3.1 Introduction

Intraseasonal and interannual variability over the tropical Indian Ocean is one of the important phenomenon which have really attracted the attention of national and international communities working for regional and global climate systems. The intraseasonal variability is enhanced due to the internal oceanic instability and the response to the intraseasonal atmospheric forcing (Reppin et al. 1999; Sengupta et al. 2001a; Waliser et al. 2003). The intraseasonal variability in the atmospheric forcing has different impacts on SST. To be more specific, the intraseasonal variability in SST has large influence on Asian summer monsoon (e.g., Wang and Xie, 1997; Lawrence and Webster, 2002; Webster and Hoyos, 2004). Apart from the intraseasonal variability, the interannual variability is also observed to have large influence on air sea interaction processes over the tropical Indian Ocean. Infact studies on interannual variability of the tropical Indian Ocean came into limelight with the discovery of IOD. In this chapter the intraseasonal variability in the Indian Ocean is studied before discussing the interannual variability IOD.

3.2 Intraseasonal signals in the daily high resolution blended Reynolds (REY) SST product over the tropical Indian Ocean and their validation

SST is an important parameter in many operational and research activities, ranging from weather forecasting to climate research. In this chapter the focus is on validation of the new available daily SST product produced at the National Oceanic and Atmospheric Administration (NOAA) as described by Reynolds et al. (2007). Weekly and
monthly Reynolds SST product was provided by Reynolds and Smith (1994) and Reynolds et al. (2002). These weekly and monthly products use IR satellite data from the Advanced Very High Resolution Radiometer (AVHRR) and in situ data from ships and buoys. The analyses were performed weekly on a 1° spatial grid by optimum interpolation (OI) with a separate step to correct any large scale satellite biases relative to the in situ data. The Reynolds and Smith (1994) and Reynolds et al. (2002) weekly OI are referred to as OI version 1 (OI.v1) and OI version 2 (OI.v2), respectively. The techniques for these analyses were originally designed in the late 1980s and early 1990s when there was only one AVHRR satellite instrument producing SSTs. The satellite derived SSTs were from the multichannel SST products that have been constructed operationally from the five-channel AVHRR by NOAA's Environmental Satellite, Data, and Information Service (NESDIS) produced at the University of Miami's Rosenstiel School of Marine and Atmospheric Sciences. Even though weekly SST's and the seasonal changes were comparable with the observations, it did not reproduce the intraseasonal and lower frequency variability on time scales longer than a few days (Senan et al. 2001).

In the late 1990s, more satellite data sets were made available and there have been frequent comparisons or study [Chelton and Wentz (2005)] of other data and analyses with the OI.v2. These results have strongly suggested that spatial and temporal improvements were needed for better understanding the ocean. The purpose of the new daily SST product (Reynolds et al. 2007) was to refine the OI.v2 analysis procedure and produce a higher resolution (.25° spatial and 1 day temporal) reanalysis product. The present study aims to examine the advantages and disadvantages of this daily product and to focus mainly on its capability to capture the intraseasonal and lower frequency variability on time scales longer than a few days which infact was absent in Reynolds weekly SST product. Further an attempt has been made to validate this product over the Indian Ocean region using the available buoy observations and TMI SST data. The present study will help the users aware of its merits, demerits and limitations in the Indian Ocean region.
3.2.1 A statistics of TMI and REY SST

Figure 3.1: (a) The correlation between TMI and REY SST. (b) RMS difference between TMI and REY SST. (c) TMI SST standard deviation. (d) REY SST standard deviation for the period 1998-2007.

