Indian Ocean dipole mode events in a simple mixed layer ocean model

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A precise knowledge of sea surface temperature (SST) is very essential for climate and oceanographic studies. In this paper a simple two dimensional mixed layer ocean model and its numerical code have been developed and used to simulate the SST fields over the north Indian Ocean (20°S-25°N and 35°E-115°E) for a period of 10 years (1992-2001). The model simulated the SST variability reasonably well. The simple model could simulate the observed dipole of 1997 and 1994 very well, especially the eastern cooling. The model study showed that the interannual SST variability in the western equatorial Indian Ocean is not only due to the variability in the surface heat fluxes, but also due to the variability in wind and sea surface height (SSH). The OLR anomaly also shows positive (negative) anomaly over the negative (positive) anomalous SST region. The variability in the latent heat flux is found to be greatly influencing the SST variability in the eastern equatorial Indian Ocean.

Simulation of sea surface temperature (SST) is the most important and yet one of the most difficult task of ocean models. It is important because except for the surface roughness, it is the only oceanic variable that affects the atmosphere directly. SST variability simulates the basic scale air-sea interaction in the Indian Ocean through internal dynamic and thermodynamic processes[1]. Earlier study indicated that except for the upwelling region off Somalia, SSTs in the tropical Indian Ocean tend to be rather high[2]. Atmospheric model studies also indicated that when the absolute temperatures are high, even small variations in temperatures may produce significant climatic impacts. Hence it is important to understand the mechanisms responsible for the interannual variability of the Indian Ocean SSTs. Though most efforts in the past had been focused on understanding Pacific phenomenon, recent studies[1,3,4] show that there are internal dynamic and thermodynamic processes that lead to basin-scale air-sea interaction in the Indian Ocean. Regarding the variability of the SST in the Indian Ocean is concerned there are two schools of thoughts suggesting the evaluation of interannual SST anomaly (SSTA). One advocates the major role of heat flux[5] in the evaluation of interannual SSTA and the other school promotes the dynamical importance of winds[2] in addition to the heat flux. The most important finding from the recent studies on the tropical Indian Ocean variability[4] is existence of an ocean-atmosphere coupled mode, the dipole structure in SSTA.

In order to understand the role of surface heat fluxes in the interannual SST variability, in the present study, we developed a simple ocean mixed layer model and used it to simulate the SST and its variability over the Northern–Indian Ocean (NIO) (Fig. 1). We examine the influence of mixed layer depth (MLD) on SST by taking a constant MLD and the spatially and temporally varying climatological mixed layer depth.

Materials and Methods

The ocean mixed layer temperature change is estimated using the simplified thermodynamic equation:

$$\frac{\partial T_m}{\partial t} = \frac{Q_{net} + Q_{cor}}{C_p \rho h} - \frac{Q_{swh}}{C_p \rho h} - \frac{\omega_e \Delta T}{h} \qquad \ldots (1)$$

where T_m is the mixed layer temperature (MLT, hereafter the model simulated MLT will be called as model SST), h is the depth of ocean mixed layer, Q_{net} is net surface heat flux into the ocean (positive if ocean gains and negative if ocean looses), $\Delta T = 0.5°C$, ω_e is the entrainment rate, ρ and C_p are the reference density and specific heat of the seawater and Q_{cor} is the surface heat flux correction to account the advective and diffusive processes.

The simple ocean model does not explicitly account for the horizontal advection of heat in the ocean, which is found to be an important process[6,7] in determining Indian Ocean SST. The model when forced by the total surface heat flux will not produce a realistic seasonal cycle in regions where advection is important. In order to achieve an accurate seasonal cycle in SST a correction to the observed surface heat fluxes needs to be applied[8]. The magnitude and pattern of Q_{cor} resembles the observed annual mean oceanic heat flux convergence[8]. The correction is done as follows. The ocean model is integrated for one time step (say t_1 to t_2) with the forcing determined from the observed meteorological parameters and climatological SST. From the simulated SST at t_2, a heat flux correction Q_{cor} is calculated as the additional heat that is required to achieve a surface mixed layer temperature equal to the climatological SST at t_2. This process is continued for 10 years after setting SST to climatological value at each time step. The daily Q_{cor} calculated by this process is used to obtain monthly mean climatological Q_{cor}, which is then linearly interpolated to each time step. The total flux through the surface is then reduced by a factor (0.75) of Q_{cor}, which is roughly half as large as net heat flux but of opposite sign. The surface heat flux components are estimated as follows:

Short wave radiative flux Q_I of absorption of solar radiation is calculated as[9]:

$$Q_I = SW(1-\alpha)(1-0.7C) \qquad \ldots (2)$$

where, SW = the net downward flux of solar radiation just above the surface in cloudless conditions in W/m^2, α = albedo at the sea surface (0.06), C = the fraction of sky covered by cloud or cloud cover in percentage. The net upward flux Q_B of long wave radiation from the ocean is calculated by:

$$Q_B = 0.985\,\sigma T_s^4\,(0.39 - 0.05e_a^{1/2})(1-0.6C^2) \quad \ldots (3)$$

where, σ = Stefan-Boltzmann's constant (5.7×10^{-8} W m^{-2} K^{-4}). In the above formula, the first term is a correction factor, 0.985 for departure of the ocean surface from black body behavior, the second term, σT_s^4 is the flux emitted by the blackbody of temperature T_s, third term is a correction factor for back radiation in the absence of clouds, and fourth term is a correction factor for the cloud effect[9].

