4.1 Introduction

Equatorial planetary waves have been studied analytically through numerical modelling (O’Brien and Hurlburt 1974; Reverdin and Cane 1984; Han et al. 1999) and their effect on the coasts was studied by Clarke and Liu (1993, 1994) and Gnanaseelan et al. (2003). They also have been observed through moored (Luyten and Roemmich, 1982) and drifting buoys (Luyten et al. 1980; Reverdin et al. 1983). In Topex/Poseidon data combined with data from two earlier altimeters (ERS-1 and Geosat), Jacob, (1994) first saw evidence of a single Rossby wave lingering in the Pacific Ocean 10 years after their origin. Le Blanc and Boulanger, (2001) studied the interannual variability in the equatorial Rossby and Kelvin waves over the Indian Ocean using Topex/Poseidon altimetry data. Chelton and Schlax, (1996) mapped intra-seasonal signals in time-longitude sections of the entire world ocean, and identified the alternating anomalies unmistakably as first mode baroclinic Rossby waves. They objectively estimated the phase speed of each wave and found a surprising discrepancy between observed Rossby wave speeds and those predicted by the linear theory. A comparison of their phase speed estimates with those obtained from the linear theory shows observational values up to four times faster than the theory. The average difference in mid-latitudes is a factor of two. The advent of satellite altimetry made radial advances in the study of oceanic process such as IOD especially due to Rossby waves propagation. The wind anomaly and associated Ekman pumping generate off-equatorial Rossby waves that travel westward, deepen the thermocline, and warm the SST in the western Indian Ocean, causing the peak of an IOD event a few months after it begins (Feng and Meyers 2003). Prasad and McClean, (2004) found that internal oceanic
processes within the Arabian Sea are primarily responsible for western equatorial Indian Ocean warming. Chamber et al. (1999) did a rather extensive analysis of the Rossby waves observed with altimetry during the 1994 and 1997 events and clearly pointed out that the Rossby waves in 1994 and 1997 appeared before the IOD warming (although they linked it to ENSO, not IOD, as IOD had not been quantified at that time), and suggested that the Rossby waves had a dynamical relationship to the warming. They suggested that the warming begins with wind - forced Rossby waves in the southeastern Indian Ocean associated with the Southern oscillation, similar to the forcing of Kelvin waves which precede El Nino in the Pacific. However, in the present chapter detail description of Indian Ocean Rossby waves in connection with IOD has been carried out.

4.2 Interannual variability of semi-annual and annual Rossby waves propagation as twin gyres in the equatorial Indian Ocean

Before discussing the westward propagating Rossby waves over the equatorial Indian Ocean in detail, the Hovmoller diagram of the T/P SSHA (a) along equator (b) latitudinal average of SSHA along -2.5°S and 2.5°N and (c) latitudinal average of SSHA along 4.5°S and 4.5°N are analyzed to understand the propagation of SSHA signals (Figure 4.1). SSHA along the equator clearly shows the eastward propagation of Kelvin waves (Figure 4.1.a), but the latitudinal average of SSHA along -2.5°S and 2.5°N has revealed that the strength of the westward propagating Rossby waves and their variability are stronger than the eastward propagating Kelvin waves (Figure 4.1b). The similar picture with much more significance was seen in the latitudinal average of SSHA along 4.5°S and 4.5°N (Figure 4.1c). Infact from figure 4.1, westward propagation of SSHA with increasing slope can be clearly seen in the higher latitude in accordance to wave guide theory. The westward propagating Rossby waves have a broad spectrum dominated by semi-annual and annual signals whose amplitudes vary significantly in time. The equatorial jet has been thoroughly examined in previous studies.
Figure 4.1: Hovmoller diagram of the T/P SSHA (a) along equator (b) latitudinal average of SSHA along -2.5°S and 2.5°N and (c) latitudinal average of SSHA along 4.5°S and 4.5°N.
However, how reflected Rossby waves from the Sumatra coast interact with the incident equatorial jet has not been investigated in detail yet. Westward propagating twin gyres in the equatorial Indian Ocean were reported by Reddy et al. (2004) from the ocean current simulation using a simple reduced gravity model driven by observed daily NCEP winds. Reddy et al. (2004) could not find the twin gyres in the raw TP SSHA and pointed out in their manuscript that SSH variations do not precisely capture the twin gyre structure. Knowing that the T/P SSHA is the best source for observing such gyres, SSHA has been filtered using FIR Filter. In view of all previous studies on equatorial Indian Ocean dynamics, the present study is unique in observing the twin gyre pattern in the equatorial Indian Ocean. The propagation of equatorial Rossby waves is seen as twin gyres, with one gyre centered in the northern hemisphere and the other in the southern hemisphere. The propagation of equatorial semi-annual Rossby waves during 1994 and 1997 (positive IOD years) is shown in Figure 4.2. During 1994 and 1997, the twin gyres form by the end of May, one just north of the equator and the other south of it. The northern gyre in 1994 was centered along 4°N and the weaker southern gyre was centered along 5.5°S. In contrast to this, in 1997 the northern gyre was centered along 3.5°N and its southern counterpart was centered along 4.5°S. These semi-annual Rossby waves (or twin gyres) are observed to reach the western Indian Ocean during August/September. Figure 4.3 shows the semi-annual Rossby waves during the years following the above positive IOD years.