A detailed analysis of daily REY SST over the tropical Indian Ocean is necessary to quantify its credibility. The accuracy of the REY SST was tested using Correlation, Root Mean Square (RMS) error and standard deviation (STD) analysis between TMI and REY SST. Figure 3.1 shows (a) The correlation between TMI and REY
SST. (b) RMS difference between TMI and REY SST. (c) STD of TMI SST. (d) REY SST standard deviation. The correlation of SST between TMI and REY SST shows values above 0.8 (above 99% confidence level) over most of the tropical Indian Ocean basin (Figure 3.1d). Correlation above 0.9 is seen in the western Arabian Sea. The correlation is observed to be above 0.60 in the eastern equatorial Indian Ocean. The RMS difference between TMI and REY SST is shown in Figure 3.1c. Over the tropical Indian Ocean, between 10°S to 10°N, most of the Arabian Sea and southern Bay of Bengal, the RMS difference is observed to be less than 0.7°C. RMS differences of the order of 0.8°C are seen in the coastal regions (especially Sumatra), where the correlation is also less. STD of TMI and REY SST shows similar pattern with almost similar amplitude. Both the products showed more prominent SST variation over the region south of 5°S and western Indian Ocean whereas very less SST variation was observed in the east and central equatorial Indian Ocean.

The basin wide average TMI SST and REY SST for the period 1998 to 2007 over 45°E to 110°E, 25°S to 25°N; hereafter Indian Ocean (IO), 65°E to 85°E, 10°S to 3°S; hereafter BOX1, 80°E to 90°E, Equator to 5°N; hereafter BOX2 and 85°E to 90°E, 10°N to 14°N; hereafter BOX3 are drawn in Figures 3.2a, 3.2b, 3.2c and 3.2d respectively. The REY SST over all regions are comparable with the TMI SST. Over IO region, REY SST amplitude was observed to be less than TMI SST. However, over BOX1, BOX2 and BOX3 REY SST is seen to be comparable with TMI SST.

3.2.2 Validation of REY SST over the Tropical IO

It is worthwhile to calibrate those products with the available in situ observations. The SST observations DS1, DS2, DS3, DS4 and DS5 buoys are utilized for this purpose. Figure 3.3 shows a comparison of SST from TMI, buoy and REY at viz five buoy locations DS1 (69.3E, 15.3N), DS2 (72.5E, 10.8N), DS3 (87E, 13N), DS4 (89E, 19N) and DS5 (82E, 16N) for the year 1998. At all the five locations, the REY SST is comparable with the TMI and DS SST. The REY SST captures the intraseasonal variability on time scales longer than a few days.
Figure 3.2: TMI (dotted line) and REY SST (dashed line) averaged over IO, BOX1, BOX2 and BOX3.
Figure 3.3: The comparison of TMI SST (dotted line), REY SST (dashed line) and DS SST (solid line) from WHOI mooring.
Premkumar et al. (2000) noticed abrupt drop in SST by more than three degrees at 15.5°N, 69.25°E in the Arabian Sea in early June 1998. The steep fall in SST seen in DS1 buoy observation in early June was due to the presence of severe cyclonic storm in the Arabian Sea during June 3-9. This fall in SST is clearly seen in the REY SST. During this time REY SST, TMI SST and DS SST show comparable results at DS1 location. The DS1 buoy is located in the eastern Arabian Sea region. The pre-monsoon warming and monsoonal cooling of SST over this region is well reproduced by the REY SST. The SST between REY and DS1 buoy observation is compared in Figure 3.3a. The SST comparison at DS2, DS3, DS4 and DS5, buoy locations are shown in figure 3.3b, 3.3c, 3.3d, and 3.3e respectively. The DS3 and DS4 buoys are located in the central Bay of Bengal region. REY SST is able to capture the seasonal evolution of SST at all the mooring locations. Bias in SST was observed based on the comparison of buoy SST with the TMI and Reynolds data. Most of the bias (both cold and warm bias) in REY and TMI SST are similar, meaning that either both of them showed warm bias or cold bias. This perhaps is due to the horizontal and temporal resolution of these products REY and TMI SST and due to their inefficiency in incorporating the rains in the algorithm. Also the inability of the microwave sensors on observing SST during rain is clearly seen in these comparisons. For example the rain fall around 25 May at DS5, the SST fell over 2°C in 2-3 days whereas both TMI and REY SST showed warming during this period. Interestingly this particular REY SST product has significant intraseasonal oscillations which are comparable with the TMI SST and this has been further exploited in the forthcoming section. Reynolds SST captures the intraseasonal variability as well as seasonal changes very closely.

