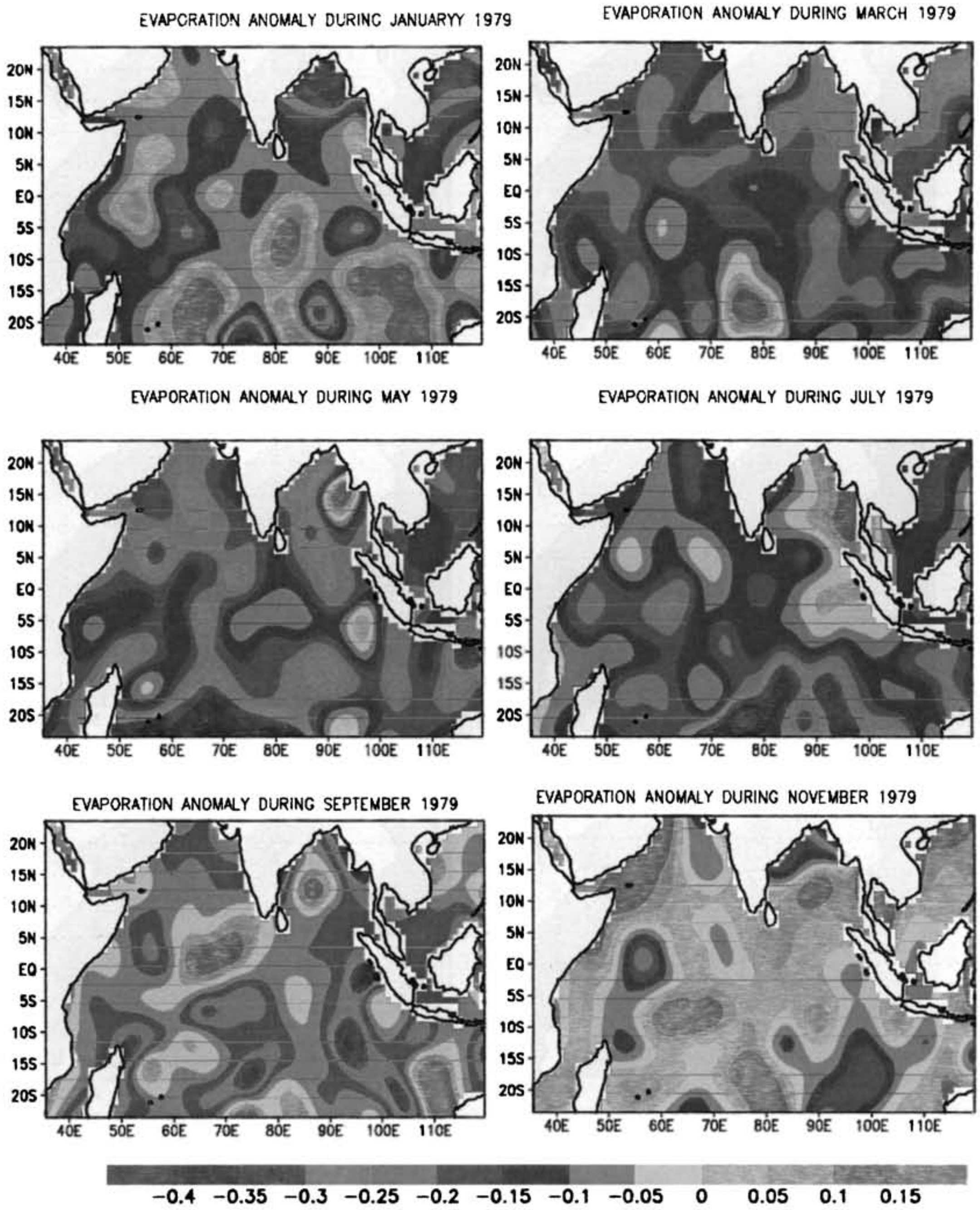


Fig: 4.44 Monthly Mean of Evaporation Anomaly (mm/ 3hrs.) in the Indian Ocean during 1979

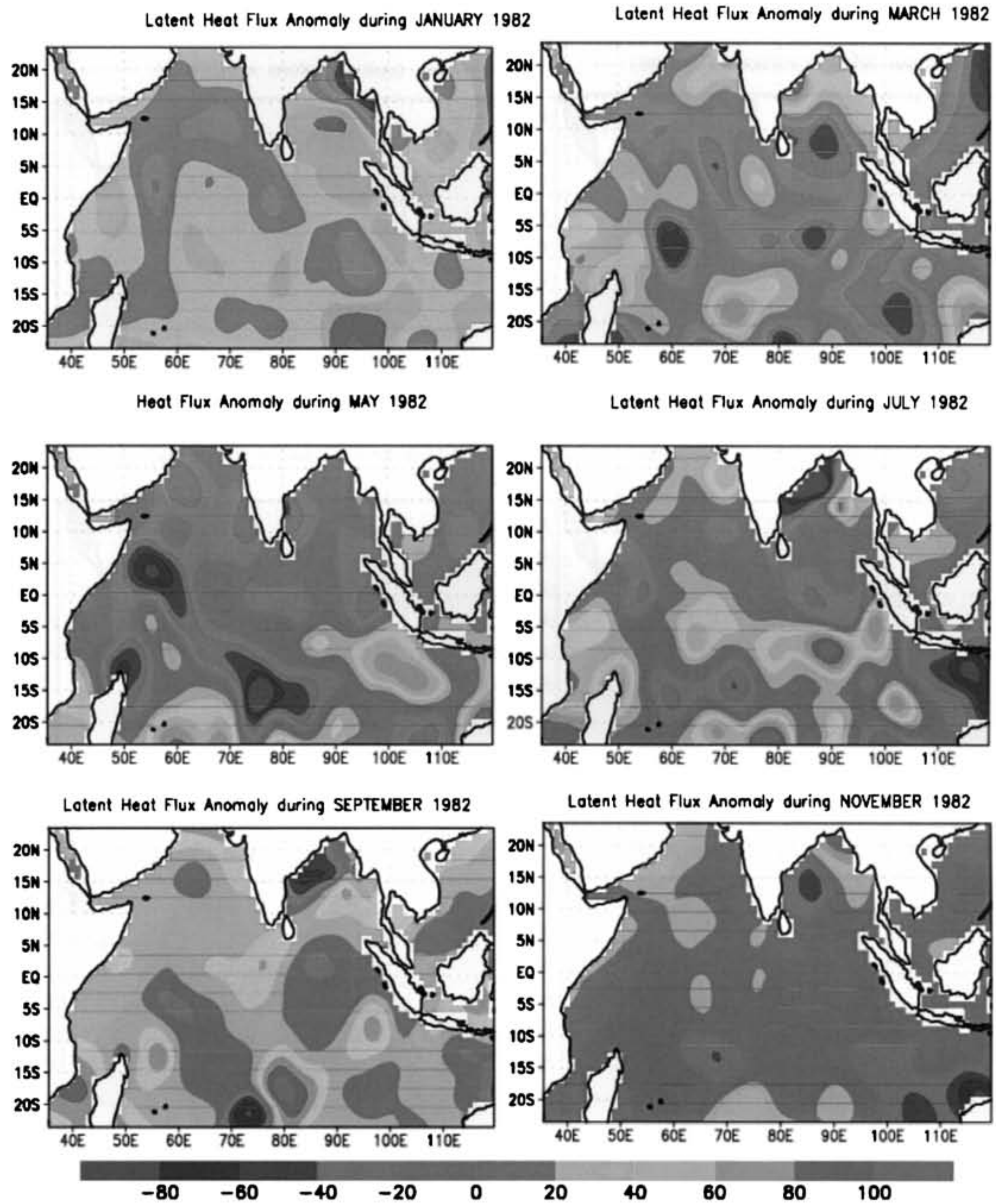


Fig: 4.45 Monthly Mean of Latent Heat Anomaly (w/m^2) in the Indian Ocean during 1982

