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# West India Coastal Current and Lakshadweep High/Low

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**Abstract.** The West India Coastal Current (WICC) flows northward during November–February and southward during April–September. At the time of formation of the northward (southward) phase of the current, a high (low) in sea level – the Lakshadweep High (Low), forms off southwestern India, and migrates westward across the Arabian Sea. The annual cycle of the WICC and that of the Lakshadweep High/Low arise from a set of circumstances that are special to the North Indian Ocean. This relatively small tropical basin is driven by seasonal monsoon winds. As a result, its wind-driven near-surface circulation consists primarily of annual and semi-annual long, equatorially-trapped Kelvin and Rossby waves, and coastally-trapped Kelvin waves. In terms of these waves, the West India Coastal Current is a superposition of annual and semiannual coastally-trapped Kelvin waves. The Lakshadweep High/Low forms when the Kelvin waves, on turning around Sri Lanka, and propagating northward along the west coast of India, radiate Rossby waves.

**Keywords.** West India Coastal Current; Lakshadweep Sea; coastal dynamics; Kelvin waves; Rossby waves.

## 1. Introduction

The North Indian Ocean exhibits some striking differences when compared to the North/South Pacific/Atlantic Ocean basins. The North Indian basin is essentially a tropical basin. The Asian continent confines it equatorward of about 25° N. The other basins almost touch the polar circles. The North Indian basin is relatively small; along the equator it measures 6500 km, compared to the Pacific's 17,000 km. The most important distinguishing feature of the North Indian Ocean, however, is the winds it experiences – the Asian monsoons. The near-surface circulation that these winds drive is distinctly seasonal. All these special characteristics of the basin – small size, proximity to the equator, seasonality of circulation – lead to features of near-surface flow that are special to the North Indian Ocean. The purpose of this article is to describe one such feature, and to identify the elements of geophysical fluid mechanics that make its occurrence possible.

