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Simulation of SST fluctuations and circulation in the Equatorial Indian Ocean

D.W.Ganer, A.A.Deo and P.S.Salvekar
Indian Institute of Tropical Meteorology, Dr.Homi Bhabha Road, Pasha, Pune – 411 008
E-mail : tsd@tropmet.res.in

ABSTRACT
Inter-annual variability in the equatorial Indian Ocean sometimes exhibits large Sea Surface Temperature (SST) fluctuations. The SST anomaly gradient between eastern and western equatorial Indian Ocean during summer season (June-October) becomes greater than 1°C. This east-west SST anomaly gradient is independent of ENSO and may be a natural oscillation of the Indian Ocean. Such fluctuations are observed in many years during the decade 1990-2000. In the present study a 2½ layer thermodynamic ocean model has been used to simulate inter-annual variability in the circulation and SST anomaly (SSTA) fluctuations in the Indian Ocean. The model simulated SST over the Tropical Indian Ocean is in close agreement with observed SST. The warm SSTA in the western equatorial Indian Ocean and cold SSTA in the eastern equatorial Indian Ocean during the years 1994 and 1997 and, opposite SSTA during the years 1992 and 1996, are well simulated by the model.

INTRODUCTION

The 40-year NCEP/NCAR reanalyses products over the tropical Indian and Pacific Oceans shows that the observed east-west SSTA gradient over the Indian Ocean is a passive response to the Asian Summer Monsoon (ASM) and ENSO. The unusually cold anomalies of surface and subsurface oceanic temperatures in XBT sections off Java reported by Meyers 1996, made researchers to find identical situations which have occurred in the past. The warm SST anomaly in the western equatorial Indian Ocean and cold SST anomaly in the eastern equatorial Indian Ocean during the years 1994, 1997 and vice versa during the years 1992, 1996 are reported as positive and negative Indian Ocean Dipole events (Saji & Yamagata 2003).. A few years later, intrigued by the observation of Reppin et al.(1999), that the Yoshida -Wyrtki jet was absent in 1994, Vinaychandran et al., (1999), investigated this event using OGCM. They showed that the anomalous Yoshida -Wyrtki jet and the accompanying SST anomalies were brought out by wind forced ocean dynamics. Behera, Salvekar & Yamagata (1999) went up a step further to suggest that this event evidenced air-sea interaction inherent to the Indian Ocean. In 1997, yet another Indian Ocean event occurred, which was accompanied by an El -Nino.

A different approach towards understanding the Indian Ocean events was taken up by Saji & Yamagata (2003). In the 41 years of observations from 1958 to 1998, they noted several of these events in all, of which equatorial zonal wind anomalies developed simultaneously with a zonally oriented dipole SST anomaly; interestingly enough the phase of these anomalies were locked always with the same phase of the seasonal cycle. Using composites of significant events, they revealed that the dipole SST anomaly and the equatorial zonal wind anomalies were strongly and interactively coupled. In this study, we will discuss about the model simulation of the Indian Ocean Dipole events and the effect of these events on the Indian Ocean circulation.

THE MODEL AND DATA

The model used in this study is a 2½ layer thermodynamic ocean model (McCreary, Kundu & Molinari. 1993) hereafter referred as MKM. In short, the model has two active layers overlaying a deep motionless layer of infinite depth. The upper two active layers interact with each other through entrainment and detrainment while conserving mass and heat of the total system. In simpler form the surface uppermost layer is a single layer with a thickness of h_1 and temperature T_1 which separates into two sub-layers i.e. well-mixed upper turbulent layer of thickness h_m and temperature T_m and a non-turbulent fossil layer of thickness h_f and temperature T_f (Fig. 1).

The uppermost sub-layer of the surface layer that can be termed as upper mixed layer entrains or detrains