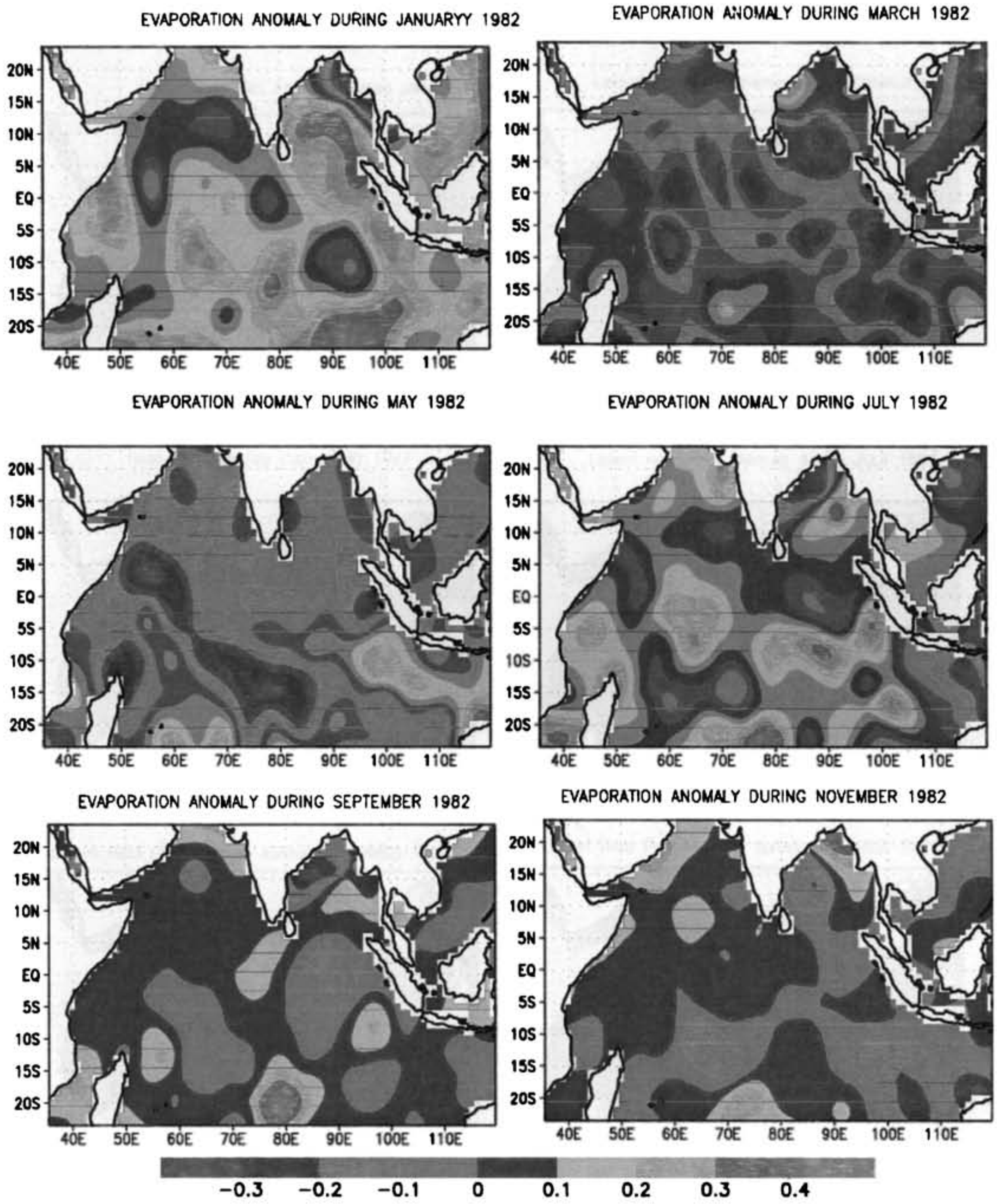


Fig: 4.46 Monthly Mean of Evaporation Anomaly (mm/ 3hrs.) in the Indian Ocean during 1982

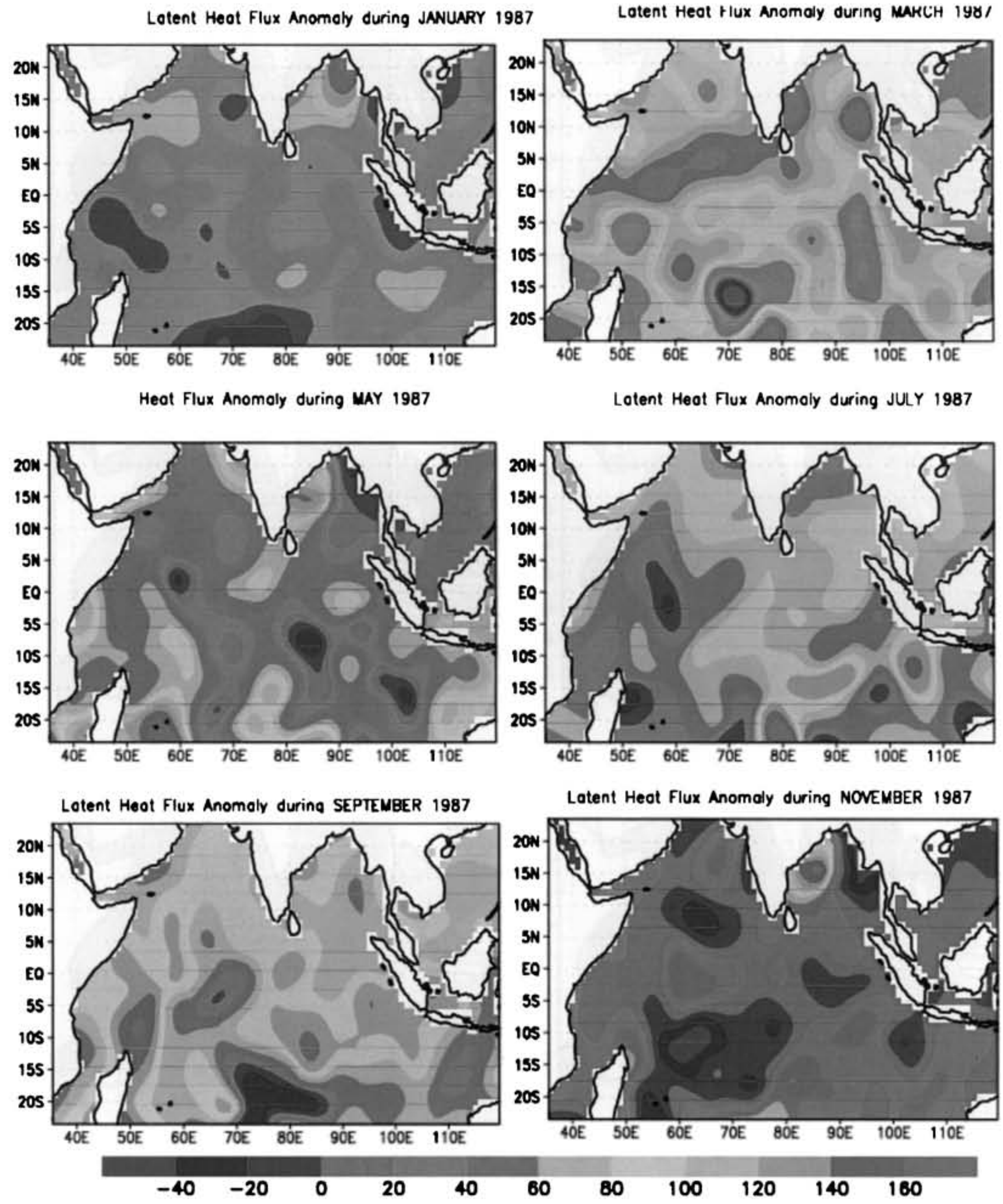


Fig: 4.47 Monthly Mean of Latent Heat Anomaly (w/m^2) in the Indian Ocean during 1987