During this time, the gyres reach western Indian Ocean during October/November. During 1995, the northern gyre was seen propagating with its center along 4°N. The southern gyre was centered along 4.5°S throughout the year 1995. In 1998 the northern gyre was found along 2.5°N and southern gyre along 4.5°S, but the northern gyre has strengthened near 60°E. The propagation of annual Rossby waves during the positive dipole years of 1994 and 1997 is shown in Figure 4.4. The southern gyres in 1994 and 1997 are dominant and centered along 6.5°S and the weaker northern gyre was centered along 2.5°N. During 1997, the strong westward propagating annual Rossby waves were observed in the eastern boundary around June and they reached the western boundary in February/March 1998.
Figure 4.2: Propagation of semi-annual Rossby waves (mm) 1994(left panel) and 1997(right panel).
Figure 4.3: Propagation of semi-annual Rossby waves (mm) 1995(left panel) and 1998(right panel).
Whereas the semi-annual signals reached the western boundary in October/November 1997. The relatively weaker annual signals of 1994 reached the western boundary in February/March 1995. In contrast, Figure 4.5 shows annual Rossby waves in 1995 and 1998 with the northern gyre centered along 2.5°N and the southern gyre along 3.5°S. The Rossby wave propagation as twin gyres was observed not only in the IOD years, but was observed in all the years during the study period, of course with considerable interannual variability in both intensity and latitudinal extent.

Time series of semi-annual and annual Rossby waves signals at different points in the equatorial Indian Ocean were analyzed to understand the variability in speed and magnitude during 1994 and 1997. Figure 4.6 shows the time series of semi-annual Rossby waves [(a) top three panel] and annual Rossby waves [(b) bottom three panel] along 90°E (line), 70°E (open circle) and 50°E (cross), along 0.5°N, 2.5°N and 2.5°S. Figure 4.6 strongly confirms that the linear theory will fail to explain the sort of changes happening in the planetary wave propagation in the equatorial Indian Ocean. The standard linear theory of these planetary waves is well known (Dickinson, 1978, Leblond and Mysak, 1978, Gill 1982) and holds for an ocean whose background state is at rest, with uniform depth. Results from T/P altimetric observations (Le Traon and Minister, 1993; Glazman et al. 1996; Cipollini et al. 1997) revealed discrepancies between observed and theoretical Rossby waves phase speeds away from the tropics. These observations show that the standard linear theory is an incomplete explanation of these recent satellite measurements. Killworth et al. (1997) suggested that the linear theory may be inadequate because (1) flow is forced (by wind and/or buoyancy) and is part of the coupled system, and so it is not free; (2) the ocean has varying buoyancy; (3) the background state of the ocean is not at rest; and (4) the response is fundamentally non-linear.

