1. Introduction

The annual reversal of the monsoon wind (Fig. 1) causes dramatic changes in the upper-northern Indian Ocean. In addition to the dynamical changes, the strong southwesterly winds during boreal summer monsoon cause evaporative and entrainment cooling of the sea surface. It is interesting to note that sea surface temperature (SST) in the tropical Indian Ocean is quite warm having comparable magnitude to that of the western Pacific warm pool with an exception during the summer cooling in the western sector. This suggests that any small change in the SST may lead to large-scale ocean–atmosphere phenomena similar to that in the tropical Pacific. Though most efforts in past decades have gone into the understanding of the Pacific phenomenon, not many studies are available to understand the variability in the Indian Ocean. Some recent studies (Vinayachandran et al. 1999; Behera et al. 1999; Saji et al. 1999) show that there are internal dynamic and thermodynamic processes that lead to basin-scale air–sea interaction in the Indian Ocean and thus underlies the importance to understand the SST variability.

In tandem with the meager number of observational studies, there are very few modeling studies that address basin-scale SST variability in the Indian Ocean on seasonal to interannual scales. One of the early attempts to understand the seasonal cycle of SST in the northern Indian Ocean using an intermediate model is due to McCreary et al. (1993, hereinafter referred to as MKM). Recently, Murtugudde and Busalacchi (1999, hereinafter referred to as MB) simulated the interannual variability of the tropical Indian Ocean SST in addition to the seasonal cycle using a reduced gravity primitive equation model.

In this study, we have addressed some of the key dynamics and thermodynamics that are responsible for the SST variability in the tropical Indian Ocean using observed SST data (Reynolds and Smith 1994, hereinafter...
referred to as Reynolds SST) and ocean model simulations. We adopt the 2.5-layer model due to MKM to simulate the interannual SST using the surface fluxes derived from the National Centers for Environmental Prediction–National Center for Atmospheric Research (NCEP–NCAR) reanalysis data (Kalnay et al. 1996). The model was successfully used by Behera et al. (1998) to simulate the interannual SST variability in the northern Indian Ocean. Simplicity and economy of the model make it possible through several model experiments to understand various processes that are responsible for the SST anomalies. Special emphasis is given to the SST variability in the southern tropical Indian Ocean (STIO), as it supplies about two-thirds of the moisture that accounts for the Indian summer monsoon precipitation (Hastenrath and Greischar 1993). This region is also shown to be highly correlated to the Australian winter precipitation (Nicholls 1989), underlying the importance of understanding SST variability in the context of the Asian–Australian monsoon. In the next section, a brief description is given about the model and the surface forcings. Section 3 deals with the discussion of the seasonal cycle and interannual variability of model SST and process studies. Section 4 provides the summary.

2. The model and surface forcing

The model has two active layers overlying a deep motionless layer of infinite depth (Fig. 2a). The first layer separates into two sublayers (Fig. 2b) with the initiation of entrainment and detrainment processes. The upper sublayer entrains or detrains water due to turbulence generated by wind stirring and cooling at the surface (Kraus and Turner 1967). The lower sublayer remains isolated from the surface forcing but can be engulfed into the upper sublayer during strong entrainment. The model’s second layer, in turn, is driven by the mass and heat fluxes derived from the first layer.

![Fig. 1. Surface winds derived from the NCEP-NCAR reanalysis data for winter (Dec-Feb) and summer (Jun-Aug) months.](image1)

![Fig. 2. A schematic diagram of the model layer structure (a) in the simpler form with single upper layer and (b) in a more complex form when upper layer is divided into a upper-mixed sublayer and a lower fossil sublayer.](image2)
The uppermost sublayer is considered as model mixed layer and its temperature is considered as the representative of model SST. For a detail discussion on model equation, initial, and boundary conditions readers may refer to MKM.

The NCEP–NCAR fields through the period 1982–1991 used for surface flux computations are net solar radiation (incoming fluxes–outgoing fluxes), air temperature, specific humidity, and wind components. It may be noted that the model-simulated SST fields are used for the sensible and latent heat flux computations instead of the observed SST. The drag coefficients used for these heat flux computations are the same as given in MKM. The wind stress fields are computed from the wind components with a drag coefficient $C_D = 1.5 \times 10^{-3}$ and air density $\rho = 1.2$ kg m$^{-3}$. For simplicity, the model calendar is considered to be of 360 days having 30 days in each month.