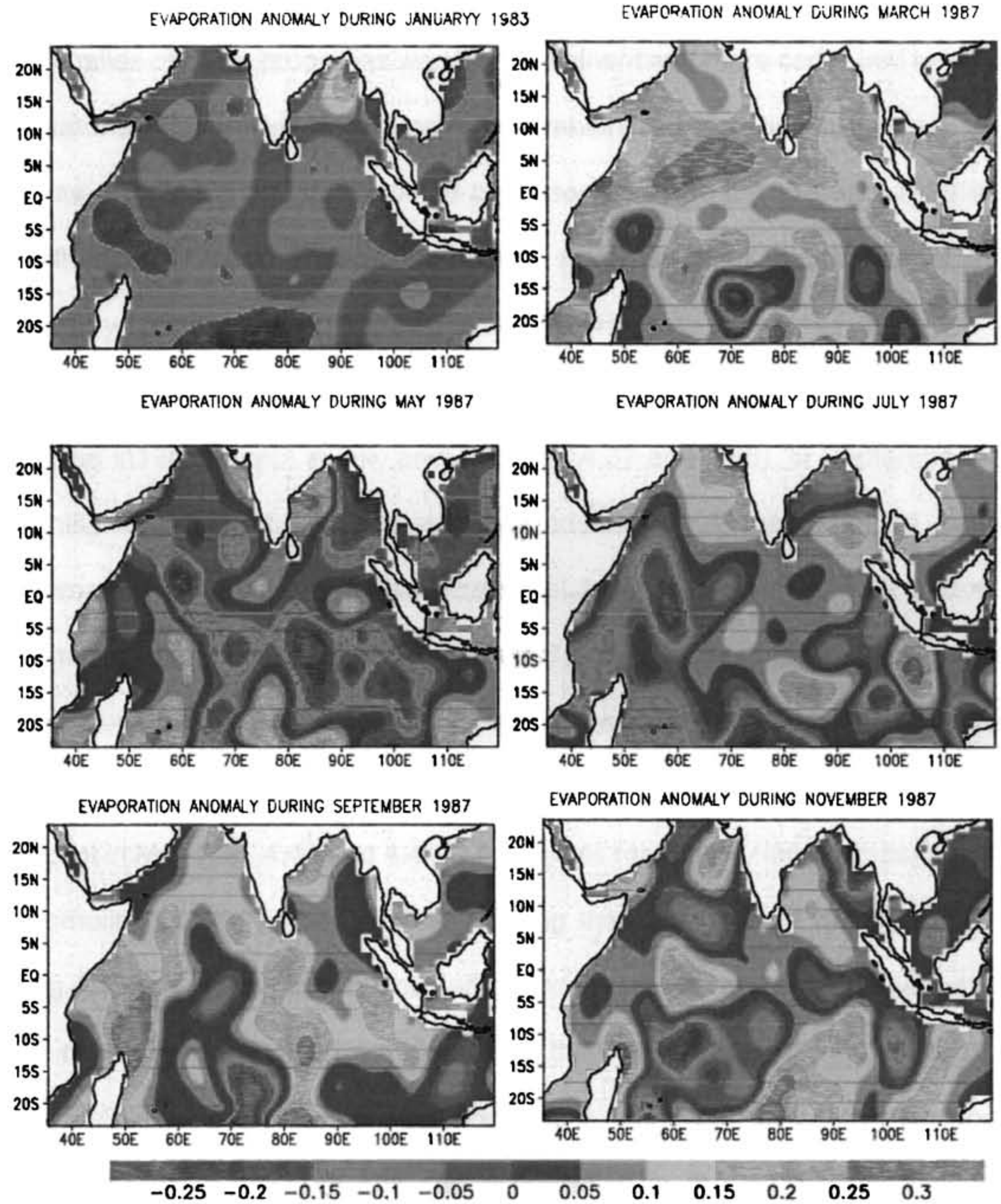


Fig: 4.48 Monthly Mean of Evaporation Anomaly (mm/ 3hrs.) in the Indian Ocean during 1987

south western Indian Ocean in 1961(Fig 4.29 & 4.30), especially in the early and later parts of the monsoon. In 1975(Fig. 4.31 and 4.32) the regions of the positive anomalies of these properties was less prominent and more continued to western equatorial Indian Ocean and parts of Arabian sea and bay of Bengal. Large areas of positive anomalies could be observed in both 1983 and 1988 in the north Indian Ocean, though there were large areas of the equatorial Indian Ocean with negative anomalies (Fig. 4.33 to 4.36).

In contrast to this, large zones of negative anomalies of these properties existed in1951 except in July and July (Fig. 4.37 and 4.38). In September 1965, significant areas of the western equatorial Indian Ocean showed negative anomalies including the central equatorial Indian Ocean (Fig. 4.39 and 4.40). Negative anomalies were prevalent in 1979, also in these areas throughout the monsoon season (Fig. 4.43 and 4.44). The regions of positive anomalies were very limited in the monsoon of 1979. In most regions this was true for 1972 also except in May (Fig. 4.41 and 4.42). There was reasonably large areas of positive anomalies in September 1987 and during the major part of monsoon of 1982 (Fig.4.45 to 4.48), in the tropical Indian Ocean. However, one notable feature is the relatively large negative anomalies in the western tropical Indian Ocean.

One of the most striking features is the occurrence of large positive anomalies of latent heat flux and evaporation in the western tropical Indian Ocean during good monsoon years and large negative anomalies in bad monsoon years. The moisture flux from that region could be one of the main sources of moisture to the ISMR. Moreover, in the central and eastern equatorial

Indian Ocean, the anomalies largely fluctuated in different years. The moisture transport in the equatorial belt associated with wind anomalies may also become the patterns of rainfall over the Indian region. A positive zonal wind anomaly can carry more moisture eastward resulting in more rain fall over eastern parts of Indian Ocean while a negative wind anomaly either inhibit or sometimes even carry moisture from east to west resulting in more rainfall over the Indian region (Gadgil et al, 2003)

A number of studies have been carried out to establish the possible link between SST variation and summer monsoon activity (Shukla, 1987, Rao and Goswamy, 1988). In the present study attempt has been made to establish probable link between the air-sea interface flux and monsoon activity. It is hoped that this attempt would give a complete picture of the link between the air-sea interaction process and summer monsoon activity.

The solar radiative flux pattern mostly controlled by the cloud cover exhibit contrasting pictures during different monsoon regime. In the pre monsoon months, a strong meridional gradient of the flux is seen over the Arabian sea and Bay of Bengal. With the arrival of summer monsoon the distribution of flux is mostly in the zonal direction. The radiative flux is mostly determined by the cloud cover under different monsoon regimes. The flux pattern exhibits an appreciable variability with excess monsoon year characterised by reduced incoming flux and vice versa.

The latent heat flux pattern and evaporation also exhibit similar variabilities. In the excess rainfall years, largest values are noticed over the tropical Indian seas due to increased wind speed.

Net heat flux is a balance between short wave flux and latent heat flux. It suggests that gives a measure of the monsoon activity over the Indian sub-continent. The tropical Indian seas are found to lose heat more during excess rainfall years compared to deficient rainfall years. This indicates that the air sea interaction prior to a good monsoon activity is increased leading to more exchange of fluxes of momentum, heat and moisture. These fluxes are crucial to drive the ocean-atmospheric models. It is felt that the net oceanic flux over the tropical Indian sea prior to monsoon onset (in May) may be considered as an advance indicator of the possible behaviour of the subsequent monsoon season. Therefore, a detailed knowledge of the variability of air-sea fluxes of heat and moisture will help in establishing a possible link between them and the monsoon activity. Further it can be seen that prior to excess rainfall year, tropical Indian seas lose heat more driving the convection activity in the atmosphere. The main hindrance in establishing this link is that adequate observational networks are not available over the tropical Indian seas. However, the present work, which provides certain crucial inputs, required for the monsoon prediction to a certain extent. This objective can be achieved in full if the observational network over the tropical Indian seas are strengthened. However, with the advance of remote sensing satellite and data buoys, there has been a spurt in the collection of observational data over the Indian seas.

CHAPTER V

EQUINOO, ENSO AND ISMR ANOMALIES (1945-1979)

5.1 Introduction

Recent advances in the understanding of the Indian Ocean and associated variability led to better understanding of Indian monsoon rainfall (Kumar et al 1999, Reason et al 1999, Saji et al 1999, Webster et al 1999; Murtugudde et al, 2000; Vinayachandran et al , 2002; Behera and Yamagata, 2003). The relationship between El Nino, La Nina, IOD and the more recently observed Equatorial Indian Ocean Oscillation (EQUINOO) on the Indian summer monsoon rainfall can lead to better understanding and better prediction of ISMR. Kumar et al (1999) observed that the relationship between ISMR and El Nino has weakened in the recent decades. For 14 consecutive years beginning 1988, there was no drought despite the occurrence of El Nino in 1992-93 and 1997-98 while El Nino of 1997 – 98 had been the strongest of the last century.