3.2.3 Annual and seasonal SST mean

Figure 3.4 shows annual mean and seasonal mean (MAM, JJAS, ON, and DJF) of SST from 1998-2007 (a) TMI SST (b) REY SST MAM, JJAS, ON, and DJF (March to May, June to August, October to November and December to February respectively). SST variations have a predominant annual variation in the tropics and give insights into the
dynamics of the underlying oceanic processes. Both the data products are observed to be consistent with each other or show fairly good agreement with each other. In the mean annual cycle, extended warm pool with amplitude of SST of the order of 29°C was observed in the equatorial Indian Ocean. On the contrary, weaker SST of the order of 26°C was seen south of 18°S. This pattern was seen in the entire seasonal mean but with a lot of variability in its amplitude and its structure.

In the equatorial Indian Ocean, out of all seasonal mean, only during MAM maximum amplitude of SST was observed. An elliptical dome of less value of SST (less than 25°C) is seen all the seasons over the southern Indian Ocean.
Figure 3.4b: Annual mean and seasonal mean (MAM, JJAS, ON, and DJF) of REY SST from 1998-2007.

These cold SSTs are extended from south to north during the monsoon and post monsoon seasons due to the influence of the southwest monsoon. It is very important to note that in all the seasons there is a steep meridional gradient between south to north from 20°S to Equator. The SST variations reflect the combined fluctuations from dynamic and thermodynamic ocean processes. The zonal gradient in SST over the western and eastern equatorial Indian Ocean during JJAS is due to the strong monsoon winds induced upwelling in the west and mass convergence associated with equatorial jets in the east. In the Arabian Sea SST is controlled by the seasonally reversing monsoonal winds. The Arabian Sea shows weak magnitude in SST during the summer monsoon in the Somali coast. The value of maximum SST in the central Arabian Sea is seen to be during MAM. This variability in the Arabian Sea is attributed by the coastal upwelling (downwelling)
during summer (winter). The pre-monsoon warming of northern Indian Ocean and the formation of mini warm pool over the southeast Arabian Sea are well evident in these products. The areal extent of the mini warm pool region (SST >29.5°C) seems to be similar in both the products.

3.2.4 CEOF analysis of TMI and REY

CEOF analysis is also carried out with TMI and REY SST for further comparison and validation.

![First and second CEOF modes of the TMI and REY SST.](image)

Figure 3.5: First and second CEOF modes of the TMI and REY SST.
Figure 3.5 shows the first and second CEOF modes of the TMI and REY. The first CEOF mode of TMI and REY explains respectively 46.49% and 46.19% of the total variance. The second CEOF mode of TMI and REY explains respectively 23.19% and 18.94% of the total SST variance in the IO. The spatial patterns in both the products show similar pattern. First mode shows almost same % variance but in second mode little differences in the % variance has been observed. Amplitude, phase and variance derived from CEOF analysis for first and second modes for TMI and REY SST has also been compared (Figure not shown). All the three parameter (Amplitude, phase and variance) are observed to be comparable with each other and hence signifies the importance of daily

![Cumulative Variance Graph](image_url)

Figure 3.6: Cumulative variance (in %) of TMI and REY SST.

Figure 3.6 shows the cumulative percentage contributions of most significant modes to the total variance of TMI and REY SST. In both TMI and REY SST, the first few modes are found to explain maximum percentage of total variance. The rest of the modes occupied a very small percentage of total variance. Moreover number of modes to explain total variance in REY SST is observed to be comparable with TMI SST.
3.2.5 Intraseasonal variability in REY SST

To checkout the intraseasonal signal presented in the daily REY SST and TMI SST, SST data from both the products are filtered into 10–90 day, 30–90 day, and 10–30 day using Lanczoc band-pass filter (Ref.: Duchon, 1979).