The sensible and latent heat flux are computed as[10]:

$$Q_s = \rho_a C_p C_H U(T_s - T_a) \qquad \ldots (4)$$

$$Q_E = \rho_a C_E LU(q_s - vq_a) \qquad \ldots (5)$$

where, Q_s = sensible heat flux, Q_E = latent heat flux, ρ_a = air density (1.25 kg/m^3), C_p = specific heat of air at constant pressure (1004 J kg^{-1} K^{-1}, C_H = drag coefficient (1.10×10^{-3}), U = surface wind speed, T_s, T_a = sea surface temperature and surface air temperature respectively, v = 0.8, C_E = dimensionless coefficient, sometime called Dalton number (1.5×10^{-3}), L = latent heat of vaporization (2.5×10^6 J kg^{-1}).
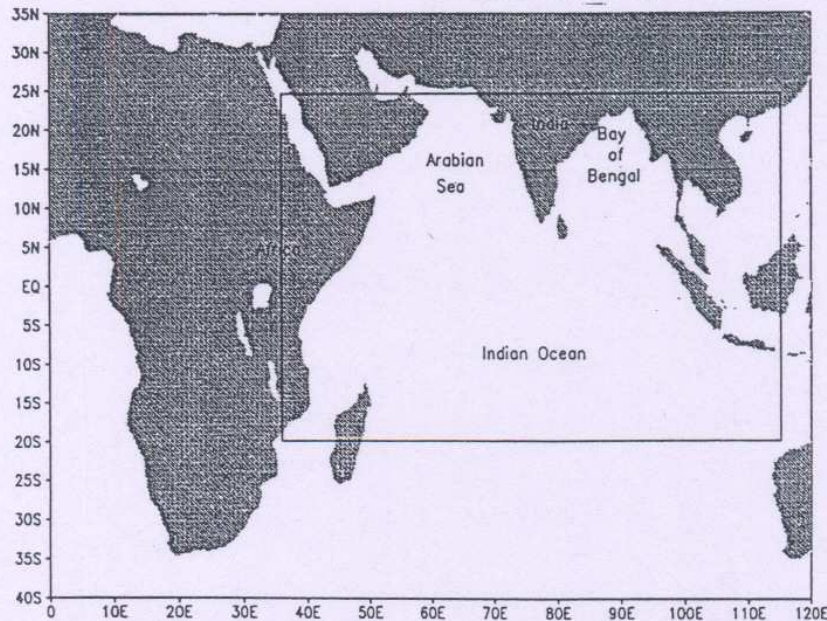

Fig. 1 – Model domain and study area (35° E to 115° E and 20° S to 25 °N)

The sensible heat flux (Q_s) depends on the difference between the sea surface temperature and air temperature, at the standard level, and the heat capacity $\rho_a C_p$ per unit volume of the air[7]. The saturation specific humidity (q_s) at the sea surface and the saturation specific humidity of the air near the sea surface (q_a) are estimated as:

$$q_\gamma = e_\gamma(T_\gamma) \times 0.622 / P_A \qquad \dots (6)$$

where, $\gamma = s$ or a, P_A = sea level pressure (1013 mb), 0.622 = ratio of the densities of the water vapor and dry air, $e_\gamma(T_\gamma) = 10^{[9.4-(2353)/T_\gamma]}$ in mb and T_γ in ° K.

The net heat flux to the ocean through the sea surface Q_{net} can be expressed as:

$$Q_{net} = Q_I - (Q_B + Q_S + Q_E) \qquad \dots (7)$$

Each component of the heat flux can be calculated by the empirical formulas discussed in the previous section.

Very shallow mixed layer was observed over the Bay of Bengal region and hence the short wave radiation received at the sea surface may penetrate beyond the mixed layer. The penetrative solar radiation Q_{swh} below the ocean mixed layer is computed as[11]:

$$Q_{swh} = Q_I \left[R e^{-h/v_1} + (1-R)e^{-h/v_2} \right] \qquad \dots (8)$$

where R is the separation constant, v_1 and v_2 are attenuation length scales. And the entrainment velocity:

$$\omega_e = \frac{2 m_0 u_*^3}{\alpha g h \Delta T} \qquad \dots (9)$$

where, u_* is the frictional velocity, m_0 is an empirical mixing efficiency parameter (0.5), α is expansion coefficient of water (2.5×10^{-4} °C).

In the above equations the MLD is fixed to be 50 m for the entire model region. However if the variable MLD is considered then the thermodynamic equation will be modified to:

$$\frac{\partial T_m}{\partial t} = \frac{Q_{net} + Q_{cor}}{C_p \rho h(x,y,t)} - \frac{Q_{swh}}{C_p \rho h(x,y,t)} - \frac{\omega_e \Delta T}{h(x,y,t)} \qquad \dots (10)$$

where $h(x,y,t)$ is the spatially and temporally varying climatological ocean MLD specified by linearly interpolating the mean monthly MLD values and ΔT is the difference between the mixed layer temperature and the temperature of water entrained in the mixed layer, which is taken as 0.5°C.