Using altimetric data, Chambers et al. (1999) pointed out that during the 1994 and 1997 events; the waves are asymmetric about the equator and could be explained as the combination of first and second mode Rossby waves. The superposition of such waves had caused higher amplitude south of the equator and smaller amplitude north of the equator.
Figure 4.4: Propagation of annual Rossby waves (mm) 1994(left panel) and 1997(right panel).
Figure 4.5: Propagation of annual Rossby waves (mm) 1995(left panel) and 1998(right panel).
Figure 4.6: Time series of semi-annual Rossby waves [(a) top three panel] and annual Rossby waves [(b) bottom three panel] at lon. 90°E (line), 70°E (open circle) and 50°E (cross) on lat. 0.5°N, 2.5°N and 2.5°S.
The semi-annual Rossby waves found to be more energetic along the latitudes 2.5°N, 4.5°N, 4.5°S whereas the annual Rossby waves are energetic along the latitudes 2.5°N, 3.5°S, and 6.5°S. In all these cases the tropical Rossby waves have a local minimum along equator with symmetric maxima between 4.5° N and 4.5°S, in agreement with the previous studies (Delcroix et al. 1991; Chelton and Schlax 1996). But during the IOD years Rossby waves are found to have a pattern different from the normal years. None of the previous studies looked into these interannual variabilities. The phase speeds of semi-annual and annual Rossby waves were estimated using the Radon Transform method (Polito et al. 2000). For semi-annual Rossby waves the phase speed along 2.5°N is -41 cm/sec, 4.5°N is -49 cm/sec and 4.5°S is -63 cm/sec. For annual Rossby waves along 2.5°N, 3.5°S and 6.5°S, the phase speeds are observed to be -33 cm/sec, -37 cm/sec and -24 cm/sec respectively. Identification of Rossby waves as equatorial twin gyres in the SSHA will add a new insight into the Indian Ocean dynamics. One possibility for the genesis of the twin gyres is the nonlinear interaction between the equatorial jet and the first mode Rossby waves at the front of the reflected packet (Reddy et al. 2004). In this view, it is the strong retarding effect of the equatorial jet that allows a soliton-like twin gyres structure to develop at the wave front. Another possibility is that the gyres are Rossby wave solitons (Boyd, 1980), and the soliton theory involves only the Rossby waves balancing non-linearity against depression.

4.2.1 Drag or slowing down of westward propagating twin gyres

In the previous section an understanding of Rossby waves: where they occur, their amplitude, and particularly their phase speed etc. are discussed. In this section, an important issue about the propagation of annual Rossby waves during the IOD year has been discussed in detail. Reddy et al. (2004) using reduced gravity model currents found that twin gyres develop every year from 1992-2001 with less interannual variability. The present study however investigates the interannual variability in the Rossby waves in the equatorial Indian Ocean during the positive IOD years 1994 and 1997-98 and found a strong interannual variability in the twin gyres. The gyres that formed in 1994 and 1997 reached the western boundary in February/March 1995 and 1998 respectively (Figure 4.4).
In contrast, gyres formed in 1995 and 1998, years following positive IOD, (Figure 4.5) propagates relatively faster and reached the western boundary in October/November of the same year. We examined the propagation of annual Rossby waves in the rest of the non IOD years during the study period and found faster propagation similar to those obtained in 1995 and 1998 (Figure not shown). From the detailed analysis of the annual Rossby wave signals during the IOD years, a drag in the propagation of annual Rossby waves was observed in the region 78E-88°E. The average annual Rossby wave phase speed during the dipole years was approximately -29 cm/sec along 3.5°S whereas in the other years it was approximately -37 cm/sec. The drag in the propagation of annual Rossby waves during the positive IOD years and the resultant late arrival at the western boundary helped to strengthen the semi-annual Kelvin waves in the subsequent years (Figure 4.7). It is important to note that all these Rossby and Kelvin waves are derived from the T/P SSHA using the FIR filter. Figure 4.7 shows longitude-time plot of semi-annual Kelvin waves (mm) averaged over 3.5°S to 3.5°N, cycles (a) 1-120 (b) 120-240 (c) 240-373. It is suggested that the late arrival of downwelling Rossby waves to western Indian Ocean during the IOD periods coincided with the sustained 1997 warming in the western Indian Ocean upto February 1998 (Murtugudde et al. 2000).