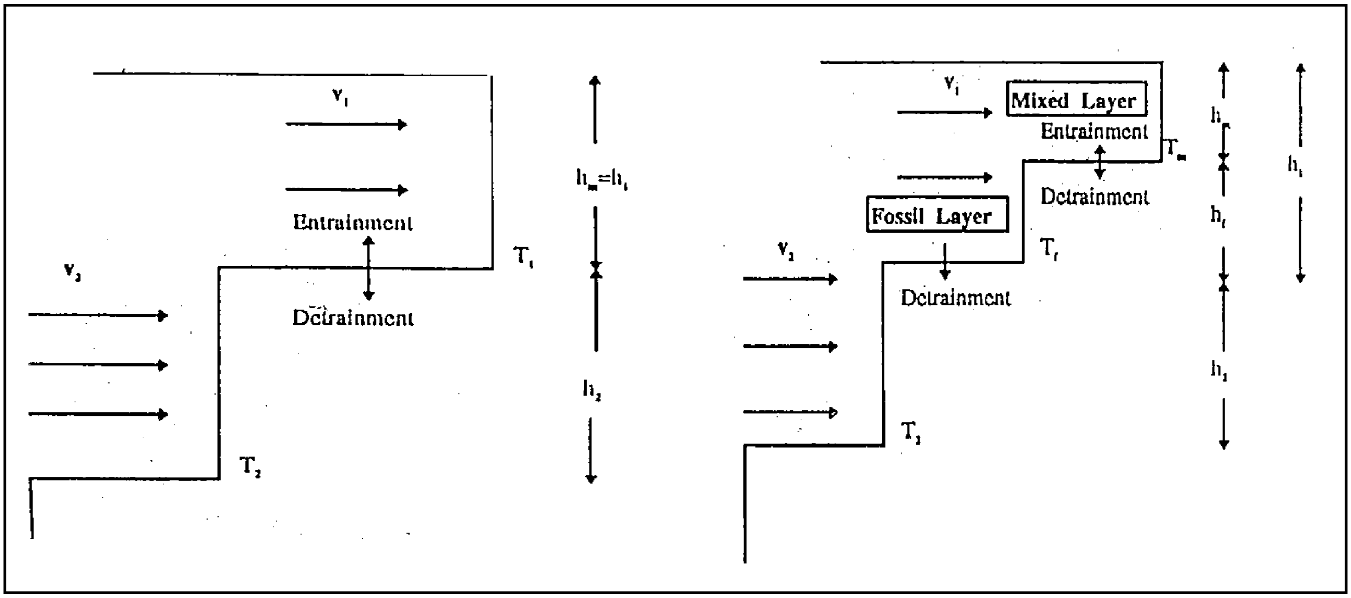

Figure 1. Model structure in(a) simple and (b) complex form

water in a process in which the mixing is maintained by turbulence generated by both wind stirring and cooling at the surface. The non-turbulent fossil layer i.e. the lower sub-layer of the surface layer being formed by the detrainment of water from the upper mixed layer is kept isolated from the surface forcings. There is also provision for detrainment of water from the upper surface layer to the model second layer to conserve the mass of the layer, as entrainment through the base of the surface layer takes mass from the second layer. The second layer is driven by mass and heat fluxes which are derived from the first layer. In this study the temperature of the uppermost turbulent sub-layer is considered as representative of SST.

The wind stress used as surface forcing, is derived from daily NCEP surface winds with drag coefficient $C_D = 1.25 \times 10^{-3}$ and air density $\rho = 1.2$ Kg m^{-3}. The surface heat flux used as a thermal forcing in the model is derived from the daily net solar radiation (incoming - outgoing), air temperature, specific humidity and scalar wind magnitudes. These fields were derived from the daily NCEP data set and are linearly interpolated to the model grid. Complete description of model equations, numerical methods, boundary conditions etc., is contained in McCreary, Kundu & Molinari (1993).

RESULTS AND DISCUSSION

Initially the model is integrated for ten years to reach the steady state with mean surface winds and heat fluxes those are obtained by averaging the ten years daily NCEP surface winds and heat fluxes over the period 1992 to 2001. Since the numerical solution reached a quasi-equilibrium state after this integration, the model solutions by the end of this integration are considered as steady state solutions for the inter-annual runs. The model equations are integrated with inter-annually varying daily NCEP surface winds and heat fluxes from the year 1992 to 2001.

The model simulates inter - annual variability over the Indian Ocean region in particularly over equatorial Indian Ocean, using daily forcings. The difference between model simulated SSTs and Reynolds SSTs is between 0.5 to 1 °C over the whole model domain. The model SST anomalies over equatorial region are in close agreement with the observed anomalies. The model circulation and SST anomalies during the positive dipole years 1994, 1997 and negative dipole years 1992, 1996 are studied.

Figure 2 shows the model simulated and Reynolds SST anomalies over the Tropical Indian Ocean region for the month of September for the positive dipole years 1994 and 1997. The warm SST anomaly in the western equatorial Indian Ocean and cold SST anomaly in the eastern equatorial Indian ocean during the years 1994, 1997 observed in the Reynolds SST anomalies are very well simulated by the model. These SST anomalies persist in the equatorial Indian Ocean from June to November. The model SST anomalies attain peak in the month of September and suddenly disappear in the next month terminating the dipole mode. The cause for termination of dipole event in absence of El-Nino event, is westerly wind bursts drive a Kelvin wave traveling eastwards. This takes one month to travel from west to east in the equatorial