The Earth Radiation Budget Experiment (ERBE)[12] mean monthly clear sky radiation, the monthly mean International Satellite Cloud Climatology Project (ISCCP)[13] cloud cover have been used in the study in addition to the National Centers for Environmental Prediction (NCEP) reanalyzed[14] monthly averaged surface wind components and air temperature. All were interpolated to 1°× 1° grid in the model region using bicubic spline technique and are used to estimate the surface heat fluxes which are then used to force the ocean model. The monthly mean sea surface height anomaly (SSHA) was calculated using 10 - day average TOPEX/POSEIDON SSHA produced by the CLS Space Oceanography Division as part of the Environment and Climate EC AGORA (ENV4-CT9560113) and DUACS (ENV4-CT96-0357) projects. The 2.5°×2.5° OLR[15] from NOAA NCEP CPC was interpolated to 1°×1° grid using bicubic spline interpolation techniques in the study.

The surface heat fluxes were computed using bulk aerodynamic relation given above. In the flux computation the simulated SSTs are used rather than using observed SSTs. The forcing data at each model time steps (of integration) was obtained by linearly interpolating the monthly values. The values of the model SST on day 15th of each month are considered as the representative of the monthly values. For simplicity the model calendar year is considered to be of 360 days. The model code has been developed and the model was used to simulate SST for a period of 10 years (1992 to 2001) in the entire model region (Fig.1). The interannual variability in the observed and simulated SST is studied for the integration period 1992-2001. However the discussion is mainly concentrated only on the dipole mode events during the period of study.

Results and Discussion

The model was initialized with the January 1992 Reynolds SST (representing 15th January 1992) and integrated from January 1992 to December 2001 with the interannually varying surface heat fluxes. Monthly anomalies are departures from the corresponding 10 year averaged monthly mean values. It is important to note that heat flux is the only surface forcing applied to the model and it is assumed that the water entrained at the mixed layer base has temperature 0.5 °C less than MLT.

Longitude -time plot (Hovmoller diagram) of Reynolds SSTA for the year 1992-2001 along the equator is shown in Fig 2B and Fig 2A,C shows the
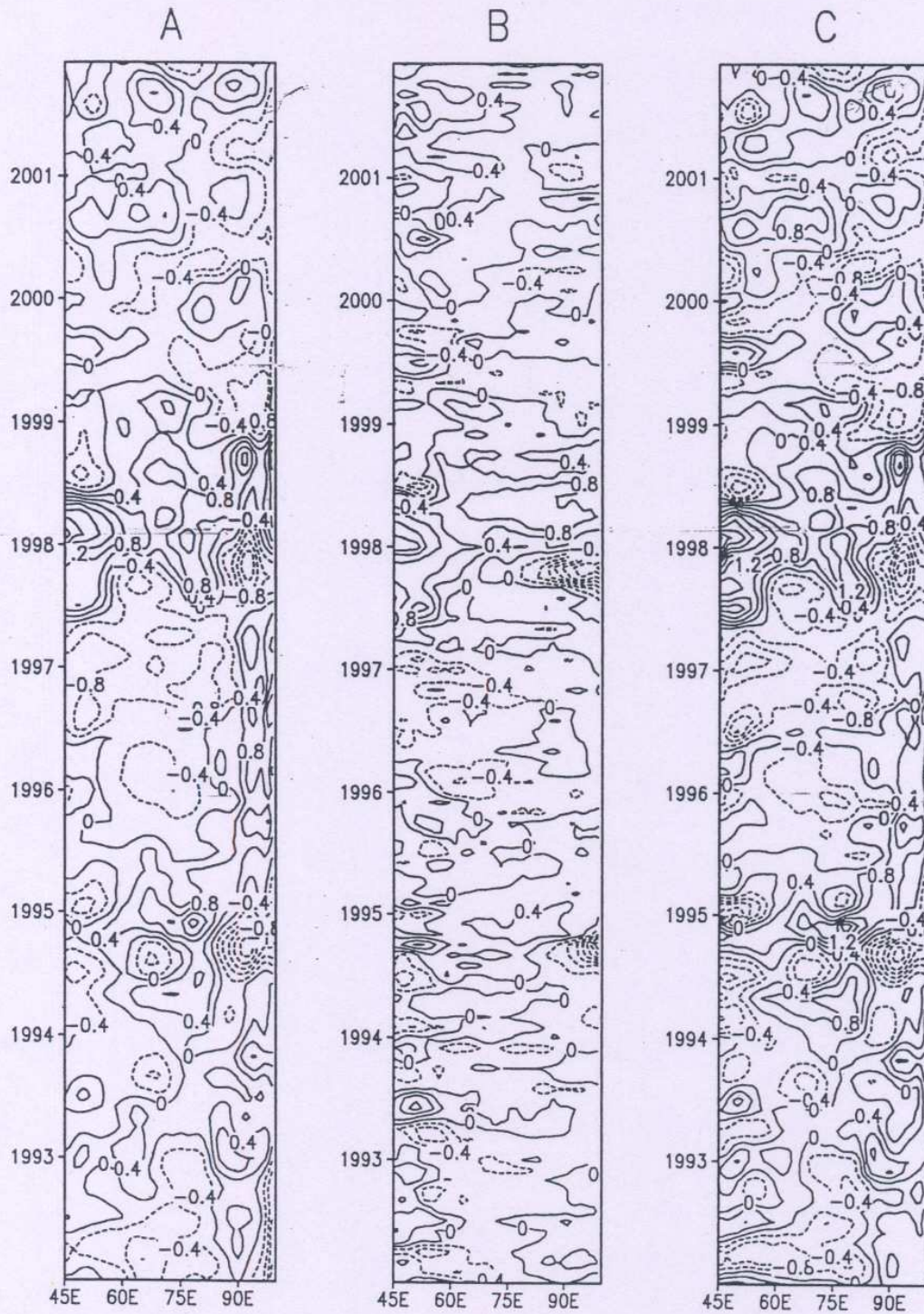
Fig. 2—Time-longitude plots of SSTA at the Equator (A) Model with 50 m MLD, (B) Reynolds observed, (C) Model with variable MLD (positive values are represented by continuous lines while negative are dotted).