![Figure 3.7](image)

Figure 3.7: Time series of 10–90 days timescales intraseasonal TMI (dotted line) and REY SST (dashed line) averaged over (a) BOX1, (b) BOX2, and (c) BOX3. Intraseasonal signals of 10 – 90 days are observed to be well captured by REY and TMI SST. Saji et al. (2006) observed the cooling event in BOX1 on 27 Jan 1999, 24 Mar 1999, 25 Feb 2000, 23 Jan 2001, 29 Nov 2001, 27 Jan 2002, 07 Feb 2003, which can be seen in the intraseasonal signals of 10 – 90 days in both the REY and
TMI products. Intraseasonal signal in TMI and REY SST in BOX1, BOX2 and BOX3 is also found to be comparable. However amplitude of intraseasonal signal using TMI was seen to be more than REY SST. It is also attempted to check 30-90 days and 10-30 days timescales signals in REY SST. Figure 3.8 and 3.9 show 30-90 day and 10-30 day intraseasonal signals respectively of REY SST and TMI SST. Both the products are found comparable to each other.

Figure 3.8: Time series of 30–90 days timescales intraseasonal TMI (dotted line) and REY SST (dashed line) averaged over (a) BOX1 (b) BOX2 (c) BOX3.
Figure 3.9: Time series of 10–30 days timescales intraseasonal TMI (dotted line) and REY SST (dashed line) averaged over (a) BOX1 (b) BOX2 (c) BOX3.

3.3 Intraseasonal variability of SST over the Indian Ocean

The tropical ocean affects the atmosphere via SST. Investigating intraseasonal SST variability induced by the intraseasonal oscillations (ISOs) is key to understand the coupled processes on intraseasonal timescales. ISOs of 10–90 day periods of the tropical Indian Ocean have been extensively studied using several sources of observational data. On 30–90 day timescales, ISOs is dominated by the Madden-Julian Oscillation (Madden and Julian, 1971, 1972, Schiller and Godfrey, 2003; Jones et al. 1998; Woolnough et al.
In the tropical Indian Ocean, the ISOs are enhanced due to the internal oceanic instability and the response to the intraseasonal atmospheric forcing (Waliser et al. 2003; Reppin et al. 1999; Sengupta et al. 2001b). Feng and Wijffels (2002) analyzed the satellite altimeter data and reported that the wavelength of the ISOs in southeastern Indian Ocean is from 600 to 900 km, the period is from 40 to 80 days, and the westward phase speed is from 15 to 19 cm s$^{-1}$.

In the past few years, a large number of modeling studies have significantly advanced our understanding of the physical processes that account for intraseasonal SST variability in the southern Indian Ocean. Harrison and Vecchi (2001) analyzed TMI SST and showed strong intraseasonal cooling over a large region of the southern equatorial Indian Ocean during boreal winter of 1999. They suggested that both air-sea heat flux and oceanic processes are important for the SST change. Empirical studies on the effects of the MJO have also been conducted using OLR and gridded reanalysis data sets which suggested that anomalous latent heat flux and surface insolation drive intraseasonal SST variations (Hendon and Glick, 1997; Jones et al. 1998; Shinoda et al. 1998; Woolnough et al. 2000). Schiller and Godfrey (2003) using an ocean general circulation model (OGCM), explained response of the Indian Ocean mixed layer and barrier layer (Lukas and Lindstrom, 1991; Sprintal and Tomczak, 1992) to atmospheric ISOs and how important these responses are in causing intraseasonal SST. Duvel et al. (2004) performed an OGCM experiment and diagnosed the processes that determine the large-amplitude cooling associated with two individual ISO events during northern winter: January and March 1999 and concluded that the thin surface mixed layer near 5°S–10°S is enhanced in 1999 because of enhanced precipitation and anomalous wind stress curl, producing the maximum cooling there. For this region, intraseasonal SST variability is mainly driven by intraseasonal surface fluxes rather than by advection or exchanges with the subsurface. Saji et al. (2006) analyzed the intraseasonal cooling of the south equatorial region on the basis of 8 years (1998–2005) satellite data, and suggested that reduced solar radiation, enhanced
evaporation and possibly strong entrainment all play a role, and local Ekman pumping may not be important. Han et al. (2006) analyzed the TMI SST and performed a series of OGCM experiments to isolate the effects of submonthly ISOs on Indian Ocean SST. They suggested that SST caused by submonthly ISOs can be as large as 0.5°C for strong events, and that winds are the primary force for the SST variability. From the previous studies it is understood that 1) realistic simulation of the ISOs is still a challenge for climate models (e.g., Slingo et al. 1996; Sperber et al. 2005; Lin et al. 2006). 2) The air-sea coupling on intraseasonal timescales can improve ISO phase and propagation which are important for both climate model simulation and prediction and this improvement is important for monsoon and ENSO prediction. 3) The relative importance of MJO and submonthly ISOs in causing Indian Ocean SST variability is not yet known.