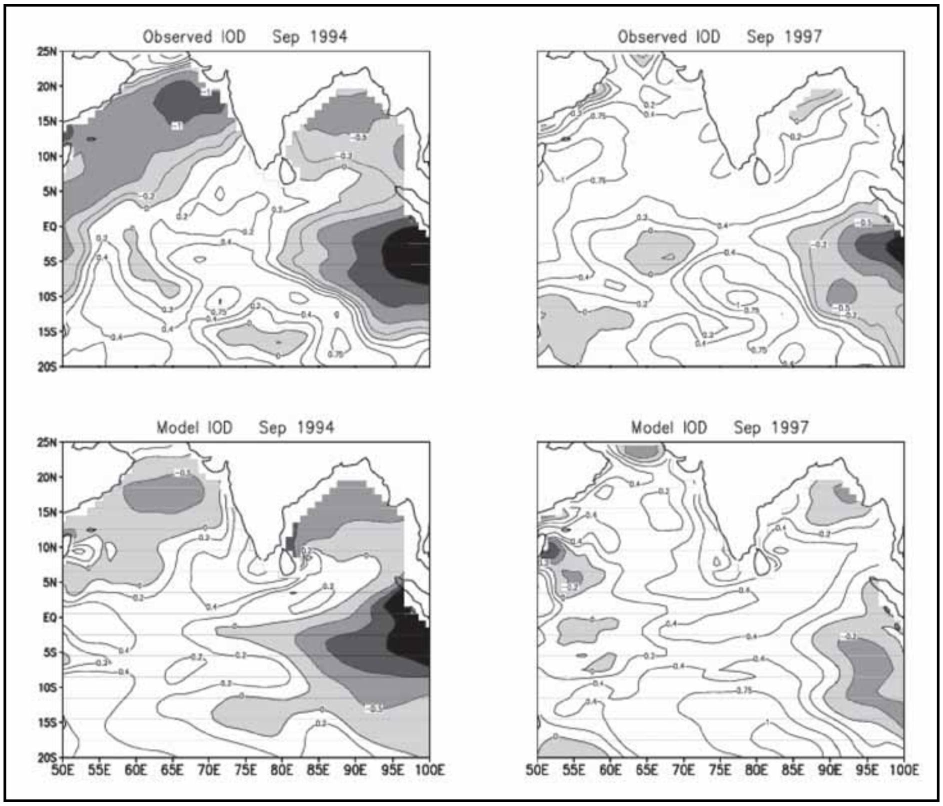

Figure 2. Observed and Model simulated positive IOD events during 1994 and 1997. Negative values are shaded.

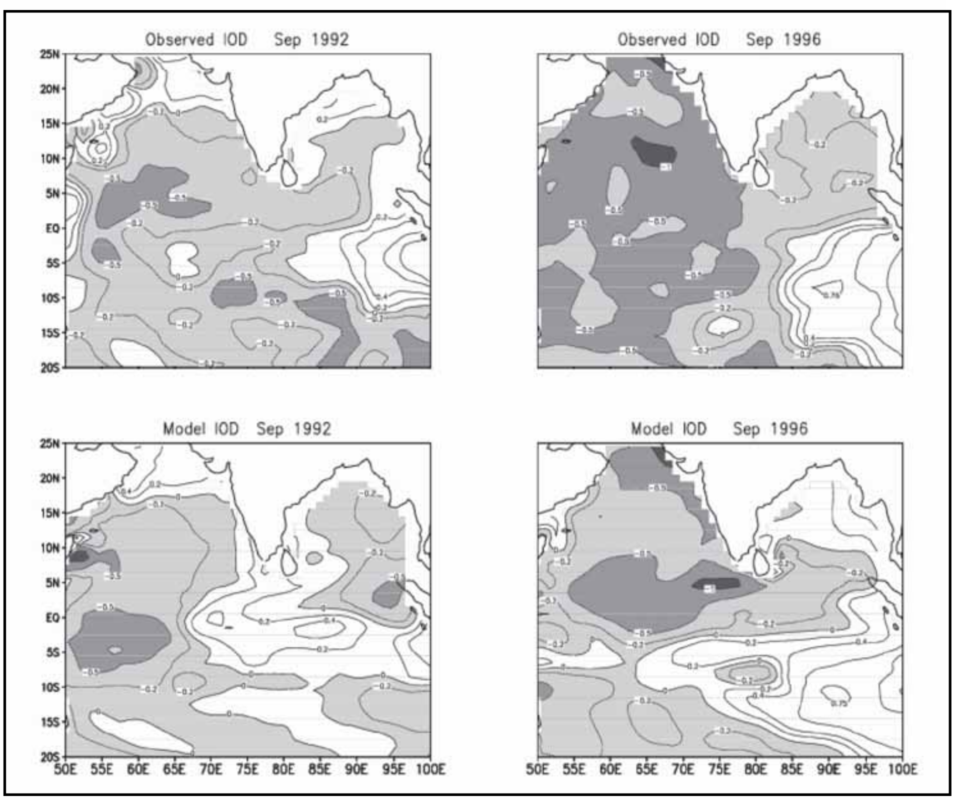

Figure 3. Observed and Model simulated negative IOD events during 1992 and 1996. Negative values are shaded.

Indian Ocean. After reaching the east coast this wave ultimately deepen the thermocline in the eastern Indian Ocean and warms the sea surface there, bringing back the equatorial Indian Ocean to normal state (Suryachandra Rao ,& Yamagata 2004). It can also be seen that the 1994 dipole event was stronger than the 1997 dipole event. The model simulated cold SST anomalies during the year 1997 are less in magnitude than the observed SST anomalies. The cold SST anomaly in the western equatorial Indian Ocean and warm SST anomaly in the eastern equatorial Indian ocean during the years 1992, 1996 (negative dipole years) observed in the Reynolds SST anomalies are also well simulated by the model (Fig. 3).

The SST dipole is strongly coupled to equatorial wind variability over the eastern equatorial Indian Ocean (Saji & Yamagata 1999). The composites of wind stress anomalies over this region (Fig. 4) show that during the positive dipole years the south easterly anomalies are stronger resulting in strong zonal anomalies along the equator. During the negative dipole years the wind anomalies are westerly.