corresponding model SST anomalies with uniform MLD and variable MLD respectively. The phase of the model anomalies agrees quite well with the observations. The time series of both model and Reynolds SSTA show periods of unusual cooling events in the Eastern Equatorial Indian Ocean (EEIO) during 1994 and 1997-98. However, the observed warming of western equatorial Indian Ocean is only seen partially in the model results. It is seen both in the model and in the observations that the EEIO cooling started in July 1997 and spread to early months of 1998 in the model, however the observations showed up to January 1998 only which may be due to the model's deficiency in accounting the influence of advection and diffusion terms through flux correction. As per IPCC, 1998 is likely to have been the warmest year[16], which was clearly seen both in the NCEP observations and in the model results. The model results with variable MLD and constant MLD show similar results however, the magnitude of SSTA in the variable MLD case is stronger in most of the integration period especially over the shallow MLD region. The model could also simulate a negative dipole observed in 1996 and 1992, however the signal of 1992 in the observations are weaker than model. It is important to note that 1997 and 1992 are strong *El-Nino* years but Indian Ocean responded completely in contrasting ways in these two years, which suggests that the dipole mode has an independent mechanism controlling it. Even though a large number of works have been produced[4, 6, 7, 17, 18, 19] on Indian Ocean dipole, the actual mechanism causing it is still to be understood clearly.

Generally cold SSTA sustain over western equatorial Indian Ocean and warm SSTA sustain in the EEIO. But exactly reverse situation was observed during 1994 and 1997-98. These unusual developments led to the recognition of an existence of a major mode, the so-called Dipole Mode (DM) in the Indian Ocean[4, 17, 18]. The SSTA in the observations and in the model demonstrate that positive DM events occurred during 1997 and 1994 and a negative DM[19] during 1996 and 1992. Figure 3 presents the SSTA during the peak phase of the DM events. The observed patterns of the anomalies (with cold SSTA off the coast of Sumatra and warm western Indian Ocean during positive DM and a reverse scenario in the negative DM) were well simulated by the in-house developed simple mixed layer model. The typical climatological scenario of high SST variability in the east, with a prevailing west-east SST gradient along

the equator, thus becomes completely reversed by November 1997 and October 1994 (Fig. 3A-D). The reversed pattern sustained up to the early months of 1998 and 1995 respectively.

Normally over the Indian Ocean south westerlies prevail in the boreal summer, north easterlies prevail in the boreal winter in the northern hemisphere. The south east trades dominate the southern tropical Indian Ocean throughout the year. The central and eastern equatorial westerlies were observed during monsoon transitions, most evident in the fall. Strong and persistent westward anomaly of the winds occurred in the equatorial Indian Ocean east of 60° E, in October 1994 (Fig. 4E). Strong south easterlies occurred along the coast of Indonesia, which extended north of the equator. These winds pushed the line of zero zonal wind anomalies north of equator, which is quite unusual during spring and fall. It was observed that winds were weak west of 60° E, especially in the southern hemisphere, which may be causing the entrainment rate which led to warm the sea surface during 1994 dipole months.

During 1997 the SE trades have displaced abnormally equatorward, which are almost similar to the wind pattern of 1994 (Fig. 4A). The warming in the western Indian Ocean was caused by anomalous winds and the resultant meridional advection[16]. This indicates that the advection is one of the main phenomena which have to be considered for clear evolution of surface warming especially in the western region. Also they described that cooling over the Sumatra coast is due to wind induced upwelling and shoaling of the thermocline. The positive DM events similar to 1997 but weaker in magnitude have occurred during 1994. Figure 4 (A,E) shows that the wind anomaly during positive DM events of 1997 and 1994. In contrast to this 1996 was one of the negative DM (event) years. The wind anomaly of October 1996 is shown in Fig 4C, which is of quite reverse pattern to the 1997 and 1994.