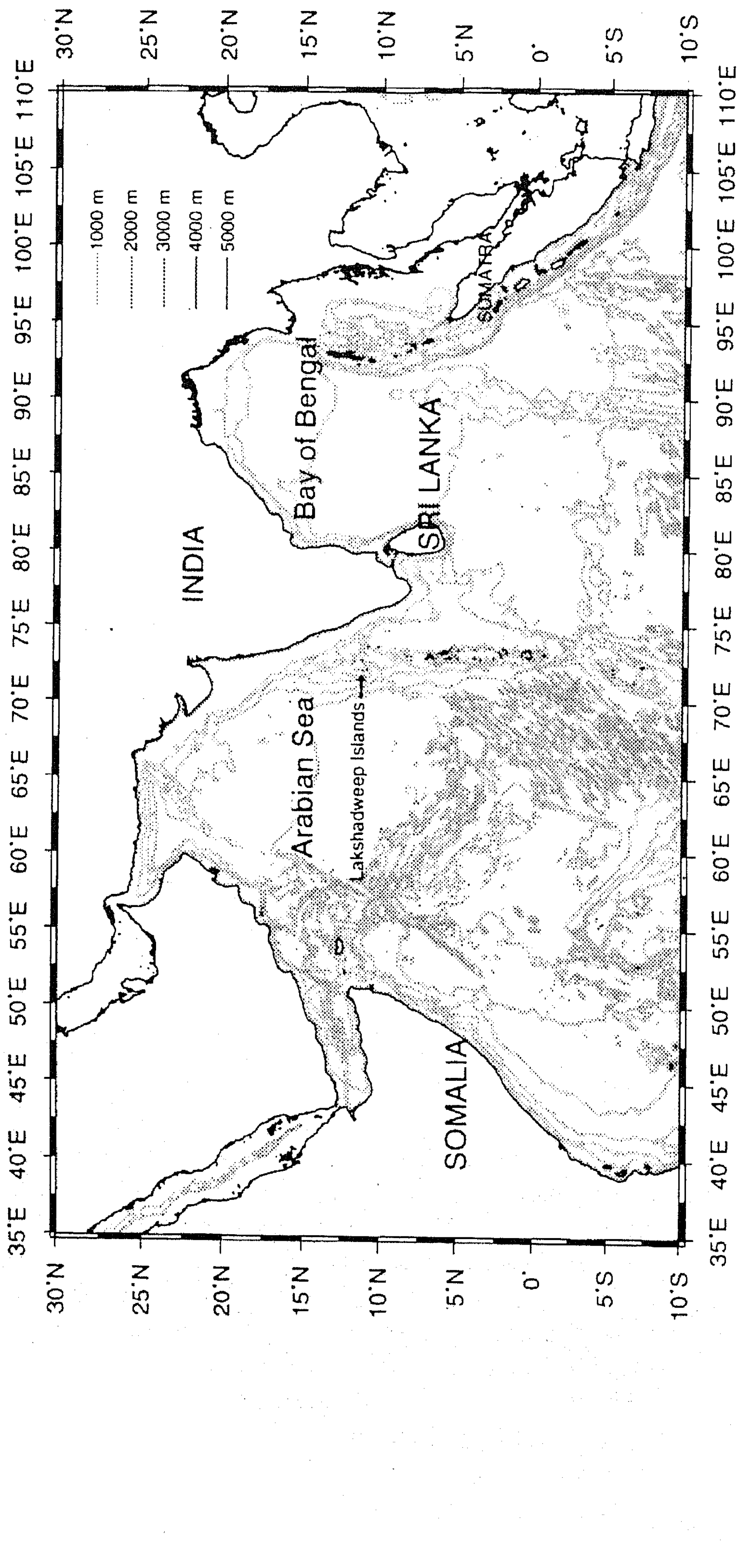


**Figure 1.** Map of the North Indian Ocean.

The feature consists of an anomaly in sea-level that builds up and decays periodically off southwestern India, in the vicinity of the Lakshadweep Islands (figure 1). The anomaly can be a "high", i.e. the sea-level gets higher, or a "low." We refer to it as the Lakshadweep High/Low (LH/LL). When it forms, there are events that take place simultaneously along the coast of India, in the form of changes in the coastal current, the West India Coastal Current (WICC).

In the next section we describe the annual cycle of events that consists of the waxing and waning of the LH/LL and reversals in the WICC. To appreciate the dynamics underlying the cycle, it is necessary to identify the agents of temporal evolution of near-surface, wind-driven circulation in the North Indian Ocean. This is done in § 3. In § 4 we discuss the dynamics underlying the LH/LL and the WICC. The results are summarized in § 5.

## 2. WICC and Lakshadweep High/Low

The WICC reverses seasonally. It is essentially a surface flow, the velocity being restricted to the uppermost few hundred metres. The core of the current is on the continental slope, i.e. close to the 1000 m isobath in figure 1. The current flows equatorward when the southwest monsoon (April–September) winds are active, and is best developed when the winds peak during July–August. At this time the current is strongest along the southwest coast of India, weaker in the middle part of the coast, and is hardly noticeable off the northwest coast. Transport of the current has been estimated to be about $4 \times 10^6 m^3 s^{-1}$ near the southwestern part of the coast (Shetye *et al* 1990).

The poleward phase of the WICC occurs during the northeast monsoon (November–February). This phase sets in when the equatorward current at the east coast of India, the East India Coastal Current (EICC), turns around Sri Lanka and starts flowing poleward along the west coast of India against the weak winds of this season. The current is about 40 km wide jet-like flow along the continental slope off the northwest coast of India, where the transport has been estimated to be $7 \times 10^6 m^3 s^{-1}$; the current is wider in the south (Shetye *et al* 1991). Thus, the poleward phase of the WICC is better developed than the equatorward phase, even though the winds during the northeast monsoon are weaker.

Associated with the seasonal cycle of WICC, there are seasonal changes that occur offshore. Sea level starts rising at the southwest coast of India with the formation of the poleward phase of the WICC. By early January, positive anomalies of sea level spread offshore and northward along the west coast, and a circular "high" in sea level forms to the east of the Lakshadweep islands. By February, the high is no longer circular: it has stretched westward. By April, positive anomalies in sea level are seen all over the Arabian Sea.

The sea-level high and the circulation associated with it were discovered by Bruce *et al* (1994). Though historical hydrographic profiles played an important role in this discovery, it is the more recent altimeter data that revealed the annual cycle of evolution of the Lakshadweep High much more succinctly, and showed the transition from the high during the northeast monsoon to the low during the southwest monsoon.

Evolution of the Lakshadweep Low proceeds in a fashion similar to that of the High. First, sea level drops at the southwest coast of India in June, when the WICC is equatorward. Second, the negative anomalies spread offshore and along the coast, and a circular low