The intraseasonally filtered (10–90 day, 30–90 day, and 10–30 day) SST is used to map the standard deviation in Figure 3.10. The STD map marks out the regions with high STD or variability over the south equatorial Indian Ocean region (60E–85E, 10S–3S; hereafter ITCZ region), western-central equatorial basin and southern East Indian Ocean. The ITCZ region is of prime importance in the context because in this region strong intraseasonal variability is observed during different seasons (MAM, JJAS, ON, and DJF).

![Figure 3.10: STD of 10–90 day, 30–90 day, and 10–30 day timescales intraseasonal SST for period 99-03.](image)
3.3.1 Impact of atmospheric ISOs on SST

To study the impact of atmospheric ISOs on the oceanic ISOs (intraseasonal variability in SST) the atmospheric ISOs over the tropical Indian Ocean are analyzed and compared with the oceanic ISO. Intraseasonal variability in wind speed, wind stress and wind stress curl fields for the period 10–90 day, 30–90 day, and 10–30 day were analysed.

Figure 3.11: Same as figure 3.10, but for wind speed.

Analysis of STD of wind speed and wind stress curl map also marks out the regions with high standard deviation or high variability over the ITCZ region (Figure 3.11 and 3.12). McCreary et al. (1993), Murtugudde et al. (1999) and Xie et al. (2002) observed that mixed layer is thin over the ITCZ region. Significant variation of wind speed and wind stress curl (Figure 3.11 and 3.12) and thin mixed layer lead the significant SST variability (Figure 3.10) over the ITCZ region. In the westerncentral equatorial ocean, wind speeds appear to be weaker but the amplitude of SST change, however, is larger. The most likely reason is that the atmosphere is more humid and closer to saturation because of its high SST (Han et al. 2007).
Apart from ITCZ region in the south East Indian Ocean, maximum SST variability associated with the maximum wind speed and wind stress curl was observed. SST variability over the ITCZ region in 30 – 90 day timescale was observed to be stronger than 10 -30 (submonthly) days. On the contrary, wind speed and wind stress curl variability was seen stronger in 10 -30 day timescale than 30 -90 days timescale. Even though the 10–30 day winds have larger variability (amplitudes) than 30–90 day winds, the latter can cause a larger SST variation because oscillating winds for a longer period allows the ocean to heat or cool for a longer time before the wind switches phase. Further the
response of SST to a local, periodic forcing is proportional to both the forcing strength and length of the forcing period (Han, 2005).

Intraseasonal wind stress curl affects the SST through various oceanic processes. Locally, positive (negative) Ekman pumping velocity associated with wind stress curl causes upwelling (downwelling) in the offequatorial regions (such as the south equatorial basin in the ITCZ region) which increases the surface cooling (warming). Intraseasonal wind stress curl can also drive strong intraseasonal currents, causing SST variations through advection. Remotely, Rossby waves forced by the wind stress associated with the ISOs propagate westward, causing variability of the thermocline and thus affect the intraseasonal SST. Additionally, intraseasonal wind stress curl can change mixed layer depth and thus affect intraseasonal SST (Han et al. 2007).

Strong intraseasonal variations of convection (Figure 3.13) can be seen in the region where mean SST exceeds 29°C (Figure not shown), and in southern tropical basin (10S–3S) where the ITCZ region is located. Infact it is found that apart from wind speed and wind stress curl, OLR forcing on 10 - 90 day, 30 – 90 day and 10 -30day scales also contribute to the intraseasonal variability in the SST. It emphasizes the role of atmospheric intraseasonal variability in producing the intraseasonal variability in the equatorial Indian Ocean SST.