Gadgil et al (2003) have shown that the patterns of outgoing long wave radiation (OLR) anomalies for July 2002 and July2003 were markedly different over the Indian Ocean, out of which 2002 was severe drought year and 2003 was good monsoon year over India. Now cloud cover over the land and ocean can be obtained from satellites, especially from daily maps of outgoing long wave radiation (OLR). Low values of OLR correspond to emission from higher levels and hence higher cloud tops (and more rain when the clouds are deep) and OLR is used as a proxy for estimation of rainfall in the tropics (Gadgil et al 2003).

Large negative (positive) anomalies in OLR are associated with large positive (negative) anomalies in rainfall over a region. Gadgil et al (2003) reported that large anomalies of OLR over most of the Indian region west of 60°E appear to be associated with anomalies of the sign over the eastern equatorial Indian Ocean. The contrasting patterns of 2002 and 2003 suggest the possibility of a link of the variation of ISMR with the variation of deep convection over the equatorial Indian Ocean.

Prior to the late 1980's, there were not many satellites observations of surface meteorological parameters especially those of OLR. Hence, in this chapter, the observed cloud anomalies are used in place of OLR. The influence of ENSO events and EQUINOO on ISMR is analysed for the period 1945-1979 and the analysis of Gadgil et al (2003) is further extended to the pre-1979 period. Except for OLR anomaly, the other meteorological parameters utilised are the same. The zonal wind anomaly averaged over 2.5°N – 2.5°S and 60 – 90°E is taken as the indicator of EQUINOO in the Indian Ocean.

Between 1945-1980, the All India Summer Monsoon Rainfall (AISMR) was very low in 1951, 1965,1966,1968,1972,1974 and 1979. The ISMR showed significant surplus in 1956, 1961,1970 and 1975. During these years EL Nino occurred in 1951,1957,1965,1969,1972 and 1976. In 1969 and 1976 eventhough there was El Nino monsoon was normal. Despite the absence of El Nino, monsoon failed in 1966,1968,1974 and 1979. Hence in this chapter the ENSO and EQUINOO influence is studied during the years shown in table 5.1.

Drought Years	Flood Years	El Nino Years
1951		1951
	1956	
		1957
	1961	
1965		1965
1966		
		1969
	1970	
1972		1972
1974		
	1975	
		1976
1979		

Table 5.1 Drought years, Flood years and El Nino years between 1945 -1980

An unfavourable EQUINOO on ISMR is characterised by positive OLR anomalies (less cloud cover) over the western equatorial Indian Ocean and negative OLR anomalies (increased cloud cover) in the eastern tropical Indian Ocean, while a ISMR favourable EQUINOO is characterised by negative OLR anomalies (increased cloud cover) in the western tropical Indian Ocean and positive OLR anomalies (less cloud cover) in the eastern tropical Indian Ocean during the southwest monsoon season. This has been demonstrated for the years 2002 and 2003 when this contrast was very clear (Gadgil et al 2003).

Since the OLR is not available for the period 1945-1978, the observed cloud anomalies for central pacific is used as an indicator of ENSO are used in this chapter. Moreover the impact of El Nino on ISMR is correlated to the equatorial pacific Ocean OLR /cloud anomalies. Gadgil et al (2003) used OLR anomalies of the same region in their analysis because this parameter is more correlated to ISMR than eastern Pacific OLR anomalies.

5.2 Indian summer Monsoon Rainfall (ISMR) and Indian Ocean Dipole (IOD)

The Indian Ocean Dipole (IOD) is characterized by anomalies of opposite sign in the SST and rainfall over the western part of equatorial Indian Ocean (50° and 70°E, 10°S-10°N) relative to that over the eastern tropical Indian Ocean. (90°-110°E; 0°-10°S). The IOD peaks in the autumn, after the summer monsoon season. The relationship of IOD and rainfall over the Indian region and surrounding areas has been studied using the dipole mode index (DMI) based on the difference of the SST anomalies in the western equatorial Indian Ocean (Ashok et al, 2001) and Eastern equatorial Indian Ocean while DMI is well correlated with rainfall over eastern Africa and western equatorial Indian Ocean. The correlation with ISMR is not significant (Saji et al 1999), but the reduced impact of El Nino on 1997 monsoon could be associated with positive IOD (Behra et al, 2000; Ashok et al 2001). It has been observed that while ISMR anomalies for all pure positive dipoles (in the absence of El Nino) are positive, for pure negative dipoles (in the absence of La Nina) the ISMR anomaly is negative in only one out of the five cases (Gadgil et al 2003). Then it is not clear whether any strong relationship exists between IOD and ISMR. There are many years

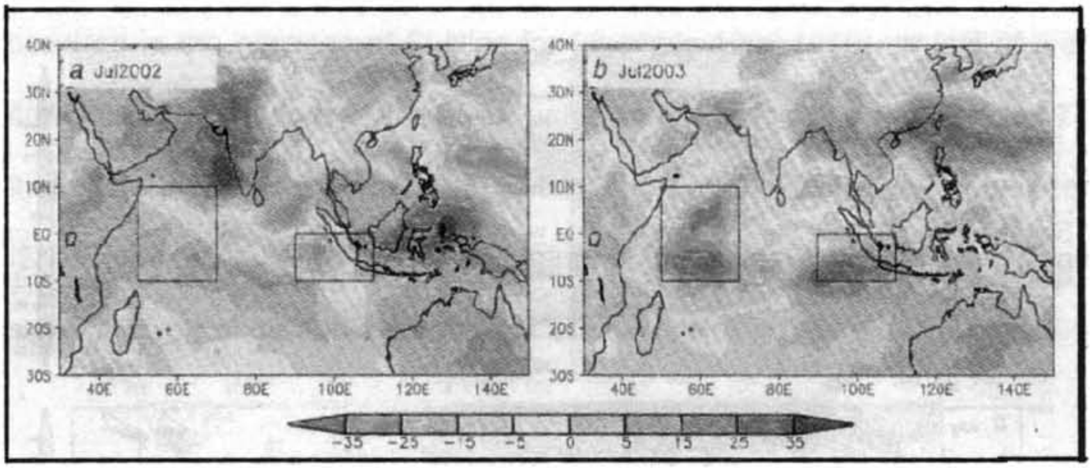


Fig: 5.1 OLR anomaly for (a) July 2002 and (b) July 2003. Note that positive anomaly of OLR corresponds to negative anomaly of rainfall. After Gadgil et al 2003

Similar results were also noticed in 1986 and 1994and also in1987 and 1988

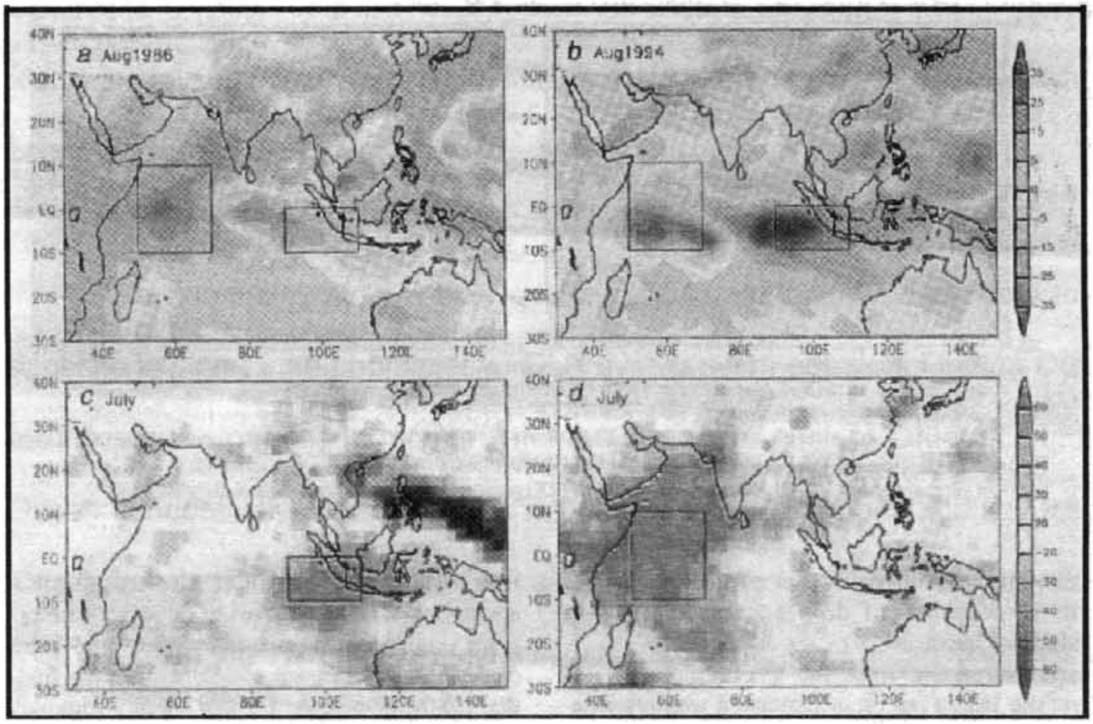


Fig: 5.2

Caption?