In the present study we have compared the wind anomaly and TOPEX/POSEIDON SSHA with SSTA. Figure 4(A, E) shows the SSHA of November 1997 and October 1994 respectively. Strong south easterlies near the coast of Sumatra might have caused a decrease in the sea level height which is indicated by the negative anomaly of SSH (light shaded) in the positive DM. The negative SST and SSH anomalies in the eastern basin are well correlated because the strong winds have caused the upwelling of cold water and directly affecting the SST as well as SSH.
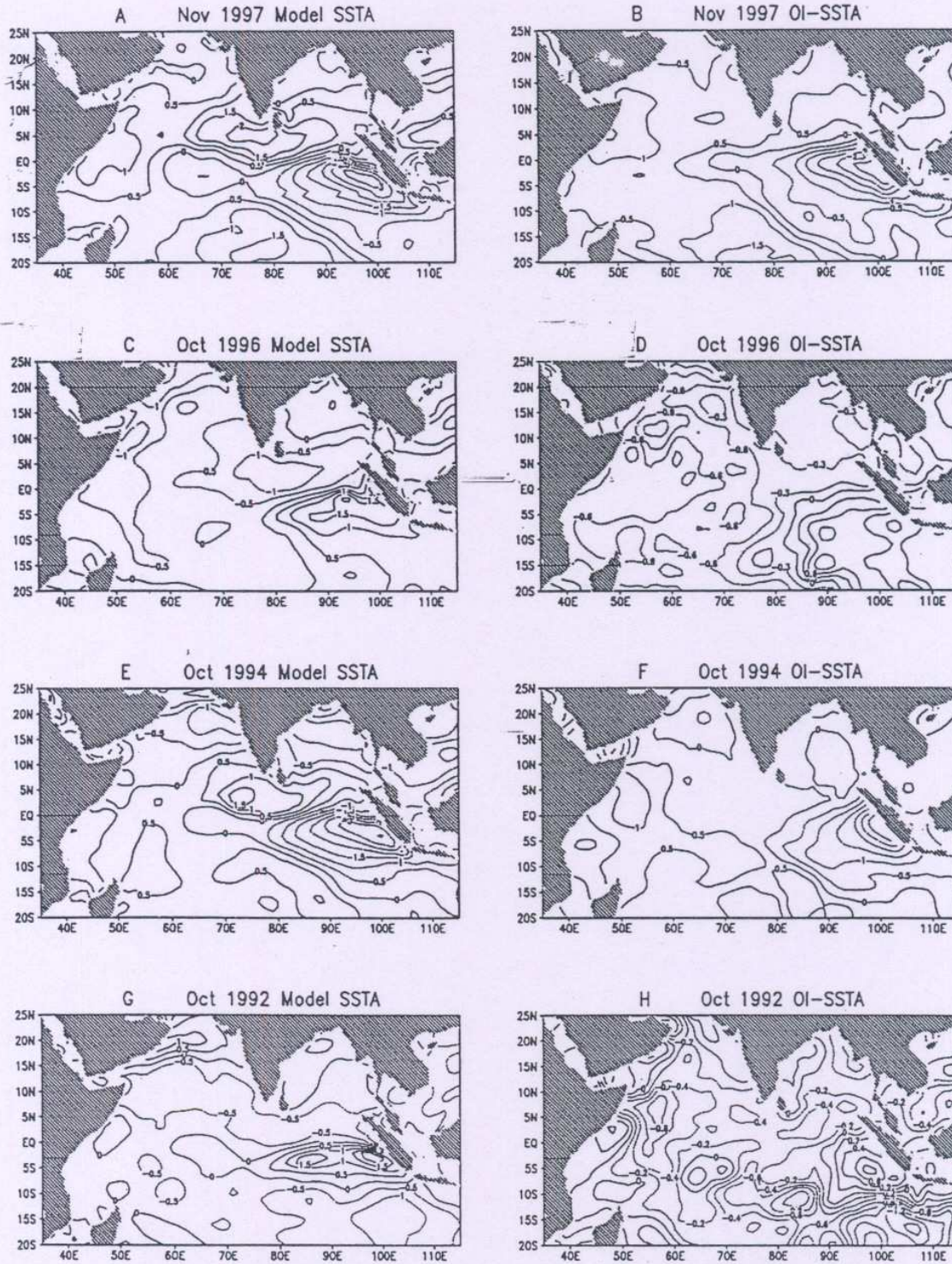