To further investigate the role of atmospheric intraseasonal signal in generating SST intraseasonal variability, correlation between wind speed, wind stress curl and OLR with SST (Figure 3.14) has been computed. This analysis further emphasis the role of wind speed, wind stress curl and OLR in the SST intraseasonal variability. These results demonstrate that intraseasonal SST variability over the ITCZ region results largely from atmospheric ISO forcing.
Figure 3.14: Correlation between wind speed averaged over 60°E–85°E, 10°S–3°S with SST over the tropical Indian Ocean for 10–90 day, 30–90 day, and 10–30 day timescales intraseasonal signal for the period 99-03 (top panel), wind stress curl averaged over 60°E–85°E, 10°S–3°S with SST over the tropical Indian Ocean for 10–90 day, 30–90 day, and 10–30 day timescales intraseasonal signal for the period 99-03 (top middle panel) and OLR averaged over 60°E–85°E, 10°S–3°S with SST over the tropical Indian Ocean for 10–90 day, 30–90 day, and 10–30 day timescales intraseasonal signal for the period 99-03 (Bottom panel). Correlation obtained above 95% significance level.
Figure 3.15: STD of 10–90 day, 30–90 day, and 10–30 day timescales intraseasonal variability in SST during boreal spring (MAM), boreal summer (JJAS), boreal fall (ON) and boreal winter (DJF) for the period 99-03.
Intraseasonal variability in SST during boreal spring (MAM), boreal summer (JJAS), boreal fall (ON) and boreal winter (DJF) was computed separately to understand its seasonal variability (Figure 3.15). Maximum intraseasonal variability was observed during DJF, whereas very less intraseasonal variability was observed during JJAS. The intraseasonal variability during DJF seems to be dominant over all the seasons (MAM, JJAS and ON). The atmospheric intraseasonal oscillation (wind speed, wind stress curl and OLR) over the tropical Indian Ocean is also analyzed and compared with the oceanic ISO (Figure not shown). Maximum STD of wind speed, wind stress curl and OLR is also observed over the ITCZ region during DJF, whereas JJAS variability is observed to be very less. Figure 3.16 shows correlation between wind speed and SST during MAM, JJAS, ON, and DJF for intraseasonal signal of 10–90 day, 30–90 day, and 10–30 day timescales prepared in the same manner as Figure 3.14. This further confirm the role of atmospheric ISO on SST intraseasonal variability.

Strong correlation between wind speed and SST was observed in the ITCZ region during DJF, ON, MAM. But during JJAS the value of correlation was seen to be very small and this in fact supports the role of wind speed in the SST intraseasonal variability over the region. The correlation between wind speed and SST on 30-60 days timescales is much higher than submonthly timescales. The results indicate that on 30-60 days time scales intraseasonal variability in SST will be more than the submonthly time scales. Similarly, the wind stress curl also contributed to the intraseasonal variability in SST (Figure not shown). Figure 3.17 shows correlation between OLR and SST during MAM, JJAS, ON, and DJF for intraseasonal signals of 10–90 day, 30–90 day, and 10–30 day timescales, which prepared in the same manner as figure 3.16. The significant correlation was observed in the ITCZ region in all time scales between 10–90 day, 30–90 day, and 10–30 day intraseasonal timescales OLR and SST during DJF, ON, MAM manifesting the interactions between the intraseasonal convection and intraseasonal SST. However, during JJAS correlation seems to be very less. Thus maximum (minimum) SST variations over the ITCZ region during DJF (JJAS) are observed to be associated with the atmospheric ISOs. In other words atmospheric ISO can control the SST variations over the ITCZ region.
Figure 3.16: Correlation between wind speed averaged over 60E–85E, 10S–3S with SST over the tropical Indian Ocean for 10–90 day, 30–90 day, and 10–30 day timescales intraseasonal signal for the period 99-03 (top panel during boreal spring (MAM), boreal summer (JJAS), boreal fall (ON) and boreal winter (DJF).
Figure 3.17: same as 3.16 but for OLR.
3.4 Interannual variability in the Indian Ocean

Intraseasonal and interannual variability over the Indian Ocean are observed to be a very important phenomenon for the global climate system. In the last sections, the intraseasonal variability over the tropical Indian Ocean has been studied. Now in this section, the interannual variability over the tropical Indian Ocean is emphasized.