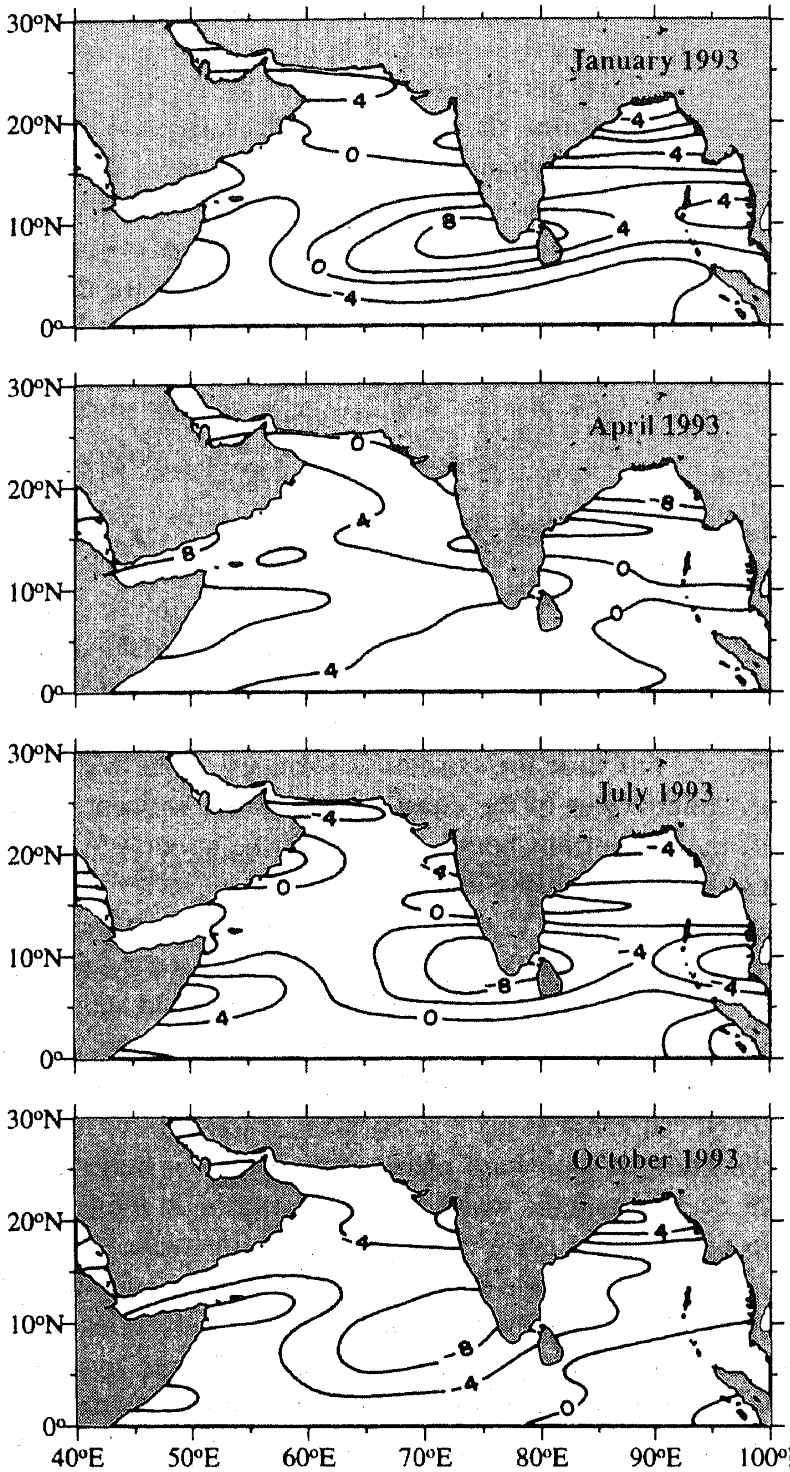


**Figure 2.** Sea level deviation for 1993 from TOPEX/POSEIDON. The mean level was defined over 1993–1995. The deviations, computed along the TOPEX/POSEIDON tracks, were averaged in $4° x 1°$ (longitude by latitude) cells to construct regular monthly grids of sea level change. Taken from Shankar & Shetye (1997).