4.2.2 Role of wind forcing in dragging annual Rossby waves

Rossby wave characteristics (amplitude and propagation speed) can be altered by wind or buoyancy forcing that is phase-locked with the wave signal (White et al. 1977). Figure 4.8 (a) shows area averaged wind stress curl anomaly over 78°E-88°E, 6.5°S-4.5°N and Figure 4.8 (b) shows zonal wind stress anomaly, meridional wind stress anomaly, wind stress anomaly averaged over 78°E-88°E, 6.5°S-4.5°N. The anomalous wind stress and wind stress curl during the dipole years seem to have caused the drag in the propagation which was observed in the annual Rossby waves in the region 78°E-88°E. The variation of the wind stress and wind stress curl makes the thermocline shallower, which therefore reduces the phase speed of the Rossby waves. Wind stress over the Indian Ocean was computed from NCEP/NCAR winds.
Figure 4.7: Longitude-Time plot of semi-annual Kelvin waves (mm) averaged over 3.5°S to 3.5°N; Cycles (a) 1-120 (b) 120-240 (c) 240-373.
Figure 4.8: (a) wind stress curl anomaly (*10^8 dynes/cm^2) averaged over 78°E-88°E, 6.5°S-4.5°N (dashed line) and (b) zonal wind stress anomaly (dynes/cm^2), meridional wind stress anomaly (dynes/cm^2), wind stress anomaly (dynes/cm^2) averaged over 78°E-88°E, 6.5°S-4.5°N.

4.2.3 Impact of slowing down of twin gyres on equatorial Indian Ocean

Figure 4.9 shows (a) OISSTA over Box1 (78°E-88°E, 6.5°S-4.5°N) and Box2 (88°E-98°E, 6.5°S-4.5°N), (b) TPSSHA over Box1 and Box2 (c) Kinetic energy (cm^2/s^2) over Box1 and Box2 (d) Heat storage (10^8 J/m^2) over Box1 and Box2. The difference in these values in the boxes is usually negligible during the normal years, whereas considerable difference is observed in the dipole years. It is observed that the drag (or slow down) in the Rossby waves during the dipole years in the region 78E-88°E gives more energy (of the order of 4 m^2/s^2) to the ocean which in fact changed the SST (increase of 1°C) and SSHA (increase of 10 cm) in this region (Figure 4.9). Heat storage anomalies
calculated over Box1 and Box2 showed maximum difference (of the order of $5 \times 10^8$ J/m$^2$) in the dipole years [Figure 4.9d]. The heat storage computation has more relevance with the strong positive correlation between rise in SSHA and the ocean warming. The heat storage anomalies in Figure 4.9d are calculated similar to Chamber et al. (1997) and details of which have already been mentioned in chapter 2. The calculation assumes that the height signal is dominated by temperature so that the haline contraction can be neglected, according to the results of Sato et al. (2000). This has been investigated by comparing the heat storage anomaly estimated from T/P SSHA and OISST anomalies in the peak IOD months of October 1994 and November 1997 [Figure not shown].

### 4.3 Annual Rossby waves in the Arabian Sea and its interannual variability

Apart from annual Rossby waves propagation in the equatorial Indian Ocean, the present study also attempted to understand the Rossby waves propagation in the Arabian Sea. In the context of observing anomalous Rossby waves and Kelvin waves in the equatorial Indian Ocean it is quite natural to assume that these variability will affect the coastal Kelvin waves along the rim of Bay of Bengal and along the coast of India. Further these might impact variability in the Arabian Sea Rossby waves. So it is felt that a brief discussion on these waves is necessary to have completeness. Westward propagation of SSHA over the Arabian Sea has been identified as twin eddies. These twin eddies have not explicitly stated (or have not been reported formally yet) in any of the previous studies even though the pattern is seen in McCreary et al. (1993) and Shankar and Shetye (1997).

The Arabian Sea is bounded by the east Africa and the Arabian Peninsula to the west, Iran and Pakistan to the north and the Indian Peninsula to the east. The distribution and the topography of the land around the Arabian Sea and the differential heating of the land result in distinct seasonal wind patterns and ocean circulation over this part of the region making it almost unique. Among all the circulation in the Arabian sea, Great Whirl, an anticyclonic gyre, with a high in the surface topography, that form between $5^\circ$N to $10^\circ$N during the onset of southwest monsoon in June has been well studied (Bruce, 1968). The occurrence of another such high in the eastern Arabian Sea, though not as energetic, was discovered by Bruce et al. (1994) using hydrographic data collected during IIOE, they showed that high [called as Lakshadweep high (LH)] forms off southwest India during the
north east monsoon and low [Lakshadweep low (LL)] forms in the same region during the southwest monsoon.