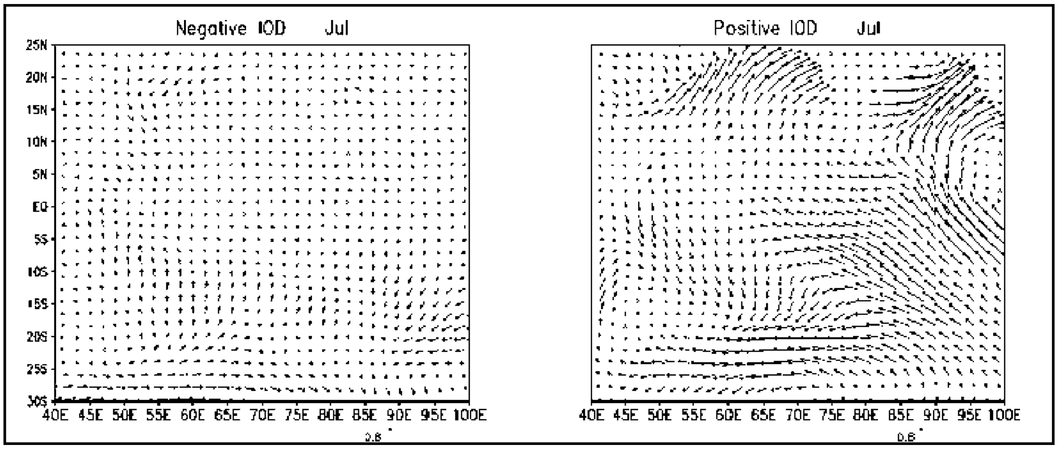

Figure 4. Composite of Wind stress anomaly (Nm^{-2}) during negative and positive IOD events

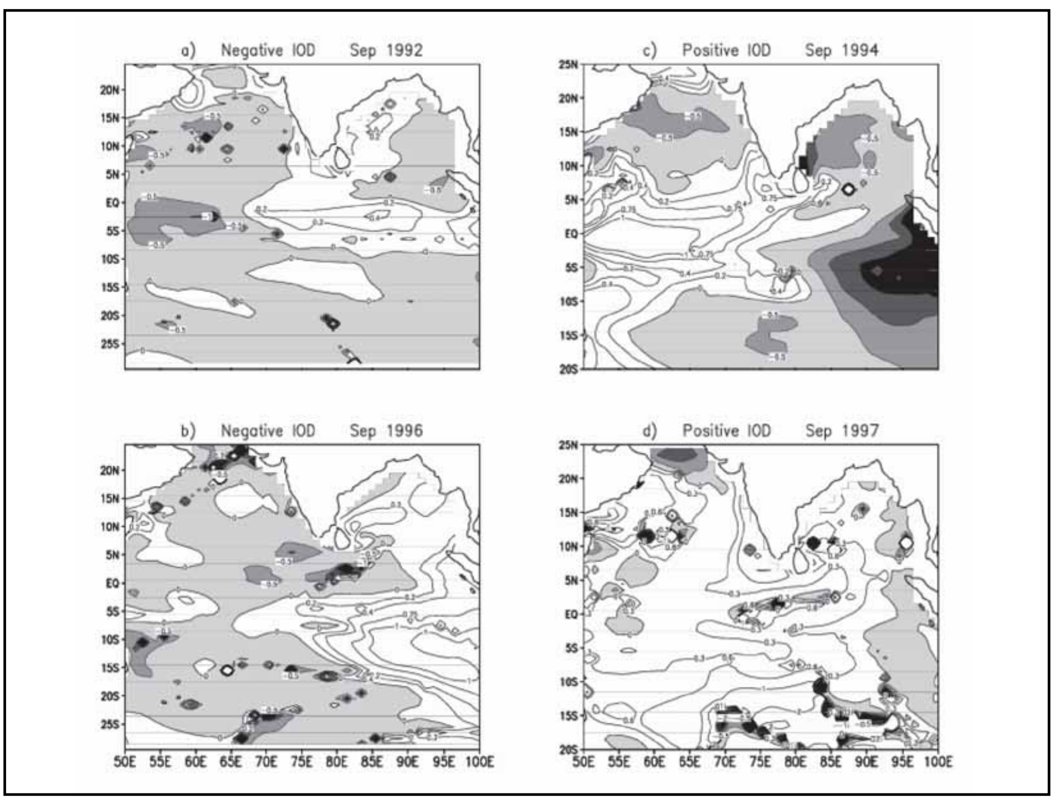

Figure 5. Model simulated sub-surface negative and positive IOD events. Negative values are shaded.