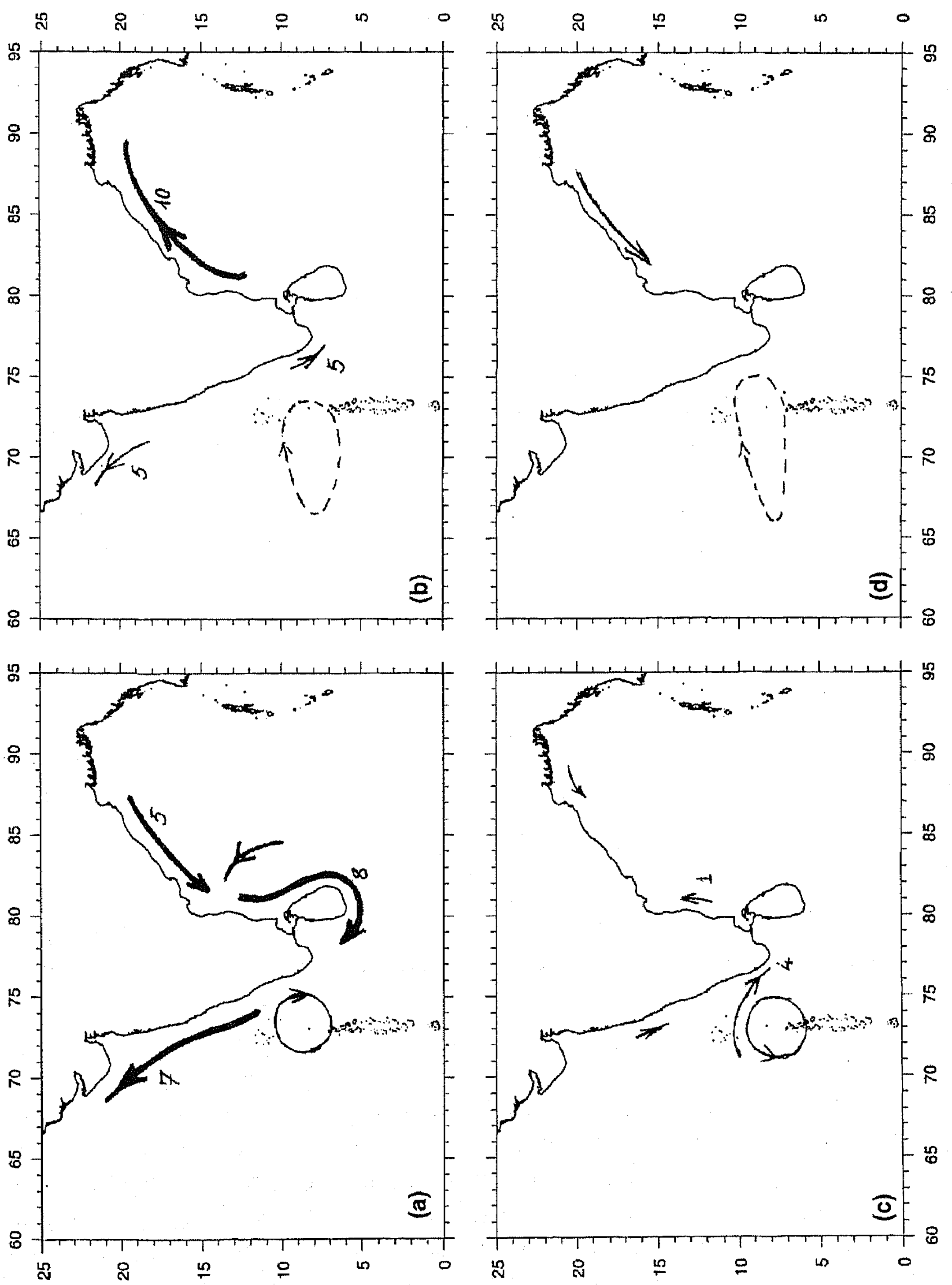


**Figure 3.** Schematic showing the characteristics of circulation in the vicinity of the coastal region of India. (a) January, (b) April, (c) July, (d) October. Numbers next to arrows showing patterns of current give estimates of transport ($10^6$ $m^3$ $s^{-1}$) of the current.

forms to the east of the Lakshadweep islands in July-August. Third, the low stretches westward, and by the end of October, negative anomalies in sea-level occur all over the Arabian Sea. Figure 2 shows the large-scale variability of sea level in the North Indian Ocean as revealed by altimeter data.

The schematic in figure 3 summarizes the events that take place along the coast of India and in the region offshore of the coast. What are the factors that lead to the formation of the patterns of circulation seen in the figure? How are these events linked to the annual cycle of WICC and LH/LL? Shankar and Shetye (1997) have argued that this cycle can be understood in terms of waves that arise in the equatorial region. We examine these waves in the next section

## 3. Long waves of the equatorial and coastal waveguides

During the last few years there have been important advances in our understanding of the working of large-scale seasonal currents in the North Indian Ocean in general, and the coastal currents around India in particular. This was made possible by a number of studies (Potemra *et al* 1991; Yu *et al* 1991; McCreary *et al* 1993, 1996; Shankar *et al* 1996; Vinaychandran *et al* 1996, and others). Of these, the work of McCreary *et al* (1993) was the most significant in identifying the following as the most fundamental elements.

(1) The coastal currents around India form an integral part of the circulation that encompasses the entire North Indian Ocean basin. Hence to understand the coastal currents, it is not sufficient to look at the coast alone, but it is necessary to consider the entire basin.

(2) The currents in the basin are primarily due to free and forced long waves of three kinds: equatorially-trapped Rossby and Kelvin waves, and, coastally-trapped Kelvin waves.

(3) The waves are triggered by the monsoon winds.

In order to understand the characteristics of the waves that are generated, it is important to understand the nature of wind stress, $\tau$, associated with the monsoons. $\tau$ is given by

$$\tau = \rho_{air} C_d \mathbf{U} U, \tag{1}$$

where $\rho_{air}$, $C_d$, $\mathbf{U}$ are the density of air, drag-coefficient, and wind velocity at a height of 10 m from the ocean surface. Figure 4 shows the wind stress during January and July. During the former, the northeast monsoon winds are near their peak, whereas during the latter the southwest monsoon winds are close to maximal. It turns out that the annual cycle of the monthly-mean wind stress over the North Indian Ocean can be represented to a good approximation as a linear combination of mean wind stress, an annual component, and a semi-annual component (Rao 1998), i.e.

$$\tau_x(\mathbf{r}, t) = \tau_x^{\circ}(\mathbf{r}) + A_x^{a}(\mathbf{r}) \sin\{\omega_a t + \phi_a^x(\mathbf{r})\} + A_x^{sa}(\mathbf{r}) \sin\{\omega_{sa} t + \phi_{sa}^x(\mathbf{r})\}, \tag{2}$$

$$\tau_y(\mathbf{r}, t) = \tau_y^{\circ}(\mathbf{r}) + A_y^{a}(\mathbf{r}) \sin\{\omega_a t + \phi_a^y(\mathbf{r})\} + A_y^{sa}(\mathbf{r}) \sin\{\omega_{sa} t + \phi_{sa}^y(\mathbf{r})\}, \tag{3}$$