3. Results and discussion

The model is spun up by a 10-yr integration with mean surface forcing for the period 1982–91. Since a quasi-equilibrium state is reached by this integration, the solutions from the end of this integration are considered as model initial conditions for the interannual run in which the model equations are further integrated for 10 yr with the interannually varying monthly surface forcing (control run). Values of the model variables at day 15 of each month are considered to represent the monthly mean values. Monthly anomalies are departures from the corresponding 10-yr averaged monthly mean values.

Figure 3 shows the plots of mean February and August model SST fields, which roughly constitute the annual cycle of SST in the basin. The figure also shows the difference between the model SST and the Reynolds SST. The tropical Indo–Pacific warm pool is seen in the eastern Indian Ocean with SST values greater than 28°C, which has less seasonal variability. In contrast, the western Indian Ocean SST has large seasonal variability caused by the intense coastal upwelling (entrainment plots of Fig. 4) and offshore advection of cold waters in response to strong monsoon winds during boreal summer. The summer cooling in the west brings out the zonal structure of the SST field in the equatorial region.

The SST variability in the STIO is found to coincide with the annual march of the radiative flux. As can be seen, the mixed layer (Fig. 4) in STIO is thinner during austral summer as compared to that in winter. Therefore, a higher-radiative flux, in the absence of strong entrainment (Fig. 4) and latent heat loss (figure not shown), produces a warmer SST. Zonal advection of warmer waters by the South Equatorial Current also contributes to the higher SST in the western part of STIO. Stronger entrainment, higher latent heat loss, and a decrease in radiative flux produce the colder SST in the austral winter. The basin gains (losses) heat north (south) of 10°S in the annual average, for example, Hastenrath and Lamb (1979). As found from the model annual heat budget, this is mainly because of the higher (lower) heat gain due to radiative flux as compared to the heat loss due to latent heat flux north (south) of 10°S. A zonal-averaged southward (northward) meridional heat advection in the upper (lower) layer balances the heat budget (similar to Fig. 5 in MKM) and helps in regulating the SST.

The difference between the model and Reynolds SST is less than 1°C in most parts of the basin (Fig. 3). Major differences occur in the eastern equatorial region and the northern Arabian Sea. The model we used in this study does not include the saline process, which may explain the colder model SST as compared to the data. The formation of a barrier layer in the region, below the seasonal thermocline due to freshwater flux from precipitation, plays an important role in the thermodynamic structure of the eastern equatorial Indian Ocean. The model SST in that region. Due to the close proximity to land regions, we expect contamination of the re-analysis data resulting in erroneous specific humidity fields but this needs further investigation.

Interestingly, the SST differences are much less in the Bay of Bengal in spite of the lack of freshwater treatment in the model. The closer agreement, thus, might be a result of some deficiency in the forcing field that compensates for the noninclusion of the saline process in the model. It is difficult to discuss further the deficiency in the forcing field and the model performance due to lack of observation in the region. Similarly, the higher differences seen in the southeast corner of the model domain might be again due to a deficiency in the forcing field as well as the model configuration, which neglects influence of the Indonesian throughflow (Godfrey 1996; Murtugudde et al. 1998).

a. Interannual variability

Figure 5 shows the longitude–time cross section of the model and Reynolds SST anomalies along the equa-
tor. Evolution of model warm anomaly is remarkably similar to that of the observed with a slight variation in the magnitude. In particular, the warm events of 1982–83, 1987–88, and 1990–91 are well captured. The only difference being the presence of a small period of cooling in the model simulation during the warm phase of 1990–91. Apparently there is a slight eastward tilt in the warm anomaly bands suggesting El Niño-like episodes (of a smaller amplitude) in the equatorial Indian Ocean in conjunction with that in the Pacific. Further analysis of long time series Reynolds SST data for the period 1950–95 confirms the concurrence of warm events in the equatorial Indian Ocean and El Niño in the Pacific. However, Indian Ocean warming only in couple of occasions exceeds 0.5°C.

A point to point comparison between the anomalies of model SST and Reynolds SST is carried out over the whole model domain by using statistical scores like root-mean-square errors (rmse) and correlations. Rmse are found to be less than 0.4°C in most parts of the basin (figure not shown), except for small regions near the Somali coast and in the central part of the basin. Lack of intense cooling features in the observed data might be a reason for the higher rms values near the coast. The correlations are also higher than 0.6 in most regions except for the Somali coast and a small region in the central-equatorial Indian Ocean.