ISMR was large in the absence of La Nina (example 1961 and 1994) and drought occurred in the absence of El Nino for (example 1966,1979). In fact of the 22 droughts occurred during1871-2001, only 11 were associated with El Nino. Thus it appears that process other than El Nino/La Nina play significant role in large ISMR anomalies. The events occurring in equatorial Indian Ocean play a significant role in droughts/excess rainfall seasons over India (Gadgil et al 2003).

5.3 Equatorial Indian Ocean Oscillation and ENSO

The OLR pattern of a typical drought year (example 2002) and a good monsoon year (2003) is similar in several typical monsoon seasons (OLR anomalies over the western part of the equatorial Indian Ocean being opposite to that over the eastern part). The OLR over the eastern equatorial Indian Ocean is found to be negatively correlated with that over the Bay of Bengal, while the correlation of the OLR over the western equatorial Indian Ocean is high over that of the eastern Arabian Sea.

The anomalies of OLR over eastern equatorial Indian Ocean tend to be opposite in sign to the anomalies of OLR over western equatorial Indian Ocean for the summer monsoon months, and the OLR over eastern equatorial Indian Ocean is negatively correlated with that over western equatorial Indian Ocean. This is well demonstrated in the OLR anomaly fields for July 2002 and July 2003 (Fig. 5.1) with totally contrasting fields such contrasting OLR anomalies over the equatorial Indian Ocean was seen in August 1986 and August 1994 (Fig. 5.2) typical in two contrasting monsoon years. The OLR anomalies in July 1987 and

August 1988 (Fig.5.3) also showed contrasting patterns over the eastern and western equatorial Indian Ocean. The anomalies of sea level pressure and zonal wind along the equator are also consistent with this information. Therefore, when deep convection is enhanced over western equatorial Indian Ocean, the pressure is reduced at the same time, deep convection is reduced over eastern equatorial Indian Ocean and the pressure increases. This pressure gradient (with pressure increasing from west to east) is associated with easterly (from east to west) wind anomalies in the zonal wind. On the other hand, when deep convection is enhanced over eastern equatorial Indian Ocean, there are westerly zonal wind anomalies along the equator. Gadgil et al (2003) called this oscillation as Equatorial Indian Ocean Oscillation (EQUINOO). An index of EQUINOO is selected as the anomaly of zonal component of surface wind at the equator (60°-90°E; 2.5°S -2.5° N), which is highly correlated with the difference in OLR of western equatorial Indian Ocean and eastern equatorial Indian Ocean. The zonal wind index (EQWIN) is taken as negative of the anomaly.

The anomalies of SST, OLR and wind are inter -related to both EQUINOO and IOD coupled ocean-atmosphere system. For the monsoon season (June-September) there is high correlation between Dipole mode index (DMI) and EQWIN. However there is strong relationship between EQWIN and ISMR compared to DMI and ISMR. Based on DMI (Yamagata et al 2001), the ISMR anomaly was negative only in one out of the pure negative dipoles. There is a one to one correspondence between large positive (negative) ISMR anomaly (Gadgil et al 2003). This difference are due to the fact that large values of

EQWIN and DMI do not always occur in the same seasons, since there are lags between changes that occur in the atmosphere and oceans.

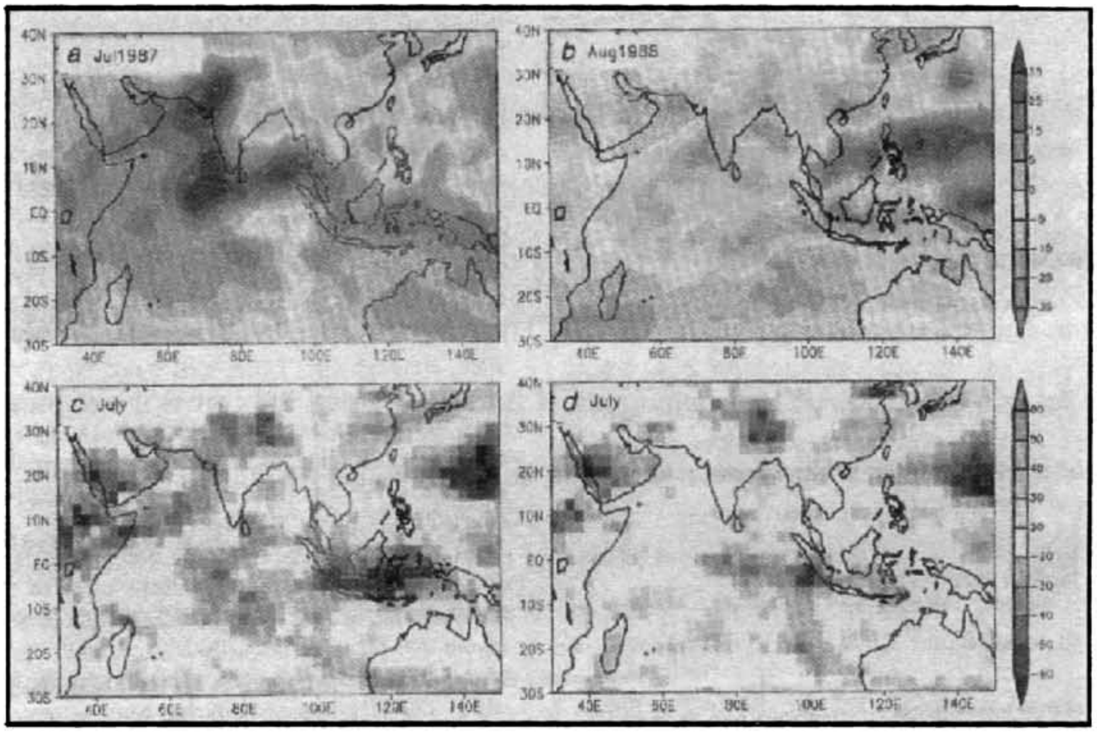


Fig: 5. 3 Anomaly patterns of OLR for (a) July 1987 and (b) August 1988. Correlation for July of OLR with the average OLR over (c) central Pacific (160^0 E - 150^0W, 5^0S – 5^0N) and (d) east Pacific (150 – 90^0W, 5^0S – 5^0N). after Gadgil et al 2003

The impact of a typical El Nino (La Nina) is the suppression (enhancement) of deep convection over the entire equatorial Indian Ocean and Indian region (Rasmusson and Carpenter, 1982; Hastenrath et al 1993; Reason et al 2000; Xie et al, 2002) and Gadgil et al (2003) used OLR anomaly over central Pacific (CPOLR) as an index of ENSO, rather than OLR anomaly over that of east Pacific, since the deep convection over central pacific is better correlated with ISMR anomalies over the Indian region.

It has been observed that El Nino (La Nina) events are associated with easterly (westerly) anomalies of zonal wind along the equatorial Indian Ocean. (Reason et al 2000, Ropelwski and Halpert, 1989). Hence it is expected that EQWIN be negatively correlated with CPOLR, but during the large fraction of the seasons it is not the same. Hence it is suggested that a significant part of the variability of the Indian Ocean coupled system is independent of ENSO (Saji et al 1999; Webster et al 1999).

5.4 EQUINOO, ENSO and ISMR (1945-1980)

The impact of EQUINOO, ENSO events on ISMR is studied for the period 1945-1980 using the same criteria adopted for the period 1979-2003 by Gadgil et al (2003). Since there was no satellite data for OLR anomalies before 1979, the cloud anomaly in the central Pacific Ocean is used to indicate the impact of ENSO. Hence cloud anomaly in the central Pacific is used as an indicator of EN°SO event, while the zonal wind anomaly in the equatorial Indian Ocean (60°-90°E: 2.5°N – 2.5°S) as an indicator of EQUINOO.