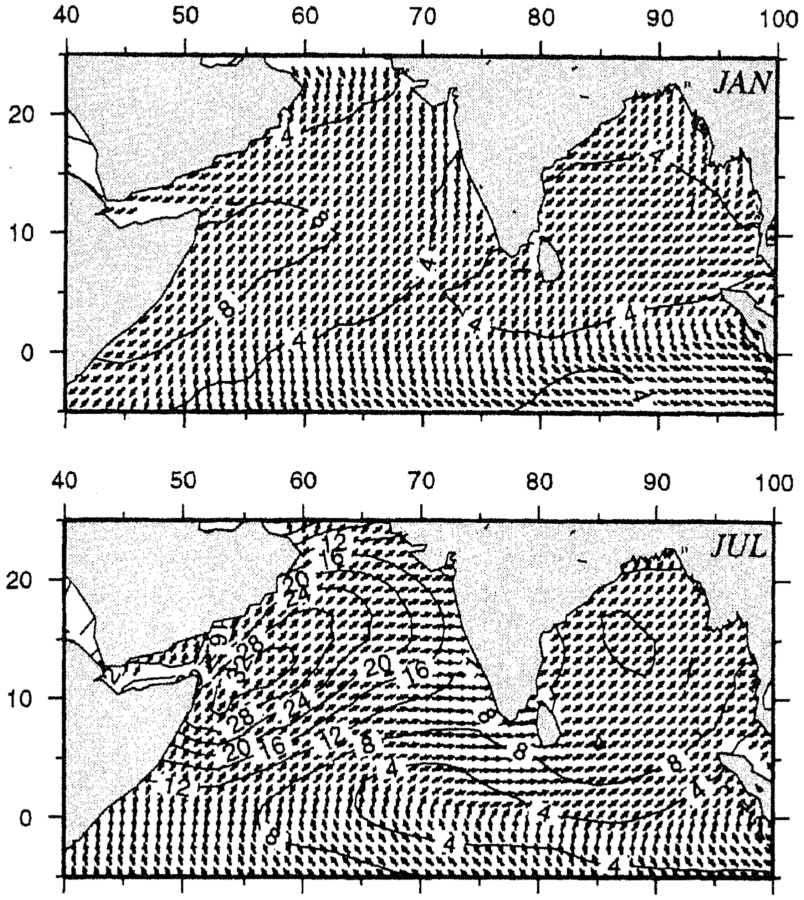


**Figure 4.** Wind stress ($10^{-1}$ dyn cm$^{-2}$) in the North Indian Ocean during January and July derived from Florida State University (FSU) wind stress climatology. Contours indicate magnitude of wind stress and vectors (not to scale) show direction. Taken from Rao (1998). (1 dyn cm$^2$ = $10^{-1}$ newtons m$^{-2}$).

where $(\tau_x\ , \tau_y)$ is the wind stress at location **r** and time $t$ and $(\tau_x^\circ, \tau_y^\circ)$ is its temporal mean; $A_x^a$ and $A_x^{sa}$ ($A_y^a$ and $A_y^{sa}$) are amplitudes of annual and semi-annual components of $\tau_x$ ($\tau_y$) respectively; $\omega_a (\omega_{sa})$ equals $2\pi/12$ month$^{-1}$ ($2\pi/6$ month$^{-1}$); $\phi_a^x, \phi_{sa}^x, \phi_a^y$, and $\phi_{sa}^y$ are phases; $\tau_x^\circ$, $\tau_y^\circ$, $A_x^a$, $A_x^{sa}$, $A_y^a$, $A_y^{sa}$, $\phi_a^x$, $\phi_{sa}^x$ , $\phi_a^y$, and $\phi_{sa}^y$ can be determined by a least-square fit to monthly-mean wind stress fields such as those depicted in figure 4. Amplitude of the annual, and semi-annual harmonics of $\tau_x$ and $\tau_y$, i.e. $|A_x^a|$, $|A_y^a|$, $|A_x^{sa}|$, and $|A_y^{sa}|$ respectively, determined from such a fit then delineate the areas where the annual and semi-annual variability is large. As can be seen from figure 5, the region of most significant variability is the western Arabian Sea, most notably the area off Somalia. It should be noted that the region of generation of the waves need not necessarily coincide with the regions of high wind stress magnitude. The forcing of the waves is a function that involves gradient of wind stress, or only one component of wind stress, or some other function of wind-stress field, depending on the kind of waves that are generated.

The equatorial region can support a variety of waves (see, for example, Moore 1968 and Gill 1982). To understand the seasonal cycle of wind-driven circulation in the North Indian Ocean we primarily need to consider waves with two periods, annual and semi-annual. The wavelengths associated with these periods can be both short (less than about 100 km) or long. However, short waves tend to get dissipated quickly due to friction, and hence do