Fig. 3 — Model (left panel) and Reynolds (right panel) SSTA during extreme dipole events (contour patterns are similar to Fig. 2).
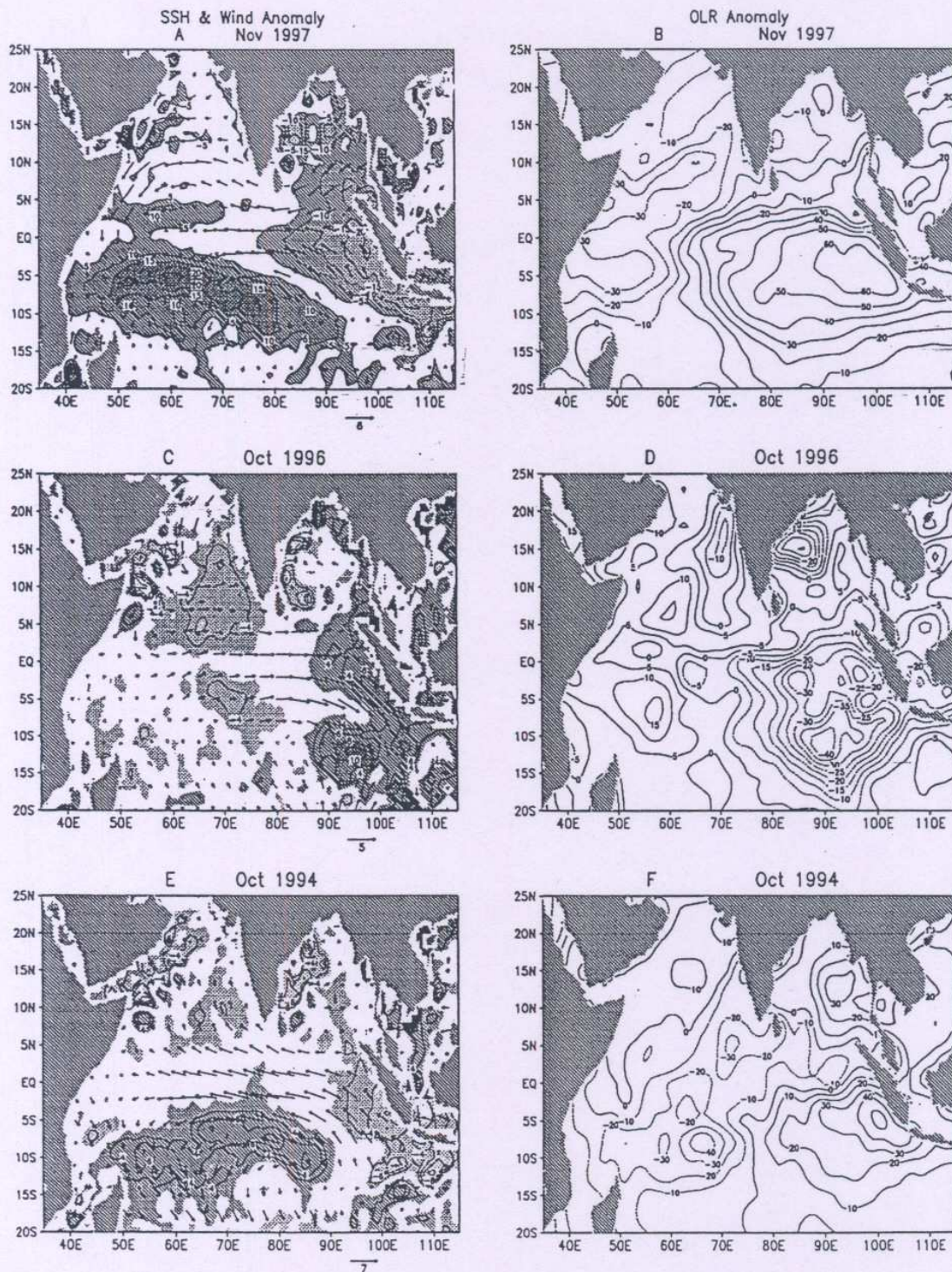
Fig. 4 — SSHA (cm), Wind anomaly (m/s) and OLR anomaly (W/m^2) during extreme dipole events (SSHA are darkly shaded if > 5 cm and lightly shaded if < -5 cm, contour patterns are similar to Fig. 2).
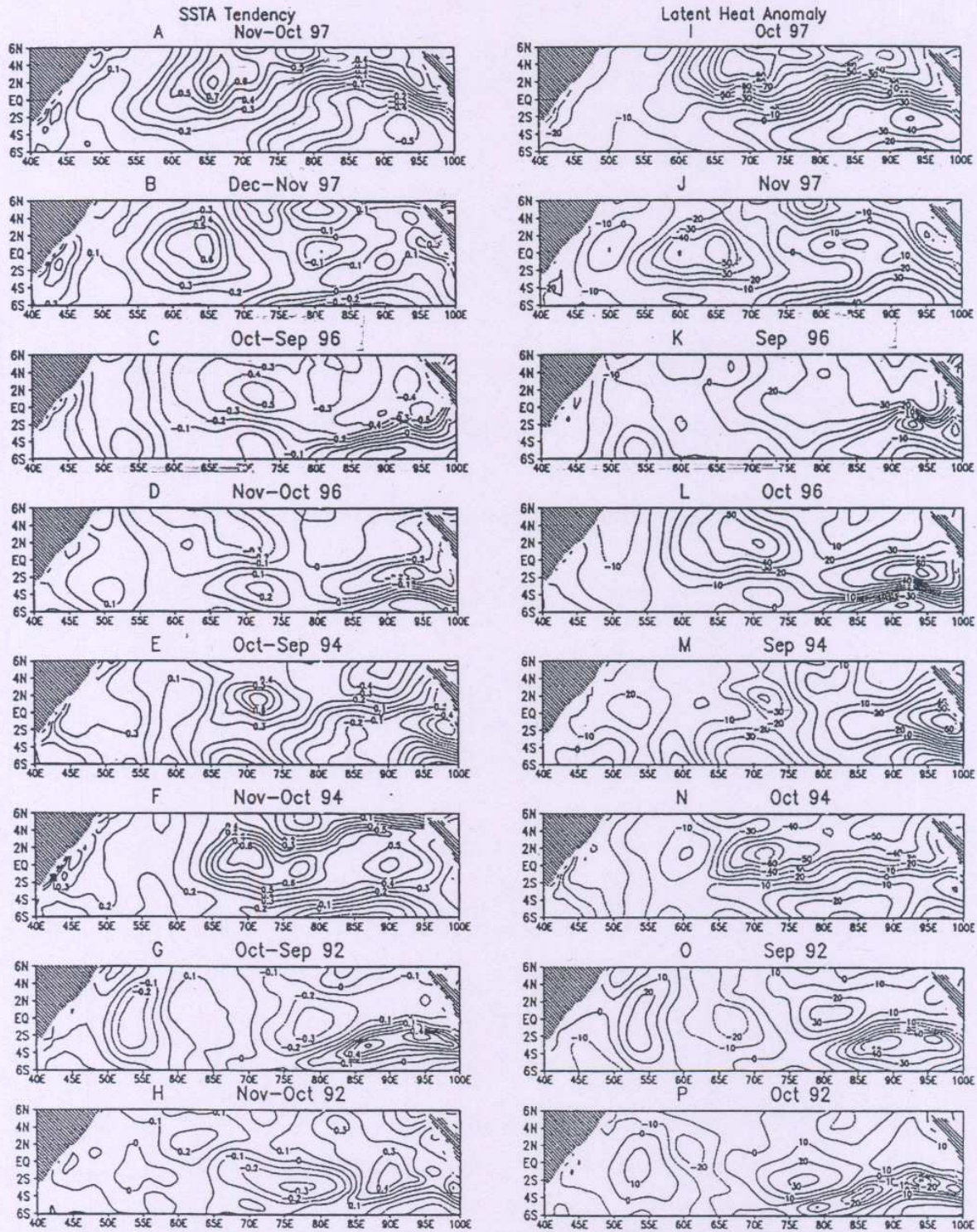
Fig. 5 — (A to H) SST tendency (°C/month), (I to P) latent heat flux anomaly (W/m²) (contour patterns are similar to Fig. 2).
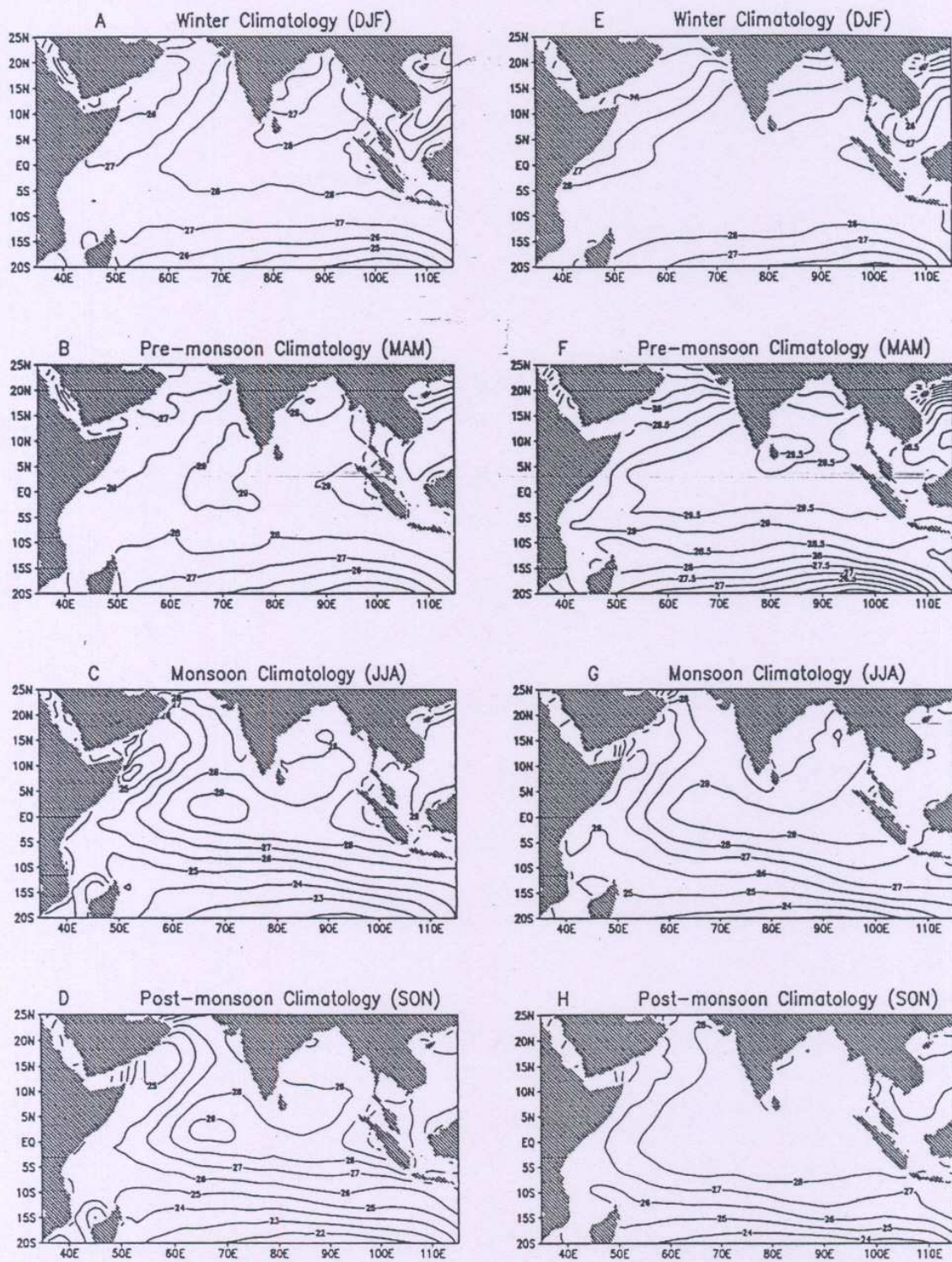
Fig. 6 — (A to D) Seasonal climatological model SST, (E to H) seasonal climatological observed SST.