5.5 Equatorial Zonal Wind anomalies

The time evolution of zonal wind anomaly along the equator averaged over 2.5°N – 2.5°S for two every good (1956, 1961) and two every bad monsoon (1972, 1979) are presented (Fig. 5.4). In 1956 negative zonal wind anomalies appeared in the eastern Indian Ocean in January, but such signal occurred in the West Indian Ocean by June only. However in 1961 continuous negative zonal wind anomaly appeared in the West Indian Ocean by May, which becomes much intense later, resulting in strong positive IOD. The positive zonal anomaly which in unfavourable for good ISMR was evident for May onwards in 1972 and early part June in 1979 (though weak) which persisted during the entire monsoon. This gives an indication that the phase of the EQUINOO can be inferred evident before the onset of monsoon except in 1956. The analysis of Gadgil et al (2003) for the years 2002 and 2003 also showed the EQUINOO signal by April - May. Between 1965 –1968 (Fig 5.5) the signal of EQUINOO during monsoon was evident from March-May with same minor variations. Between 1973 – 1976 (Fig. 5.5) the EQUINOO during monsoon was weak though there is some indications of it appeared before monsoon. Probably only in those years EQUINOO is relatively strong, early signals of its favourable (unfavourable) phase on ISMR may be clearly visible before the onset of monsoon.

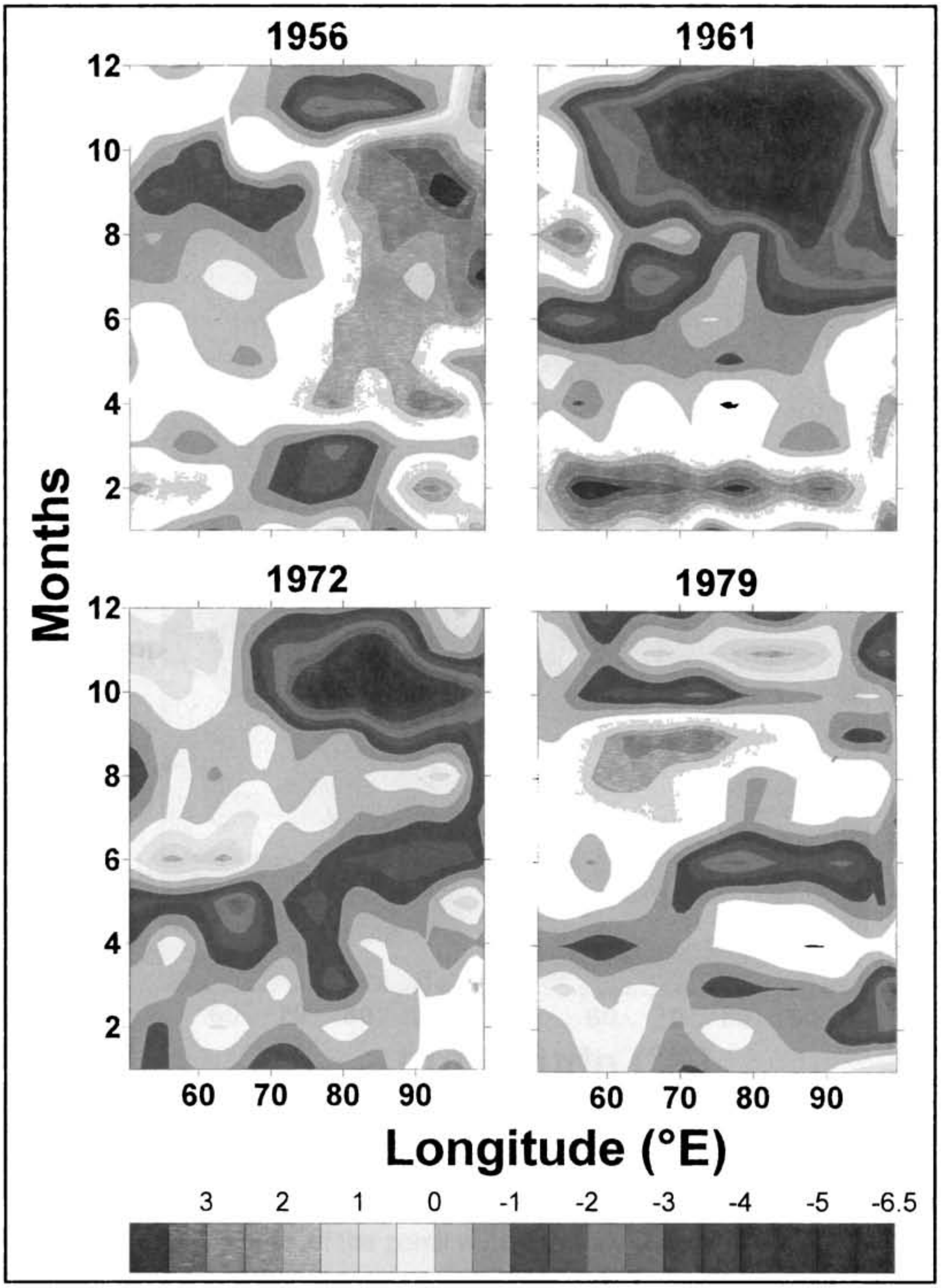


Fig: 5.4. Time evolution of the zonal wind anomaly along the equator (averaged over 2.5°N to 2.5°S) from 1956,1961, 1972, 1979.

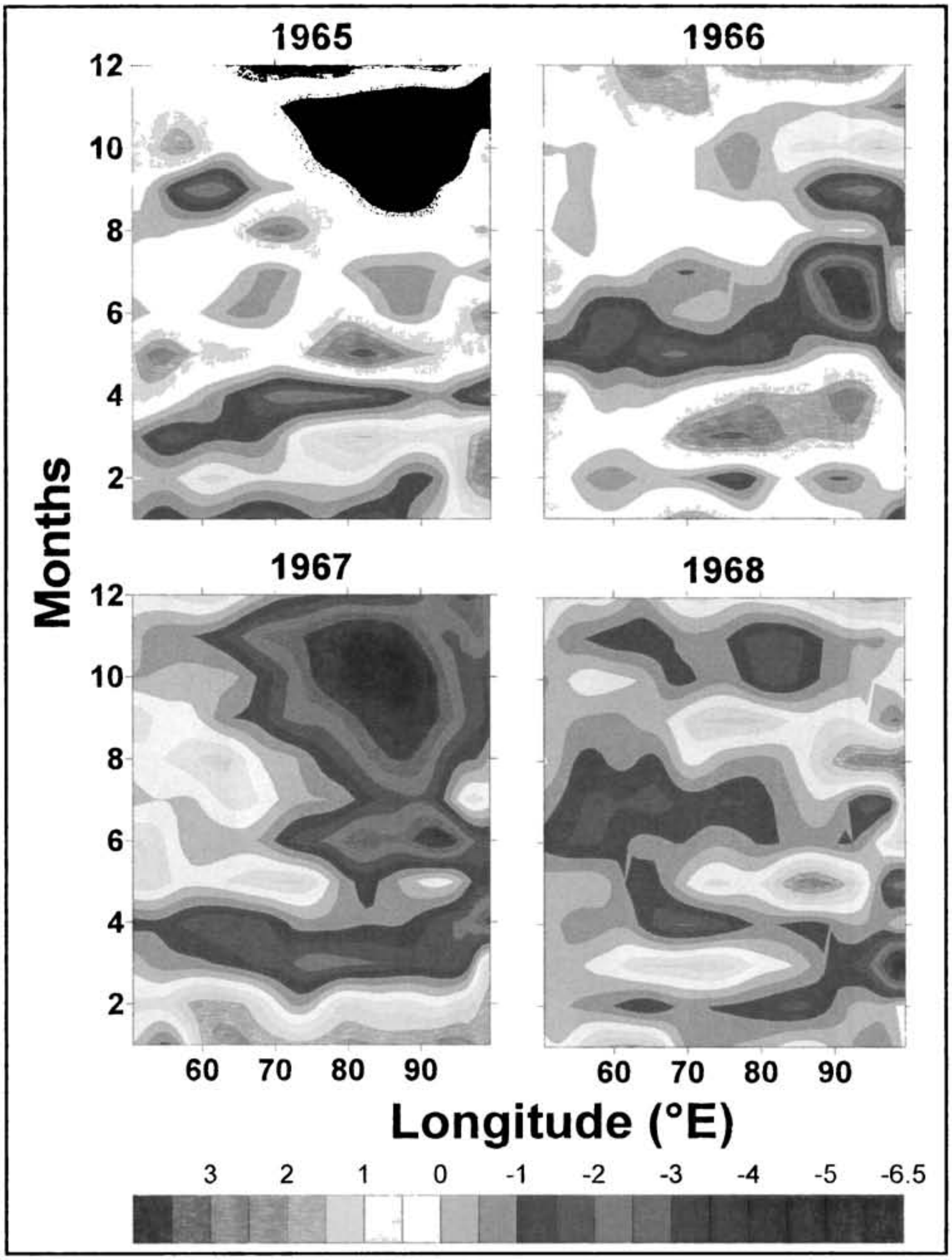


Fig 5.5 Time evolution of the zonal wind anomaly along the equator (averaged over 2.5ºN to 2.5ºS) from 1965, 1966, 1967, 1968.