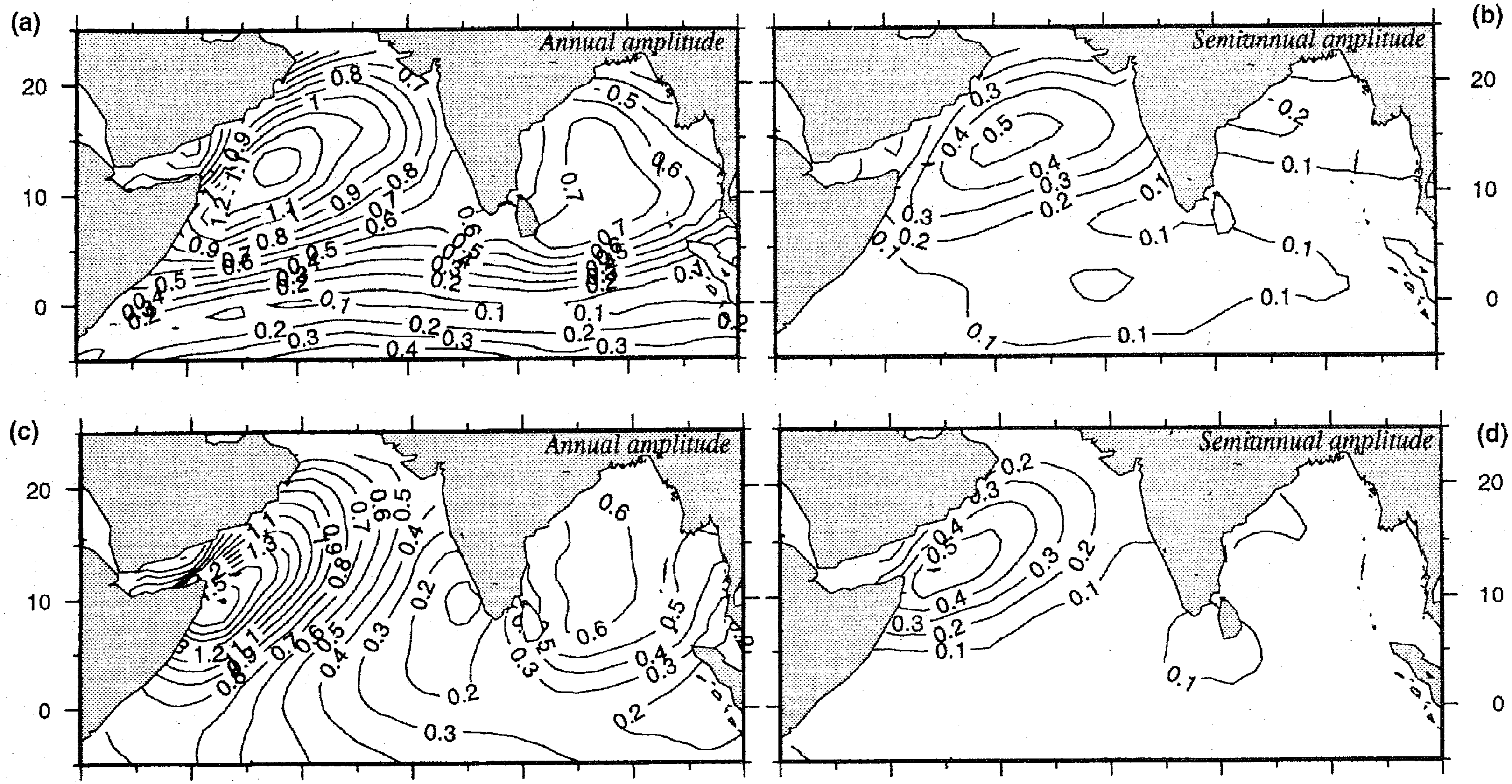


**Figure 5.** Annual and semi-annual components of wind stress (dyn cm$^{-2}$) in the North Indian Ocean. (**a**) Annual amplitude of eastward wind stress, $|A_x^a|$ in (2). (**b**) Semi-annual amplitude of eastward wind stress, $|A_x^{sa}|$. (**c**) Annual amplitude of northward wind stress, $|A_y^a|$. (**d**) Semi-annual amplitude of northward wind stress, $|A_x^{sa}|$. Taken from Rao (1998).

not make any significant contribution to large-scale circulation. Hence we need to consider only the long waves. Without getting into the details of how the disturbances triggered by the wind-stress propagate around the basin, it is possible to sketch the characteristics of this propagation from the knowledge of the waves that are triggered. We next examine these waves.

It has been found that the seasonal cycle of surface circulation can be simulated reasonably well with a $1\frac{1}{2}$-layer reduced-gravity model on equatorial $\beta$-plane (see, for example, Shankar and Shetye 1997). Consider a linear, frictionless version of such a model. The equations for the active upper-layer are:

$$(\partial u/\partial t) - \beta y v = -g\prime(\partial h/\partial x), \tag{4}$$

$$(\partial v/\partial t) + \beta y u = -g\prime(\partial h/\partial y), \tag{5}$$

$$(\partial h/\partial t) + h_{\circ}[(\partial u/\partial x) + (\partial v/\partial y)] = 0, \tag{6}$$

where $u$ is the velocity in the $x$-direction (eastward); $v$ is the velocity in $y$-direction (northward); $t$ is time; $\beta = \partial f/\partial y$, where $f$ is the Coriolis parameter. The appropriate values for $g\prime$ and $h_{\circ}$, as determined from experiments by Shankar and Shetye (1997) are, respectively, 0.035 m s$^{-2}$ and 100 m. In the discussion below, a quantity that appears often is $c = (g\prime h_{\circ})^{1/2}$, the phase velocity of an equatorially-/coastally-trapped Kelvin wave. Using the values of $g\prime$ and $h_{\circ}$ we get $c$ to be approximately 1.8 m s$^{-1}$.

Equations (4)–(6) permit a number of waves. Below we summarize the properties of low-frequency (with periods of the order of a few months or longer) and low-wavenumber (wavelengths longer than a few hundred kilometres) waves that are most important to understand the seasonal circulation.

### 3.1 *Equatorially-trapped Kelvin waves*

These waves are trapped in the vicinity of the equator, with the trapping radius, the equatorial radius of deformation ($R_e$), given by $(c/2\beta)^{1/2}$. For the choice of $c$ made earlier, $R_e$ is about 200 km. The phase velocity of these waves is always eastward. The advective particle velocity associated with the wave is strictly along the equator and can be both eastward or westward. The cross-equatorial advective velocity is zero. For these waves the equatorial regime acts as a waveguide centred on the equator, with a width of about 400 km (see figure 6).