However, Figs 3 and 4A, C, E show that the maximum SSHA (positive) were not correlated to the maximum SSTA (positive). The SSHA during November 1997 and October 1994 have similar negative anomalies in the eastern Indian Ocean and positive anomalies in the western Indian Ocean. But this pattern was very strong in 1997 than in 1994. In contrast to this, SSHA in October 1996 (Fig. 4C) was positive in the Eastern Indian Ocean and negative in the north western Indian Ocean. Figure 4 (B, D, F) shows that OLR (NOAA) anomaly over the EEIO was positive where the SSTA was negative. This clearly indicates that over the EEIO, during 1997 and 1994, there was a reduced convective clouds and hence reduced rainfall, whereas the western Indian Ocean had negative OLR anomaly and hence more rainfall. Thus the OLR distribution in November 1997 and October 1994 indicate greater (than average) rainfall in the western and north western Indian Ocean and less rainfall in the eastern Indian Ocean during the positive dipole years. A complete reverse OLR anomaly pattern was seen in the negative dipole year 1996 (Fig. 4D).

Figure 5(A-H) shows the model SSTA difference between two adjacent months and Fig. 5 (I-P) shows the latent heat flux anomaly. The patterns of latent heat flux variation are found to be in good agreement with those of month to month SSTA tendency. This further strengthens the role of latent heat flux in determining the SST in the equatorial Indian Ocean region especially during DM event years.