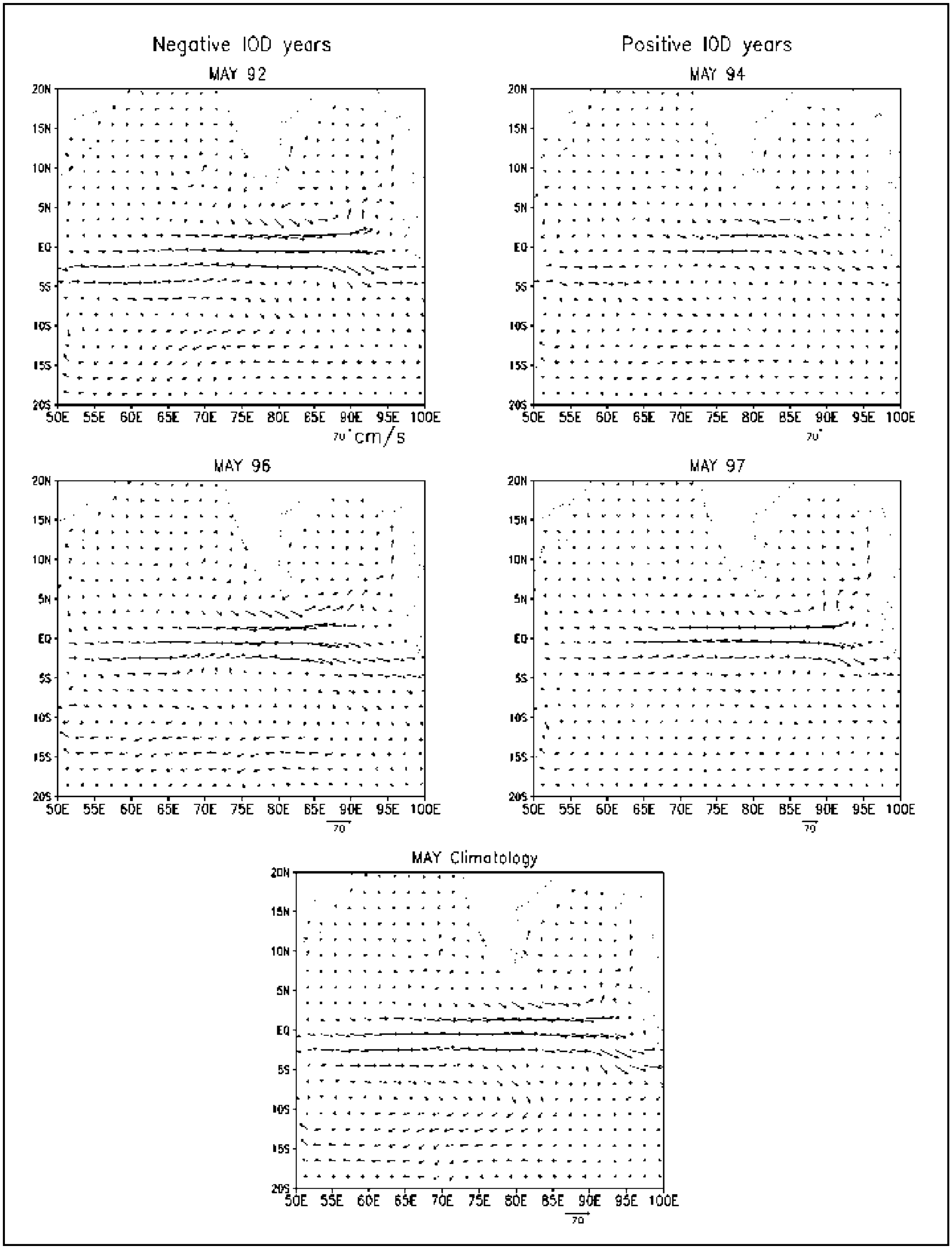

Figure 6. Model simulated Equatorial jet for May

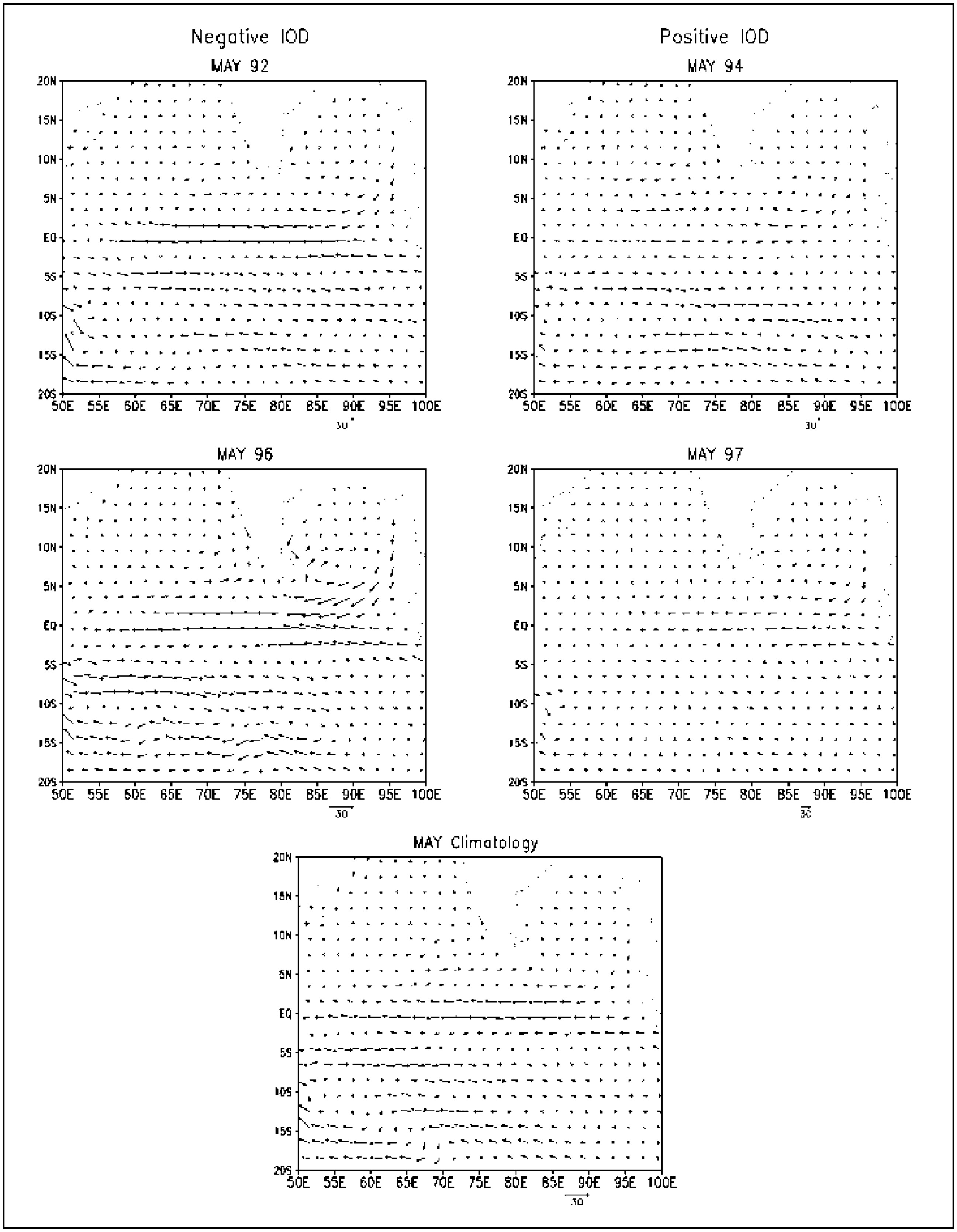

Figure 7. Model simulated Lowr layer currents (cm/s) for May

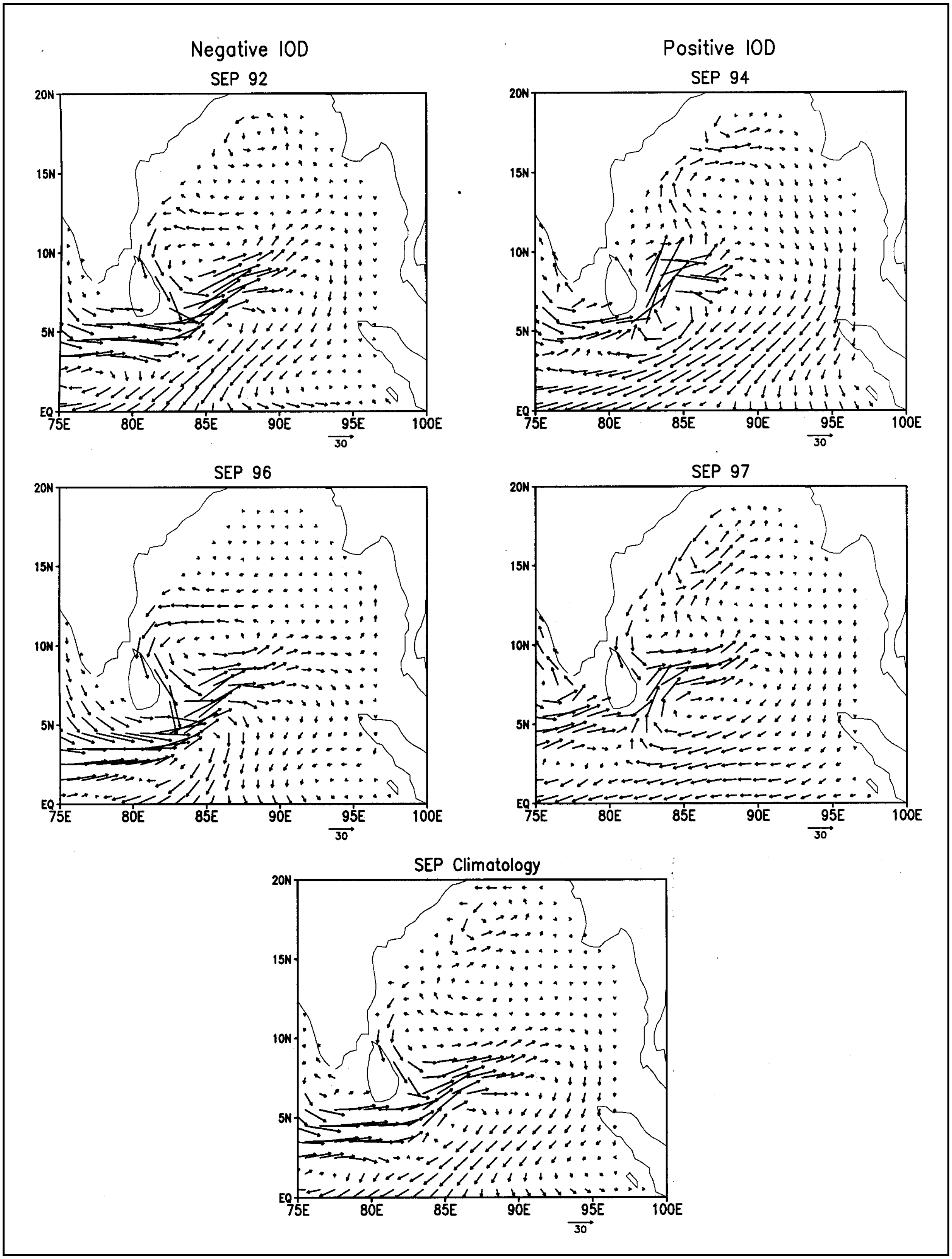

Figure 8. Model simulated currents (cm/s) in Bay of Bengal

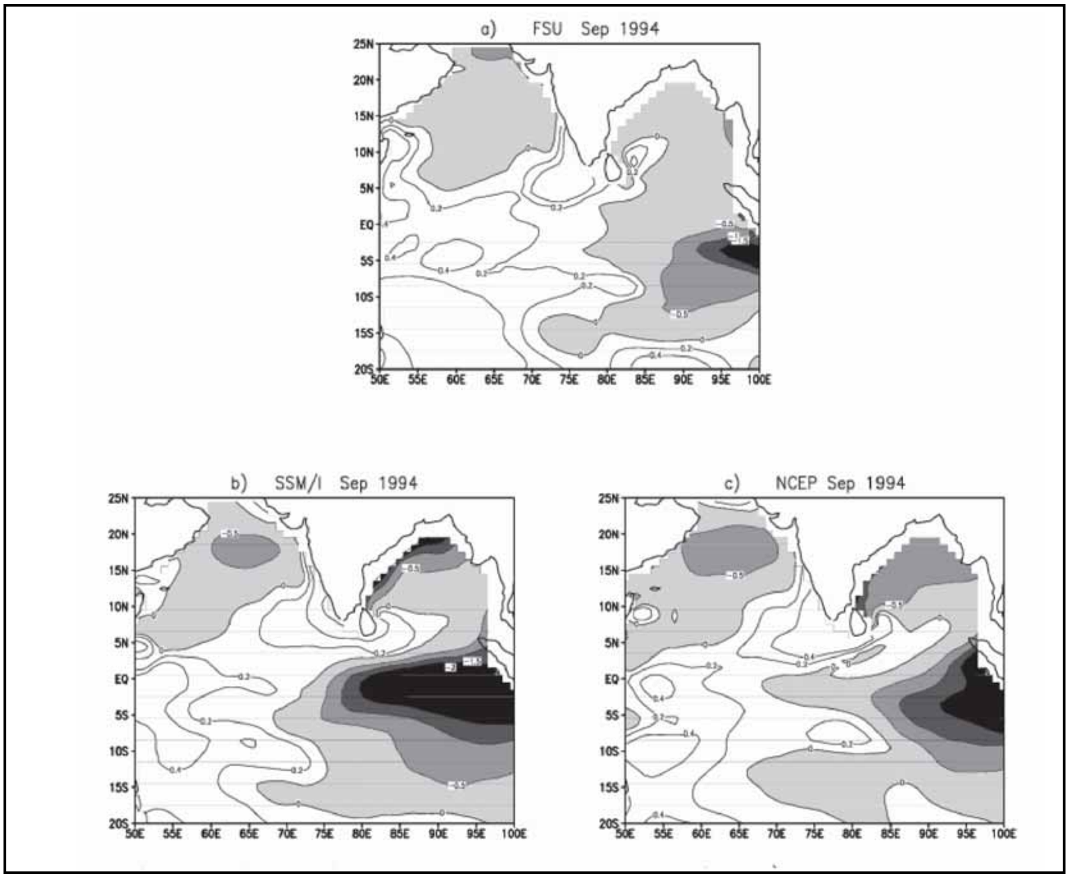

Figure 9. Simulation of 1994 IOD event with a) FSU monthly, b) SSM/I 5 days and c) NCEP daily winds. Negative values are shaded.

The fluctuations in the temperature are also noticed in the sub-layer of the model. The sub surface temperature also exhibits dipole structure (Fig. 5). The depth of the model subsurface is approximately 100 – 120 meters which is in agreement to some model simulation studies and subsurface oceanic temperatures in XBT sections (Meyers 1996). The structure of the subsurface dipole is similar to the surface dipole. There is little difference in model simulated sub surface dipole and surface dipole structure during 1997. The region