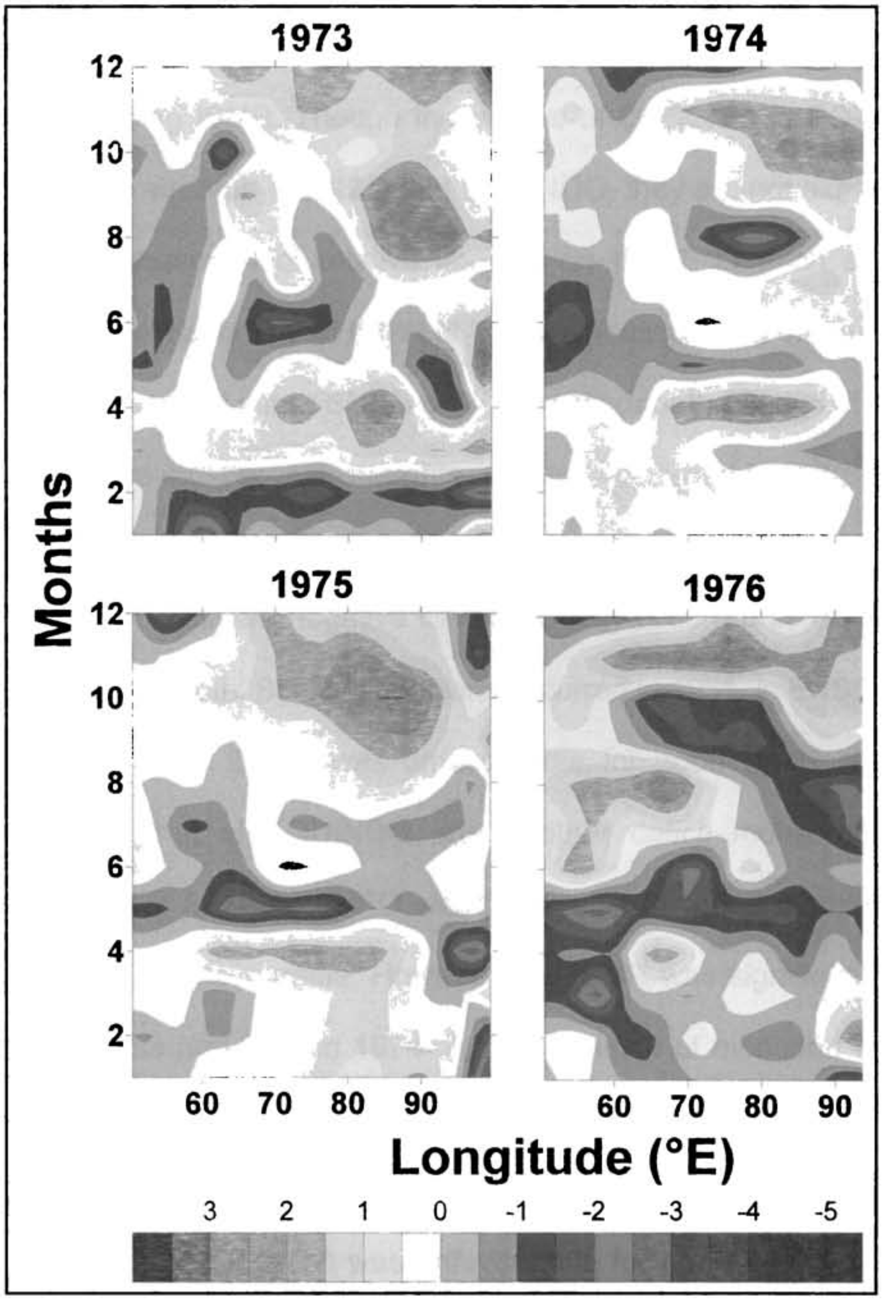


Fig: 5. 6. Time evolution of the zonal wind anomaly along the equator (averaged over 2.5°N to 2.5°S) from 1973, 1974, 1975, 1976.

The ISMR anomaly, EQWIN and cental Pacific Ocean cloud anomaly (CPOCA) for the entire summer monsoon seasons with large ISMR anomaly and a typical El Nino year with normal monsoon of 1974 are shown in (Fig. 5.7) in the order of increasing ISMR. Though the indices of CPOCA and EQWIN give an indication of the strength of ENSO and EQUINOO, they are not expected to give quantitative assessment of relative impacts on ISMR. In each season large negative (positive) anomaly of ISMR is in general associated with unfavourable (favourable) phases of either ENSO or EQUINOO or both (Gadgil et al 2003).

One of the most severe droughts over India occurred in 1972. In that year, the ENSO phase was unfavourable for good monsoon but EQUINOO was also unfavourable. The severe negative ISMR anomaly occurred in 1972 due to the combined effect of both. Similar conditions occurred in 1951, in ENSO EQUINOO and ISMR. In 1965, ENSO was unfavourable for good ISMR and its effect combined with that of EQUINOO and another deficit monsoon over India occured. However in 1966, ENSO was favourable for good monsoon, EQWIN was certainly not unfavourable. However, the ISMR was significantly negative. Similar conditions prevailed in 1974 also. The failure of monsoon in both 1966 and 1974 is difficult to explain with the favourable (unfavourable) phases of ENSO and EQUINOO.

In 1968 the EQUINOO was unfavourable for good monsoon but ENSO was favourable. However the ISMR anomaly was negative indicating bad monsoon. In 1979 the EQUINOO phase was unfavourable and ENSO phase was also marginally unfavourable, but one of the worst summer monsoon occurred in

1979. The index of ENSO and EQUINOO was only marginally unfavourable and hence the severe drought of 1979 also is difficult to explain with the present index for EQUINOO and ENSO. One reason for this reason could be the quantitative impact of ENSO and EQUINOO on ISMR is not very clear.

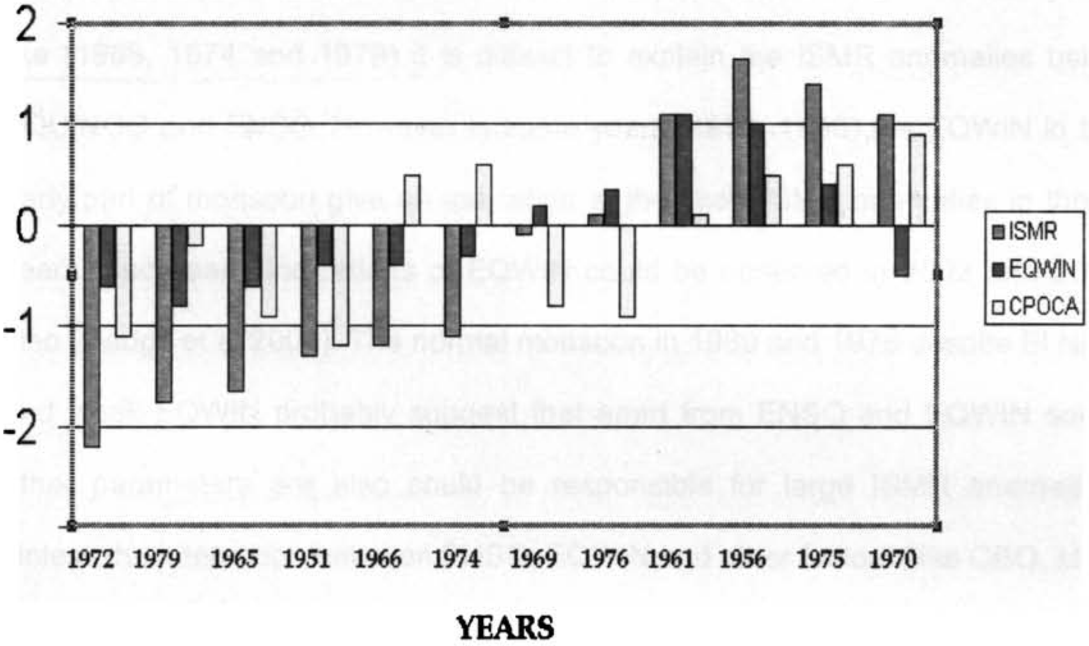


Fig.5.7 ISMR anomaly, EQWIN and CPOCA anomaly (each normalised by the standard deviation) during June – September in the order of increasing ISMR for years with large deficit in ISMR, 1969,1976 excess in ISMR