Figure 6 (A-D) shows the seasonal climatological model SST (1992 to 2001 mean) and Fig. 6 (E-H) shows the mean Reynolds[20] observed SST. The model climatology is well comparable to the observations in most of the study area (Fig. 6 A-H), except in winter season, when the maximum difference was found to be around 1°C near the Somali coast (Fig. 6A, E). Both the model as well as the observations showed large seasonal variation in SST in the western Indian Ocean, which is caused by the intense upwelling in response to strong monsoonal winds during boreal summer. In contrast the eastern equatorial Indian Ocean showed very less seasonal variability with SST over 28 °C throughout the year due to the positive net heat flux and the weak entrainment.

The annual mean of the Q_{cor} applied to the surface heat flux, varies from -60 W/m^2 to 30 W/m^2 in the entire model region (Fig. 7). The dipole mode index (DIM) is defined as the difference in the SST anomalies between the western and eastern parts of the Indian Ocean[7]. SST anomalies are averaged over 50° E -70°E, 10° S-10° N for the western Indian Ocean and over 90° E-110° E, 10° S-Equator for the eastern Indian Ocean[7]. Figure 8 compares DMI (five months running mean) from the model (the solid lines) with that of the observations (the dotted lines). The model DMI agrees well (especially in phase) with
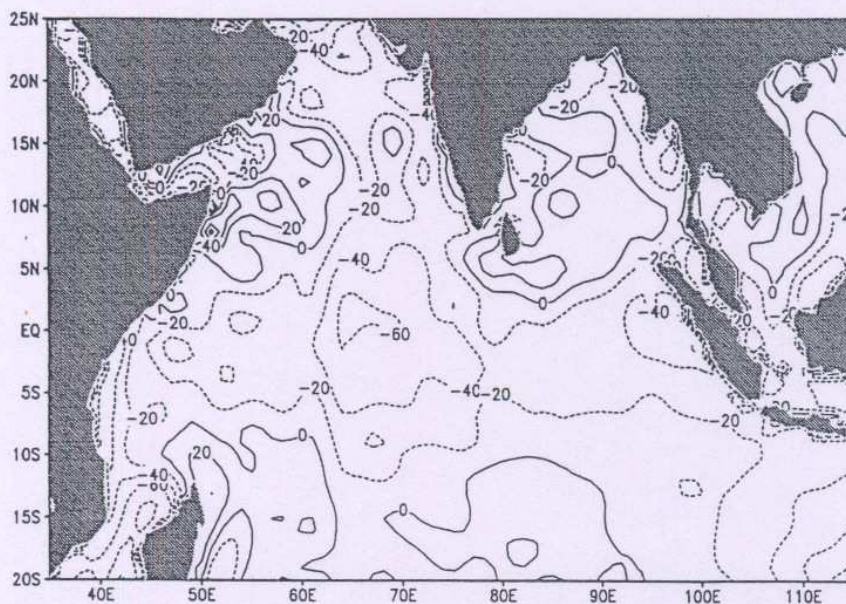

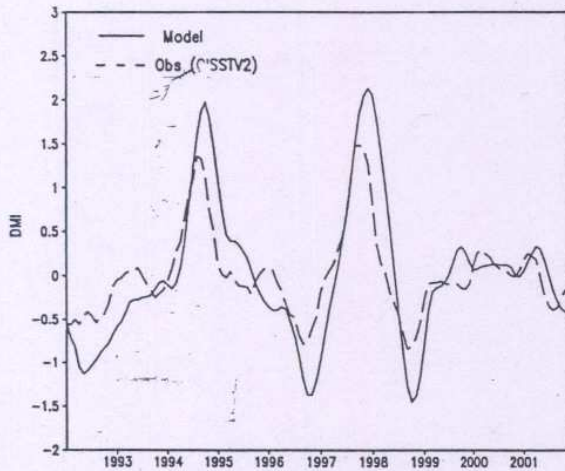Fig. 7 — Annual average surface heat flux correction (W/m^2).

Fig. 8 — Dipole mode index for the period 1992-2001

the observed index during 1994 and 1997, however the observed signals are little weaker than the model DMI.

The simple numerical model is capable of simulating SST fields over the north Indian Ocean ($35°E$-$115°E$, $20°S$-$25°N$). The model could successfully simulate the observed dipole of 1997 and 1994 especially the eastern cooling indicating that eastern cooling is mainly controlled by the surface heat flux. It was found that SSTA over the EEIO is directly related to the latent heat flux anomaly. The latent heat flux variation is found to be in good agreement with those of month to month SSTA tendency. SSTA was found to be negatively correlated with the OLR anomaly. The simple model can be easily coupled to a regional atmospheric model to study the air-sea interaction processes. The model will be further improved by including the advection term in it. The model will then be used to understand the role of surface heat flux, advection and entrainment in determining SST in different regions.

Acknowledgment

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