of cold temperature anomalies is small as compared to the surface one in the year 1997. Also warm temperature anomalies are extended from west to east up to the region of cold temperature anomalies.

The positive dipole years 1994 and 1997 are strongly associated with wind variability. The model circulation shows that the equatorial jet is absent in the year 1994 and weak in the year 1997 (Fig 6), which is also reported by Reppin et al., (1999) and Vinaychandran et al., (1999) whereas, during negative dipole years 1992 and 1996, the equatorial jet is present which is similar to the Climatology of the model (Fig 6). The absence of equatorial jet in the positive dipole years may be due to the weak westerly winds in the equatorial region during these years. The model lower layer currents get reversed in direction and the magnitude of the currents reduces to half of the surface currents in both the positive and negative dipole years (Fig 7). The lower layer currents show the absence of strong westward equatorial currents in the year 1994 and weak westward equatorial currents in the year 1997.

The response of the positive dipole years is not only confined to the equatorial region but also extended to the Bay of Bengal region. Fig 8 shows the model currents in the Bay of Bengal for the month of September 1994, 1997, 1992 and 1996. The model currents show that the upwelling coastal Kelvin wave during the positive IOD events in 1994 and 1997 influences the circulation in the Bay of Bengal in the month of September. This was also found by Vinaychandran et al., (1999). This strong propagating Kelvin wave along the perimeter of the Bay sets up southward currents along the eastern boundary of the Bay and reflected Rossby wave sets up northward currents along the western boundary of the Bay forming an anti-cyclonic circulation in the Bay.

During negative IOD events in 1992 and 1996, the circulation in the Bay of Bengal is similar to the climatological one and remains unaffected.