Standard deviations of model and Reynolds SST anomalies are plotted in Fig. 6 for further comparison. The spatial patterns of the two plots are in general agreement; however, the magnitude of the standard deviation in the data is higher compared to that of the model. Higher (lower) variability is revealed in the western (eastern) equatorial Indian Ocean and STIO. Interestingly, the model’s second-layer depth (which can be considered as the model thermocline depth) has larger variability (Fig. 6) in the western STIO that links to the SST variability. A similar connection between the SST variability in the region and the thermocline depth movement related to Ekman pumping was also sug-
gested by MB. SST variability in the Somali coastal region (Fig. 6) is also seen to be related with the variability in the second layer. In contrast, the second-layer depth shows far less variability in the eastern STIO as compared to that in the mixed-layer depth (Fig. 6). Thus, mixed-layer physics has a greater role in the SST variability seen in that region.

Figure 7 shows the time series of the model and Reynolds SST anomalies in four individual regions of interest for further comparison. These regions correspond to the western equatorial, eastern equatorial, southeastern tropical, and southwestern tropical Indian Ocean. As it is seen, the phase of the model anomalies agrees quite well with the data and the difference in the magnitude is less than 0.5°C. The time series of both model and Reynolds SST anomalies show periods of warming (cooling) during 1982–83, 1987, and 1990–91 (1984–86 and 1988–89). In the western equatorial region, the peak in the warm anomaly in the model is less than that in the data during 1983. However, the corresponding peak in the southeastern tropical Indian Ocean is well captured by the model. A strong cold anomaly is also noted in this region prior to the warming of 1983 and 1987.

Though the time series of SST anomalies in the two equatorial regions are in phase, the corresponding anomaly time series between the two STIO regions are sometimes out of phase with a correlation of $-0.56$ ($-0.51$) for model (data). These out of phase periods are found to be related with a dipole mode of variability in the STIO SST anomalies as discussed in the section 3c. The model SST anomalies are produced by the anomalies in surface forcings, subsurface entrainment, and horizontal advections. Wind forcing is one component of surface forcings. It is found that the SST anomalies are in opposite phase with the corresponding anomalies of wind stirring (proportional to $|\tau|^{3/2}$). A detailed discussion on the relative importance of different processes in producing the SST anomalies is provided in the following sections.
Fig. 5. Longitude–time plots of model and Reynolds SST anomalies along the equator. A 3-month running mean is applied to smooth the Reynolds SST anomalies. Contour interval is 0.3°C and positive values are shaded.
b. Process study

The terms in the model mixed-layer temperature equation (see MKM for details) are diagnosed to obtain quantitative picture of the underlying processes:

\[ T_{ml} = Q/h_m - \text{WENT} - v_1 \cdot \nabla T_m + \kappa_\tau \nabla^2 T_m. \]

Here \( Q \) is the total heat flux that includes radiative flux, latent heat flux, and the sensible heat flux; \( \text{WENT} \) represents the entrainment term; \( \kappa_\tau \) is the horizontal mixing coefficient; \( T_m \) and \( h_m \) are the mixed-layer temperature and depth, respectively; and \( v_1 \) is the velocity of the first layer (see Fig. 2b). Each term on the right-hand side of the equation determines a particular process, namely, surface heating, entrainment, horizontal advection, and diffusion. These terms determine the temperature tendency term on the left-hand side of the equation.

Figure 8 shows the time series of the three dominant processes in the four regions corresponding to that of the SST anomalies shown in Fig. 7. Anomalies of the latent heat flux are mostly seen to dominate the heat budget in all four regions. The effect of the entrainment process is negligible in the equatorial region, leaving the other two processes to determine the SST anomalies. Comparison of the two figures (Figs. 7 and 8) suggests that the anomalous warming during 1982–83 is mostly explained by the positive anomalies in the latent heat flux. Reduction in the mean seasonal latent heat loss associated with lower wind speed leads to the positive anomalies. A strong oscillation seen in the anomalies of latent heat flux of the southeastern tropical Indian Ocean during 1982–84 corresponds to similar trends in the SST anomalies. This region has large seasonal evaporative flux as discussed in section 3. Changes in boreal monsoon winds, thus, produce large interannual variability.