In order to study the interannual variability, CEOF analysis were carried out with T/P SSHA and HSA. The detail of CEOF analysis has been provided in Chapter 2.
Figure 3.18 shows the first and second CEOF modes of the SSHA and HSA. The first CEOF mode of T/P SSHA and HSA explains 30.13% and 43.12% respectively of the total variance. The second CEOF mode of T/P SSHA and HSA explains 21.66% and 22.48% respectively of the total sea-level variance in the tropical Indian Ocean. The spatial pattern of the second mode clearly showed dipole structure. Infact both the first and second modes of the HSA had shown more % variance than SSHA. Moreover phase also showed westward propagation in the tropical Indian Ocean. This is in agreement with the seesaw oscillation observed between the eastern and western equatorial Indian Ocean thermocline depths identified previously by Masumoto and Meyers (1998); Basu et al. (2000) and Murtugudde and Busalacchi, (1999). Moreover it is noticed that SSHA and HSA to the south of the equator in the western Indian Ocean is elongated southeastward and its intensity is stronger than its northern counterpart (Figure 3.18b). It also reveals that the significant amount of interannual variability is confined to the north of 15°S. Using the extended EOF analysis of T/P SSHA, Rao et al. (2002) illustrated how thermocline oscillations in the western Indian Ocean region take the form of Rossby waves propagating from the east. They referred second mode as a quasi-biannual mode. Therefore it can be concluded that IOD has been the leading mode of the interannual variability of the SSHA and upper ocean HSA.

Figure 3.19: Cumulative variance (in %) of T/P SSHA and HSA.
Figure 3.19 shows the cumulative percentage contributions of most significant modes to the total variance of T/P SSHA and HSA. In both T/P SSHA and HSA, the first three modes are found to explain maximum percentage of total variance. The rest of the modes occupied a very small percentage of total variance. Moreover number of modes to explain total variance in HSA is very less as compared to TP SSHA.

3.4.1 IOD structure in SSHA, HSA, HADISST anomalies and in D20 anomalies

Figure 3.20 shows HSA (in $10^8$ J/m$^2$) for 1994, 1996 and 1997. In October and November 1996 the positive HSA was of the order of $2 \times 10^8$ J/m$^2$ and $4 \times 10^8$ J/m$^2$ respectively in the region 10°S - Equator, 90-110°E (hereafter eastern IOD box) and low HSA in the region 10°S - 10°N, 50-70°E (hereafter western IOD box). Contrasting features were evident in October and November 1994 and 1997. It is clear from Fig. 3.20 that low HSA has been seen in eastern IOD box and high in the western IOD box. Further in 1994 peak cooling is seen in October (of the order of $-4 \times 10^8$ J/m$^2$) and this cooling extend has been found to cover the longitudinal distance from 110°E to 85°E of equatorial IO. But in 1997 peak value of cooling is found in November 1997 (of the order of $-15 \times 10^8$ J/m$^2$) and extended from 110°E to 68°E. As far as western IOD box is concerned, very high values of HSA were seen in October – November 1997 and October 1994. But 1997 values were higher than 1994, which support the fact that 1997 event is stronger than 1994. Infact it has been noticed from previous studies that the 1997 IOD event is intense and was associated with severe floods in eastern Africa and droughts over Indonesia. It was the strongest dipole recorded in history and was accompanied by massive fires in western Indonesia and the widespread death of coral reefs in the Indian Ocean.