Sensitivity experiments are carried out by using monthly FSU winds and five days SSM/I winds. In both the experiments, the NCEP heat fluxes are used. The model runs are carried out with inter-annually varying forcings and the coefficients used in the model are modified according to the forcings. Figure 9 shows the model simulated dipole event for the month of September 1994 with monthly FSU, 5 days SSM/I and daily NCEP winds. The dipole structure in all the three experiments is well simulated. There is little difference in the spatial structure of cold anomaly. In FSU wind case, the spatial structure of cold anomaly in the eastern equatorial region is not well organized. The region of coldest anomaly is small. The spatial structure of cold anomaly in SSM/I case is extended westward and the magnitude of cold SST anomaly is higher. The spatial structure in daily NCEP case is similar to the observed one. The spatial structure of warm anomaly in all the three cases is more or less similar and is in agreement with observed one. The warm anomaly in the southwest corner of the model domain in the FSU wind case differs from the observed anomaly. The model is able to simulate the dipole event for the year 1994 with different surface winds.

CONCLUSIONS

The fluctuations in the model SST anomalies and Reynolds SST anomalies and the variability in the model currents is studied using intermediate 2½ layer thermodynamic ocean model. The seasonal and annual cycle is well simulated by the model. The warming in the western equatorial Indian Ocean region and cooling in eastern equatorial Indian Ocean region during the years 1994 and 1997 and opposite during the years 1992 and 1996 which is found in the Reynolds SST anomalies is very well simulated by the model using daily forcings. The positive dipole events during 1994 and 1997 are strongly associated with south easterly wind stress anomalies and negative dipole events during 1992 and 1996 are associated with westerly wind stress anomalies in the equatorial region. The model sub-surface temperature also shows both positive and negative dipole structure.

The sensitivity of the model to the inter-annually varying winds produces the interannual variability in the model currents in the equatorial region. The equatorial jet is absent in the positive dipole year 1994 and is weak in the year 1997. The negative dipole years 1992 and 1996 are characterized by the presence of equatorial jet. The model lower layer currents are in the opposite direction. The magnitude of the lower layer currents reduces to half. This may be due to the combined effect of Ekman pumping and radiation of waves from the equatorial Indian ocean. The response of the positive IOD events during 1994 and 1997 is not only confined to the equatorial region but also seen in the Bay of Bengal. The strong propagating Kelvin wave along the perimeter of the Bay sets up southward currents along the eastern boundary of the Bay and reflected Rossby wave sets up northward currents along the western boundary of the Bay forming an anti-cyclonic circulation in the Bay.

During negative IOD events in 1992 and 1996, the circulation in the Bay of Bengal remains unaffected. The sensitivity experiments carried out by using different forcings such as monthly FSU winds and five days SSM/I winds and daily NCEP winds and by keeping NCEP heat fluxes common, helped to test the sensitivity of the model to the different forcings. The spatial structure of the positive IOD event during 1994 is well simulated by the model. The spatial structure of 1994 IOD simulated by using daily NCEP forcings is in close agreement with the observed one. The region of cold anomaly is not well organized in the monthly FSU wind case and in the five days SSM/I case this region is extended westward as compared to the observed event.

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Name : Mr. D.W. Ganer

Educational Qualifications : Junior Scientific Officer, Indian Institute of Tropical Meteorology, Pune-411008

Research Experience : Experience of working in ocean modeling fr the last 11 years. Mainly working on $2^1/_2$ layer thermodynamic ocean model to study the international variation in circualtion and SST over Indian Ocean using different wind forcings. In particular SST anomaly fluctuations in the equatorial Indian Ocean for the positive and negative dipole years are studied. The model simulations of SST anomaly fluctuations are in close agreement with the observations. This thermodynamic ocean model ahs also been sued to simulate the mini warm pool in the east Arabian Sea during pre monsoon. It is observed that many times the onset vortex forms over this region of warm pool. The ARMEX - II was conducted for the period 14 March to 19 June 2003 in connection with this warm pool and XBT surveys during May 2002 - April 2004 to understand the process of monsoon. The variability of cold pool simulated by the model at the southern tip of India during the year 2002 to 2004 is studied. This variability is also seen in the observed SST.

Name : Mrs. Asmita Amrut Deo

Educational Qualifications : M.Sc. (Mathematics), M.Phil.)(Physics)

Research Experience : For the last 9 years I have the experience of working in ocean modeling. I am working on reduced gravity models for the study of intra-seasonal and inter-annual variability in the circulation and SST fields, over north Indian Ocean (35ºE - 115ºE, 30ºS-25ºN). As an application of these models the ocean response to moving cyclones in the Indian Ocean is investigated numerically. Several sensitivity experiments are performed by changing cyclone parameters as well as model parameters, such as intensity, size, speed symmetry, tract direction, model resolution, entrainment, initial conditions. etc., and the variations of ocean parameters in these cases is studied. The model results are compared with observations. Conversly the effect of ocean parameters such as SST, on the cyclone is being considered and the numerical experiments are planned accordingly.

Dr. (Mrs.) P. S. Salvekar has obtained post graduate in Applied Mathematics, and Ph.D. in Atmospheric Physics from Pune University, Pune. She has 30 years Research experiences in Meteorology. Presently, she is working as Scientist –F, Head of Theoretical Studies Division, Indian Institute of Tropical Meteorology, Pune. She has to her credit 86 scientific papers in renowned International and National journals. She has been recipient of two awards: WMO young Scientist Award (International) and Mausam Award (National). She has been also working as visiting Prof. in Pune University since last 20 years. She has examined many Ph. D thesis form Pune University and IITs. Under her guidance six students have been awarded Ph.D. She is specialized in Monsoon Dynamics, Instability studies, Atmospheric Energetics and Modelling in Atmospheric and Oceanic Circulation.