The cooling in 1982 and 1986, which is prominent in the southeastern tropical Indian Ocean, is due to a higher latent heat loss while anomalies in the radiative flux and entrainment counteract each other. It is found
that the stronger trade winds carried away the moisture from the region and the outflow of moisture kept the region cloud free explaining the positive anomaly in the radiative flux. Interestingly, combined effects of the radiative and latent heat fluxes are responsible for the 1987 warming in the eastern equatorial region. On the other hand, the anomalies of the radiative flux and zonal advection (not shown) contribute to the western equatorial warming. The latent heat flux process dominates the warming of the two STIO regions during 1987 and 1991.

The model and Reynolds SST anomalies are used in empirical orthogonal function (EOF) analysis for determining the dominant modes of variability. An El Niño–Southern Oscillation (ENSO) related warming (Wallace et al. 1998; Latif et al. 1998) is found covering the whole basin in the first mode of the EOFs. Interestingly, the second EOFs (Fig. 9) reveal a dipole mode of variability in the STIO. The eastern pole has cold anomaly during 1982, 1986–87, and 1990–91. The cooling in the southeastern tropical Indian Ocean in the latter two occasions proceeds a warm ENSO event in the Pacific. However, during 1982 when the ENSO event is evolving toward a warm phase in the Pacific there is a polarity reversal in the Indian Ocean dipole. The beginning of the cooling, as seen in the time series of the principal components (Fig. 9), corresponds to the start of the negative phase in the Southern Oscillation index (SOI). This brings out a possible close link in the Indo–

**c. EOF analysis**

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Pacific region prior to an ENSO event. At the same time, a correlation of only $-0.5$ (6-month lag) with the SOI gives some independent characteristics to the STIO dipole mode.

The connection of Indian Ocean variability with ENSO, through the atmospheric processes, like shifts in the Walker circulation and associated changes in radiative flux and wind-forced evaporation, is discussed in Venzke et al. (1997). There also exists a possible oceanic connection through changes in the transports of the Indonesian throughflow (Meyers 1996). Lack of the throughflow effect in the present model probably explains the lower values seen in the model EOF (Fig. 9). The ENSO-related changes in the equatorial Indian Ocean could also remotely affect the sea level changes in the Indonesian throughflow region as indicated by Yamagata et al. (1996). In addition to the ENSO signal, Meyers (1996) found non-ENSO surface cooling off Java related to the equatorial Indian Ocean winds.

Some recent studies (Vinayachandran et al. 1999; Behera et al. 1999; Saji et al. 1999) show a dipole in the SST anomalies in the near-equatorial region. This dipole is characterized by a cold eastern pole off Sumatra. Some of the dipole events in the Indian Ocean are found to occur during non-ENSO years (Saji et al. 1999) differentiating them from the Pacific events. However, the dipole in the present study is found to be more southerly (near-subtropical) and is closer to a dipole pattern suggested by Nicholls (1989) that explains Australian winter precipitation, not closely associated with ENSO. It is quite possible that the near-equatorial dipole and the near-subtropical dipole are two distinct phenomena interacting with each other in some years (e.g., in 1982, when both occurred) but it needs further elaborate study.
The process studies in the previous section show the dominance of latent heat flux in determining the SST anomalies. EOF analysis of all the processes further reiterated the dominance of the latent heat flux anomalies for the evolution of the dipole in the model SST. Figure 10 shows the second mode of EOFs for the anomalies in the latent heat flux computed from the model and Reynolds SST. A dipole is evident in the two plots with comparable spatial pattern as seen in the EOFs of the SST anomalies (Fig. 9). It is found out that the stronger (weaker) trade winds lead to the anomalous heat loss (gain) due to higher (lower) evaporation in the southeastern (southwestern) tropical Indian Ocean.

d. Sensitivity experiments and process index

In order to understand the relative role of the interannual variability present in wind and heat forcings on the evolution of model SST anomalies, two sensitivity experiments are carried out. In each of these experiments, interannual variability in one of the forcings is dropped in order to evaluate the importance of the other forcing in the model solution. In the first sensitivity experiment (SE-1), the interannual run is carried out by replacing the interannual heat flux with the climatological (10-yr averaged) heat flux. The magnitude of the SST anomalies in this case on an average is reduced by more than 50%—60%—being closer to the control experiment only in some part of the basin as discussed below.