A very good monsoon occurred in 1961. A very strong positive IOD event also occurred towards the end of the year resulting in heavy rainfall over equatorial eastern Africa and drought over Indonesian region. The influence of ENSO was marginal in that year but EQUINOO favoured a good monsoon. A strong favourable EQUINOO formed in the good monsoon in 1956 also. Despite an unfavourable EQUINOO, the favourable ENSO conditions favoured a good monsoon in 1970 and also in 1974. However, normal monsoon occurred in 1969 and 1976 despite unfavourable ENSO events though the index of EQUINOO only slightly favoured good monsoon rainfall.

The favourable (unfavourable) phases of EQUINOO and ENSO or both appears to explain the ISMR anomaly in many years. But in some other years like (1966, 1974 and 1979) it is difficult to explain the ISMR anomalies using EQUINOO and ENSO. However in some years (1961, 1956) the EQWIN in the early part of monsoon give an indication of the likely ISMR anomalies in those years. Such early indications of EQWIN could be observed in 2002 and 2003 also (Gadgil et al 2003). The normal monsoon in 1969 and 1976 despite El Nino and weak EQWIN probably suggest that apart from ENSO and EQWIN some other parameters are also could be responsible for large ISMR anomalies. Unless the interaction between ENSO, EQWIN and other factors (like QBO, MJO and high frequency variabilities) is fully understood quantitatively, a realistic prediction of ISMR anomalies will continue to be a challenging task.

CHAPTER VI

SUMMARY AND CONCLUSIONS

The thesis entitled " An Investigation on Air-Sea Interaction in the Indian Ocean during Contrasting Monsoons" is primarily based on the analysis of monthly anomaly fields of surface marine meteorological data and computed surface fluxes over the Indian ocean. The anomalies of the surface marine data and heat fluxes are studied in relation to typical very good and very bad monsoon years between 1945 and 1993. Based ISMR anomalies the good monsoon years chosen are 1956, 1961, 1968, 1973, 1975, 1983 and 1988. The bad monsoon years are 1951, 1965, 1966, 1972, 1979, 1982 and 1987. The data set used is the surface marine data (Da Silva et al., 1994), which is based on COADS observations.

The analysis of anomalies of surface meteorological field such as SST, wind speed and direction, sea level pressure and cloud cover for contrasting monsoons yielded interesting results. The SST anomalies during pre-monsoon or summer monsoon do not indicate the nature of monsoon in that year, as there are many good monsoon years with negative SST anomalies and positive SST anomalies over the Indian Ocean and Vice-versa.

- The negative SST anomalies over Indian Ocean are generally associated with La Nina.
- Positive SST anomalies with good monsoon rain fall (eg. 1983 and 1988) follow a severe El Nino and bad monsoon in the previous year

The largest scale Indian Ocean SST anomalies are closely linked with ENSO events. Hence regions of warm SST may not be regions of negative OLR in the Indian Ocean (Webster et al, 1998). This could be one of the major reasons for any one to one direct relationship between Indian summer monsoon rainfall (ISMR) anomalies and SST in the Indian Ocean.

The anomalies of the surface wind suggests that

1) In general a strengthening of south westerlies (of over 3 to 5 m/s) in the Arabian Sea occurs during good monsoon years and weakening of SW take place during bad monsoon years. However 1983 remains as an exceptional case during early part of the monsoon.
2) Cross equatorial flow in the western Indian Ocean is weak in bad monsoon years compared to good monsoon years.
3) Some bad monsoon years are distinguished with the strengthening of south westerlies in the eastern part of the Bay of Bengal, eg. 1965.
4) The wind anomalies are strongest in the eastern equatorial Indian Ocean especially associated with EQUINOO or IOD.

SLP anomaly of monsoon months suggests that bad monsoon years have positive anomalies over the Indian seas and some part of South Indian Ocean. The positive anomalies indicate that SLP is high for bad monsoon months indicating high-pressure zones. The high pressure zone suppresses the air-sea interaction and convective activity reducing the moisture convergence in the boundary level. The pressure anomalies are positive in major part of northern Indian Ocean during bad monsoon years.

However, during good monsoon years, the pressure anomalies are negative indicating a fall in SLP during pre-monsoon and monsoon months. The condition is favourable for the updraft and moisture convergence in the atmosphere leading to an increased monsoon activity. Thus SLP anomalies can be used as a possible indicator to assess the performance of monsoon activity. This study also reveals that, in the pre-monsoon season and monsoon season, during excessive rain fall years, the cloud cover is more in the equatorial and north Indian Ocean compared to same period of drought years.

Of the four anomalies of surface meteorological parameters studied, the anomalies of wind were strongest in the eastern equatorial Indian Ocean compared to those of SST, SLP and cloud cover. In the wind anomaly, the zonal component was predominant. It is found to the strongest between 0^{o} - 10^{o} S and $85^{o} - 95^{o}$ E. The study indicates that the IOD events can be better inferred from the index of zonal wind anomaly compared to SST, SLP or cloud anomalies. The filtered (periodicities up to one year removed) zonal wind anomaly during 1945 – 1993 shows that the oscillations were strongest in the 1950's and 1960's and the amplitude decreased till the early nineties. FFT anomalies showed a periodicity of 2.5 years indicating QBO. Possibly IOD events are manifestations of QBO.

Having studied the anomalies of surface meteorological parameters the analysis of anomalies of net heat flux, short wave flux, latent heat and evaporation were carried out. This study showed the occurrence of large positive anomalies of latent heat flux and evaporation in the western tropical Indian Ocean during good monsoon years and large negative anomalies in bad monsoon years. The moisture flux from that region could be one of the main sources of moisture to the ISMR. Moreover, in the central and eastern equatorial Indian Ocean, the anomalies largely fluctuated in different years. The moisture transport in the equatorial belt associated with wind anomalies may also influence the patterns of rainfall over the Indian region. The tropical Indian seas are found to lose more heat during excess rainfall years compared to deficient rainfall years. This indicates that the air sea interaction prior to a good monsoon activity is more leading to increased exchange of fluxes of momentum, heat and moisture. Further, it can be seen that, prior to an excess rainfall year, tropical Indian seas lose more heat driving the convection activity in the atmosphere.

Recent studies (Gadgil et al., 2003) have shown that all the major ISMR anomalies during 1979 – 2003 can be explained in relation to favourable/unfavourable phase of EQUINOO or ENSO or both. This concept has been extended to the period 1945-1979. Though the phases of EQUINOO and ENSO explain most of the good and bad monsoon years, during the pre-monsoon and monsoon phases of ENSO and EQUINOO, it is difficult to infer the intensity of monsoons of 1966, 1974 and 1979. In 1979 though the phases of EQUINOO and ENSO were unfavourable the signals were too weak. But in many years (1951, 1956, 1961, 1965, 1972, and 1975) there where very clean early indications of EQUINOO/ENSO effect on ISMR anomalies.

Future scope of the study

The interaction of the marine atmosphere with tropical Indian Ocean and its influence on ISMR continue to be an area of active research. There is still considerable uncertainty on the quantitative relationship between these. Though the effect of ENSO events on ISMR is qualitatively known, there is lot of uncertainty in this aspect also. With the recent discovery of IOD and EQUINOO much more advances have been made on the relationship between the Indian Ocean and ISMR. The combined effect of ENSO, EQUINOO, IOD, MJO etc on ISMR needs to be understood quantitatively for modelling and forecasting of ISMR well in advance. With the availability of near real-time information on several factors affecting ISMR with more accurate and real-time coverage over oceans employing satellites, buoys and Argo floats and with accurate analysis techniques, the relationship between the features in the Indian Ocean, ENSO and ISMR will soon be well understood.

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