Index to quantify the IOD has been defined (Saji et al. 1999) as the SST anomaly difference between the western IOD box and eastern IOD box. When the IOD index is positive (west is warmer than east), it leads to droughts (locally known as Tuarang) in the Indonesian region and heavy rains and floods in the east Africa. When the sign reverses, these anomalous fluctuations also swing to the opposite phase. Figure 3.21 shows (a) IOD Index for Heat storage anomalies ($10^8$ J/m$^2$) (b) IOD Index for Heat storage rates ($10^2$ W/m$^2$) (c) IOD Index for D20 anomalies (m) (d) IOD Index for HADISST
anomaly (degree C). It can be illustrated that calculated IOD index for (a), (b), (c) and (d) are well comparable with the IOD index calculated by Saji et al. (1999).

Figure 3.22 shows the T/P SSHA (cms), HADISST anomaly (degree C) and D20 during October /November 1994, 1996 and 1997. Dipole structure was clearly seen in T/P SSHA, HADISST anomaly and D20 anomalies and is comparable with the dipole structure found in HSA (Figure 3.20). Further Figure 3.22 clearly shows evidence of upwelling Rossby waves in T/P SSHA and D20 anomalies during 1994 and 1997 (which will be further explored in chapter 5). From Figure 3.22 it can be said that in October 1994 the upwelling Rossby waves are found extended from 110°E to 85°E. While in November 1997 these extended upto 68°E.

Figure 3.20: Heat storage anomalies (10^8 J/m^2) for 1994, 1996 and 1997.
Figure 3.21: (a) IOD Index for Heat storage anomalies ($10^8 \text{ J/m}^2$), (b) IOD Index for Heat storage rates ($10^2 \text{ W/m}^2$), (c) IOD Index for D20 anomalies (m), (d) IOD Index for HADISST anomaly (degree C).
From the above discussions it has been concluded that IOD is the leading mode of the interannual variability of the upper ocean SSHA and HSA. It is well known through the previous studies that the two features are common to T/P SSHA, at most latitudes: the
basin-scale non-propagating variability and the meso-scale to large-scale westward propagating signal. To understand better the complex ocean-atmosphere coupled system (e.g., IOD) one needs a good understanding of planetary waves — where they occur, their amplitude, and particularly their speed of propagation, which determines the timescale for ocean-atmosphere feedback. This has motivated the present study to filter SSHA into different wave’s components (e.g. Rossby and Kelvin waves) for understanding their behavior and role during in IOD. The nature and role of these waves during IOD has been given in the forthcoming chapters.

3.5 Summary

The new high resolution SST analysis products developed using OI has been used in this chapter for its validation. The analyses have a spatial grid resolution of 0.25° and temporal resolution of 1 day. SST from REY is validated using TMI SST and observations from buoys in the tropical Indian Ocean. The most encouraging result is that, for the first time a REY SST product has become available that captures the intraseasonal and lower frequency variability on time scales longer than a few days. Thus the REY SST can be a useful tool for studies of intraseasonal variability. In addition to the intraseasonal variability in SST, the intraseasonal variability in QuikSCAT winds, outgoing long wave radiation and their association with the oceanic intraseasonal variability are studied. The results demonstrate that intraseasonal SST variability over the region 65°E to 85°E, 10°S to 3°S results largely from intraseasonal variability in surface winds. Analysis of intraseasonal variability in SST during different seasons (MAM, JJAS, ON and DJF) shows maximum intraseasonal variability during DJF, whereas very less intraseasonal variability during JJAS. Maximum (minimum) variability of SST during DJF (JJAS) is influenced by atmospheric intraseasonal variability. Results of this chapter have important implications on air-sea interaction over the southern Indian Ocean, especially over the ITCZ region.

IOD has been shown to be a leading mode of the interannual variability of the upper ocean HSA and SSHA. The SSHA and HSA showed the existence of the dipole structure in the equatorial Indian Ocean in 1994 and 1997. The 1997 dipole mode structure
was observed to be stronger than 1994 and that can be clearly seen in HSA, T/P SSHA, thermocline depth (D20) anomaly derived from SODA and in HADISST anomaly. Dipole structure in SSHA motivated the present study to filter the SSHA into different components (e.g. Rossby and Kelvin waves) for understanding their behavior and role during in IOD.