In the second experiment (SE-2), the interannual wind forcing is replaced by the climatological wind forcing, while retaining the interannual thermal forcing. The SST anomalies in this case are found closer to the control experiment in most parts of the basin. This suggests the dominance of the surface heat flux in the evolution of the model SST anomalies. Though these results are consistent with that of MB, the interannual variability in
wind forcing in their case seems to have a slightly greater role in shaping the SST anomalies. A complete analysis with similar surface forcings will be helpful for understanding the differences. It may also be noted that the latent and sensible heat flux computations in the SE-2 experiment use interannual scalar winds, thus retaining the influence of wind in the model thermodynamics. However, the lack of interannual variability in wind-forcing terms of the momentum equations and the entrainment parameterization (by the turbulent kinetic energy production term) removes direct dynamical effect on model SST, for example, through upwelling and turbulent mixing. So as can be seen in the following discussions, the impact of the wind and heat forcings is largely disconnected in the dynamically active regions.

The correlation between the anomalies of the two sensitivity runs and the control run for individual months provide an index for understanding the direct influence of each of the two forcings. Plots of the correlation for winter and summer months are shown in Fig. 11. In winter months, the regions of less correlation in the case of SE-2 correspond well with the regions of high correlation in the case of SE-1. However, such simple asymmetry in the model responses is not clearly apparent during summer months in some regions, for example, around 5°S in the central part. The region in the central STIO, south of 5°S, is strongly dominated by the wind forcing during boreal winter with values greater (smaller) than 0.6 in case of SE-1 (SE-2). This is the region where thermocline movements associated with the variability in Ekman pumping (see also MB) influence the SST anomaly.

The coastal regions in the Arabian Sea and around Sri Lanka have shown winter to summer changes in the response to the two forcings. During boreal summer, the influence of wind is more pronounced especially near the African coast. The strongest influence of wind is evident around 5°–10°N with 0.9 correlation in the SE-1 case and a corresponding less than 0.6 correlation in
SE-2 case. This region is dynamically active during the summer monsoon season and the wind-induced entrainment dominates the surface cooling. In addition, the region east of Sri Lanka has also shown influence of wind (Fig. 11) during the summer season. Vinayachandran and Yamagata (1998) recently discussed the importance of the summer monsoon winds in the generation of a thermal dome off Sri Lanka.

4. Summary

SST variability in the tropical Indian Ocean is studied using an intermediate 2.5-layer ocean model and Reynolds SST. Regions in the western and southeastern tropical Indian Ocean have large seasonal and interannual variability in the model and observed SST. Coastal upwelling due to monsoon winds in boreal summer causes the seasonal SST variability in the western equatorial region and the northwestern Arabian Sea. Seasonal changes in the solar radiation, latent heat flux, and entrainment produce the annual cycle of SST in the STIO.

It is found that the interannual heat flux plays a greater role in the evolution of simulated-SST anomalies in most part of the model domain—latent heat flux dominates the heat budget. Direct inference of interannual wind forcing through upwelling is only apparent in the coastal region of Arabian Sea and around Sri Lanka during boreal summer and a small region in the central STIO during boreal winter. SST anomalies from both the model and data show three warm episodes during 1982–83, 1987, and 1990–91 and two cold episodes during 1984–86 and 1988–89. The warm events correspond to the Pacific El Niño.

An interesting dipole mode of variability is seen in the second EOFs of the model and Reynolds SST anomalies. The eastern pole in the southeastern tropical Indian Ocean is anomalously cold prior to an ENSO event. The cooling begins along with the negative phase of the SOI but peaks 4–8 months prior to the peak of the SOI (Fig. 9). In addition, a correlation of only −0.5 suggests the dipole evolution might be independent of ENSO. In one of the events, during 1982–83, the two poles in the STIO actually swap their roles by the time the SOI reaches its peak. This shift of the dipole mode from one regime to another during the ENSO period clearly indicates a proper mechanism within the Indian Ocean for the dipole evolution.

A similar dipole mode in the latent heat flux anomalies is related to the dipole in the SST anomalies. As it is inferred, stronger winds activate evaporation and entrainment leading to the formation of cold pole in the STIO at the beginning of the event. The offloading of the moisture into the western STIO by seasonal southeasterly winds during austral winter (Fig. 1) leads to the decrease in seasonal evaporation and the occurrence of the warm pole. Some aspects that can affect the evolution of the dipole, for example, Indonesian throughflow, variability in intermediate layers, and ocean–atmosphere interactions remain unanswered. This should be studied with more sophisticated models in the future.

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Planck-Institut für Meteorologie, Bundesstrasse 55, D-20146 Hamburg, Germany.