Bruce et al. (1994) noted that the LH appears to be the consequence of Rossby waves radiation by coastal Kelvin waves off southwest India. They also speculated another possible mechanism for the LH, a patch of anticyclonic wind stress curl off southwest coast of India during the northeast monsoon; and wind stress curl is cyclonic in this region during the southwest monsoon. Though the occurrence of the LH and LL off southwest India was established using traditional hydrographic data, its annual cycle vis a vis the annual cycle of the sea level perturbations in the rest of the Indian Ocean was appreciated better only after data from satellite altimetry become available (Tapley et al. 1994; McCreary et al. 1993). Shankar and Shetye (1997) did extensive study on these high and low. They investigated the dynamics of the high and low with an analytic model and numerical simulations using a reduced-gravity ocean model for the north Indian Ocean and concluded that the LH and LL do not owe their existence to nonlinearity. Moreover they also suggested that the formation of high (low) off southwest India during northeast (southwest) monsoon is one manifestation of an annual cycle of events that are linked not only to the coastal currents around India, but also to the circulation in the southern Arabian Sea as a whole. Further the processes separating the contribution of remote and local forcing in the formation of LH and LL is well documented in Shankar et al. (2002).

Current system along the west coast of India has been studied in detail in the previous studies but how it interact with westward propagating Rossby waves in the Arabian Sea has received less attention in the literature and deserves more discussion here.

Figure 4.10 illustrates the annual Rossby waves in the Arabian Sea during 1993. The propagation of annual Rossby waves can be seen as twin eddies; one centered in the north of 12.5°N [i.e. 13.5°N to 15.5°N] and other in the south of 12.5°N [i.e. 9.5°N to 11.5°N]. The propagation of twin eddies, i.e. their generation to decay during 1993 may be seen from Figure 4.10. These eddies were established in March and continued their westward propagation through out the period and dissipated by December. They were seen symmetric about 12.5°N. Positive values indicate downwelling Rossby waves and negative values indicate upwelling Rossby waves. Following the downwelling Rossby waves
(during the northeast monsoon), upwelling Rossby waves (during the southwest monsoon) were also observed. The signature of twin eddies was also seen in the original T/P SSHA. The twin eddies are evident in every year from 1993-2001 during both northeast (downwelling) and southwest monsoon (upwelling). The possible mechanism accounting for genesis of twin eddies system can be considered as the circulation system along the coast of India. In December, the East India Coastal Current (EICC) is equatorward; it turn around Sri Lanka and flows as a poleward West India Coastal Current (WICC). In January, a high forms off southwest India. The westward northeast monsoon current flows around the high, joining the west coast around 12°N. The EICC flows into the northeast monsoon current, which extends into the eastern bay. The WICC is poleward north of 12°N and equatorward to the south; the equatorward coastal flow implies coastal upwelling.

The high then stretches westward, and the northeast monsoon current and the high extends across the southern Arabian Sea by March. By May, the WICC is equatorward all along the coast; it flows along the eastward equatorial jet, which bifurcates near the eastern boundary of the basin. A low forms off southwest Indian by July. The WICC is equatorward to the north of 12°N, the current flowing on along the western slope of the low and the joining the eastward southwest monsoon current (also called Indian Monsoon current). South of 12°N, the WICC is poleward, implying coastal downwelling. The low too stretches westward, extending across the southern Arabian Sea by November. This current system makes 12°N being the critical latitude separating these twin eddies. Other genesis of twin eddies may be considered as manifestation of the decrease in Rossby wave speed with latitude. Rossby wave speed (on an equatorial beta plane) is inversely proportional to square of the distance from the equator. The rapid decrease in speed leads to the LH and LL off southwest India radiating off much faster from the Indian west coast. Hence, when the high (low) forms off southwest India, the low (high) generated earlier is still seen propagating westward farther north along the coast. The rate of delay of speed leads to ~ 12°N being the critical latitude separating these twin eddies. This (1/y^2) variation with latitude is what leads to the Rossby wave propagating much faster off southwest India than farther north along the west coast. It is this that leads to the Lakshadweep high (and low) in the southeastern Arabian Sea. These prompt to form a twin
Figure 4.9: (a) OISSTA over Box1 (78°E-88°E, 6.5°S-4.5°N) and Box2 (88°E-98°E, 6.5°S-4.5°N), (b) TPSSHA over Box1 and Box2 (c) Kinetic energy (cm²/s²) over Box1 and Box2 (d) Heat storage (10⁸ J/m²) over Box1 and Box2.
Figure 4.10: Annual Rossby waves as twin eddies in the Arabian Sea during 1993.
eddies circulation in this region. These study has been extended for long term series SSHA data from SODA. Filtered annual Rossby waves from long term series SODA SSHA from 1958-2001), also showed the presence of twin gyres through out the period (Figure not shown). The periodic occurrences of these gyres suggest that the ocean is not wholly chaotic and unpredictable, as generally believed.