### 3.2 *Long equatorially-trapped Rossby waves*

These waves propagate westward approximately nondispersively with wave speed,

$$\sigma/k = -c/(2n+1), \tag{7}$$

where $\sigma$ is the frequency (equal to $2\pi/T$, $T$ being the period), $k$ is the zonal wavenumber (equal to $2\pi/\lambda$, $\lambda$ being the wavelength), and $n = 1, 2, 3, \ldots$. Hence the waves propagate westward with a phase velocity that is at least three times slower than that of the Kelvin wave. The advective particle velocity can have both eastward and northward components. It can be shown that away from the equator a local dispersion relation can be written for these waves,

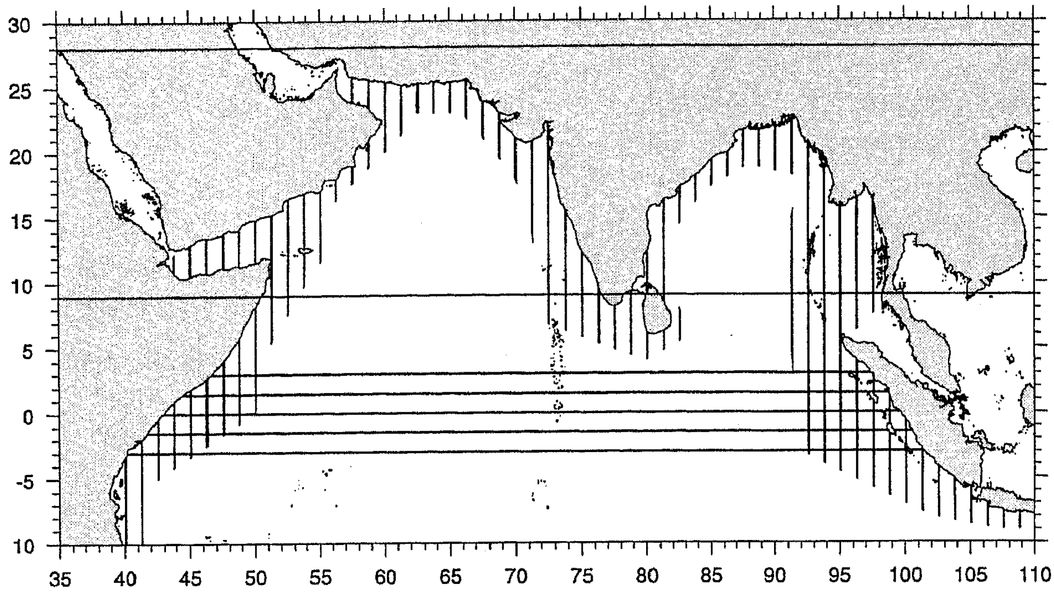


**Figure 6.** Coastal and equatorial wave-guides in the North Indian Ocean. The band between 3 N and 3 S shows the equatorial Kelvin waveguide. Vertical lines give the approximate extent of the coastal waveguide. Zonal lines at 28 N and 9 N give $y_c$ (9) for waves with periods of 180 and 60 days respectively.

$$\sigma/k = -c^2/\beta y^2, \tag{8}$$

which states that the local phase velocity increases as distance from the equator decreases.

The long equatorially-trapped waves have a trapping scale given by

$$y_c = c/2\sigma. \tag{9}$$

For the annual wave $y_c$ is approximately 56°. For the semi-annual waves, it is 28°. Hence the North Indian basin lies equatorward of turning latitudes of both these waves. However, for a wave with period of 60 days, $y_c$ is about 9°, and hence the basin has regions both poleward and equatorward of the latitude up to which the waves get trapped (see figure 6).

### 3.3 *Coastally-trapped Kelvin waves*

The coastally-trapped Kelvin waves propagate with the coast to their right in the northern hemisphere (Gill 1982). They are non-dispersive, and are trapped at the coast with a trapping scale, the Rossby radius of deformation ($R_c$), given by ($c/f$). The trapping makes the coastal zone a waveguide of these waves. Of course the width of the waveguide varies with latitude because $f$ depends on latitude. Using the value of $c$ defined earlier, the value of $R_c$ is about 75 km at 10°N, and 150 km at 5°N. The advective velocity associated with these waves is along the coast and it can be either in the same direction as that of the phase propagation, or opposite. The cross-shore component of the advective velocity is always zero.

All these features hold in the region poleward of the critical latitude ($y_c$) defined by (9). If a Kelvin wave propagates along a meridional eastern boundary which is equatorward of $y_c$, it radiates westward propagating Rossby waves and the coastal waveguide turns "leaky". In this case the Kelvin wave is not separable from the Rossby wave, and the two remain coupled to form a mode. Annual and semi-annual coastally-trapped Kelvin waves propagating along the eastern boundaries of the Bay of Bengal and the Arabian Sea fall in this category. In figure 6 we mark the approximate extent of the coastal waveguide.