4.3.1 Interannual variability of Rossby waves over the Arabian Sea

To understand the interannual variability within the annual Rossby waves in the Arabian Sea, the annual Rossby waves were averaged over the latitudes where these waves are most energetic. This procedure enhances the signals that are symmetric with respect to 12.5°N. Figure 4.11 shows the longitude-time plot of annual Rossby wave (a) averaged over 9.5°N to 11.5°N (b) 12.5°N (c) averaged over 13.5°N to 15.5°N. Interannual variability is evident in (a), (b) and (c). During the positive IOD years (i.e. 1994 and 1997), weak annual Rossby wave propagation was seen along 12.5°N and in the southern eddy. Figure 4.12 illustrates the longitude-time plot of wind stress curl anomalies (a) averaged over 9.5°N to 11.5°N (b) 12.5°N (c) averaged over 13.5°N to 15.5°N. The weak propagation of annual Rossby waves along 12.5°N and in the southern eddy during 1994-1995 and 1997-1998 are caused by the wind stress curl anomalies during these periods. Moreover wind stress anomalies during 1994-1995 and 1997-1998 were also observed to play an important role in weakening of annual Rossby waves along 12.5°N and in the southern eddy (Figure not shown).
Figure 4.11: Longitude-Time plot of annual Rossby waves (mm) (a) average over 9.5°N to 11.5°N (b) 12.5°N (c) average over 13.5°N to 15.5°N.
Figure 4.12: Longitude-Time plot of wind stress curl anomalies (*10⁶ dynes/cm²) (a) average over 9.5°N to 11.5°N (b) 12.5°N (c) average over 13.5°N to 15.5°N.
4.4 Summary

T/P SSHA have been used to study the semi-annual and annual Rossby wave propagation in the Indian Ocean. The propagation of equatorial semi-annual and annual Rossby waves is observed as twin gyres with minima along equator and maxima in either side of the equator. This study provides first observational evidence for the existence of twin gyres from T/P SSHA in the equatorial Indian Ocean. The formation of twin gyres is due to the nonlinear interaction between the equatorial jet and the first mode Rossby waves at the front of the reflected packet. During the IOD years of 1994/1995 and 1997/1998, the annual Rossby waves propagate slower (than normal) and reach the western boundary three to four months late. The average annual Rossby wave phase speed during the dipole years is approximately -29 cm/s along 3.5°S whereas in other years it is approximately -37 cm/s. A drag in the propagation of annual Rossby waves is observed in the 78°E-88°E during the IOD years. The anomalous wind stress and its curl seem to play an important role in dragging the propagation of annual Rossby waves in the 78°-88°E and hence the annual Rossby waves arrive the western Indian Ocean late in February/March during the IOD years. The drag in the Rossby waves during the dipole years in this region 78°E-88°E gives more energy to the ocean, which in fact changed the SST and SSHA in the region.

The westward propagating annual Rossby waves over the Arabian Sea were found as twin eddies, one centered in the north of 12.5°N (i.e. 13.5°N to 15.5°N) and the other in the south of 12.5°N (i.e. 9.5°N to 11.5°N), with considerable amount of interannual variability. In normal years, the annual Rossby wave along 12.5°N and the southern eddy were seen to have considerable amplitude but during the positive IOD years of 1994/1995 and 1997/1998, they were weaker. Wind stress and wind stress curl anomalies were seen to play an important role in the weakening of annual Rossby waves along 12.5°N and the southern eddy. The existence of these eddies has also been validated using SODA SSHA during 1958-2001.