### 3.4 *Propagation of a disturbance around the basin*

We can use figure 6 to track qualitatively how a disturbance generated anywhere in the North Indian basin will propagate. Note that the phase velocities of all the three waves are independent of $\sigma$. However, the nature of the waves, for example the value of $y_c$, depends on $\sigma$. Hence we need to define the time-scale associated with this disturbance. Let us assume that it is six months (semi-annual period) and that the disturbance is located in the Arabian Sea at a latitude of 10°N and 60°E. This location being outside the coastal or equatorial waveguide can excite only westward propagating Rossby waves. At 10°N their speed of propagation is $0.11\ \mathrm{m\,s^{-1}}$. Hence in about 104 days the disturbance reaches the coast of Somalia. When long Rossby waves reflect off a coast, they excite short Rossby waves that get dissipated quickly. Moreover, it has been pointed out (Cane & Gent 1984) that for all western boundary geometries the incoming mass flux normal to the coast due to the long Rossby waves cannot be balanced by the short waves. Hence the short waves do not play any significant role and can be neglected. Coastally-trapped Kelvin waves are needed to balance the mass flux.

Off Somalia, coastally-trapped Kelvin waves can only propagate towards the equator in the coastal waveguide. The disturbance can now propagate much faster, with velocity $c$, and takes 6 days to travel from Somalia to the northern edge of the equatorial waveguide. Once in the waveguide, the disturbance travels eastward as an equatorially-trapped Kelvin wave, again with velocity $c$. It takes 35 days to reach the coast of Sumatra from the western edge of the equatorial regime.

Reflection of the equatorially-trapped Kelvin wave from the coast of Sumatra produces coastally-trapped Kelvin waves that move away from the equator both to the north and the south, and Rossby waves that propagate westward. Note the $y_c$ for 6-monthly disturbances being approximately 28° of latitude, the coastally-trapped Kelvin waves and the Rossby waves remain coupled. The coastally-trapped Kelvin waves move along the periphery of the Bay of Bengal at speed $c$ and hence reach the northern tip of Sri Lanka from Sumatra in 28 days.

During this time the slower Rossby waves travel with a velocity that is dependent on the latitude, (8). At 10°N, the speed of these waves is $0.11\ \mathrm{m\,s^{-1}}$, hence they take 127 days to reach the north coast of Sri Lanka, i.e. about 4 times longer than Kelvin waves to go around the Bay of Bengal and reach this coast. Note that once the Rossby waves reach the coast, they excite a coastally-trapped Kelvin wave in a fashion similar to that off Somalia seen earlier.

A coastally-trapped Kelvin wave present on the east coast of Sri Lanka can bend around the island and reach approximately 10°N on the west coast of India in approximately 11 days. It can then travel all the way around the rim of the Arabian Sea and reach the coast of Somalia in 40 days. Hence if a disturbance is generated along the coast of Somalia, the waves emanating from it can go around the North Indian Ocean in the coastal and equatorial waveguides and reach the original spot of disturbance in 4 months. This is true for both annual and semi-annual disturbances.

However, the spot of original semi-annual disturbance we are examining was not in the coastal waveguide, but outside of it in the open Arabian Sea. For the waves caused by this disturbance to reach the original location of the disturbance, they need to travel westward as Rossby waves from the west coast of India. The Rossby waves take about 140 days to travel from the coast of India to 60°E, which is the location of the original disturbance. Hence the total time required for the waves to reach the original spot of generation after going around the basin is almost 330 days.

We note two points. First, the coastal waveguide off the southern tip of Sri Lanka is well separated from the equatorial waveguide. Had this not been the case, the disturbance propagating as the coastally-trapped Kelvin wave along the east coasts of India and Sri Lanka would have entered the equatorial waveguide and then propagated eastward as an equatorially-trapped Kelvin wave. The nature of propagation in this case would have been different from that described above. Second, the travel times for semi-annual waves are identical to those for the annual wave. In fact the estimates of travel times are the same for all waves with frequencies such that $y_c$ lies to the north of the northern boundary of the basin.

## 4. Events off the west coast of India

We will now use the waves discussed above to understand the events off the west coast of India as depicted in the schematic in figure 3. Shankar and Shetye (1997) have noted that the dynamics in the North Indian Ocean are, to a very good approximation, representable by the linear model given by (4)–(6). They demonstrated this by showing that a linear model forced with realistic wind stress reproduces virtually all the features just as well as a nonlinear model, except in the region off Somalia, where nonlinear terms were needed to reproduce the well-known eddy, the Great Whirl, that forms during the southwest monsoon. In the rest of the North Indian Ocean it should be possible to understand the circulation in terms of the waves that were discussed in the last section.

An important implication of the discussion in the previous section is that, because the North Indian Ocean basin is sufficiently small, and sufficiently close to the equator, the waves generated by winds at a location here can reach other parts and have an impact at locations far away from the region of generation within an year. McCreary *et al* (1993) have carried out numerical experiments to show how wind-generated disturbances present in the Bay of Bengal can travel to the west coast of India setting up circulations there. The only way these disturbances can reach the west coast is as coastally-trapped Kelvin waves.

But, as these Kelvin waves propagate northward along the west coast of India, they must radiate westward propagating Rossby waves. The question then is, can the Rossby waves together with the Kelvin waves set up circulations that have been associated with the seasonal cycle of the WICC and the LH/LL? Shankar and Shetye (1997) have argued that they can. This was shown by carrying out the following numerical experiment. The authors used a model essentially based on (4)–(6). The model was forced, not by imposing winds, but by generating a coastally-trapped Kelvin wave by imposing a velocity perpendicular to the coast at the northern end of the Bay of Bengal. The imposed flow had a period of one year and was defined by

$$v(x, 20^\circ, t) = -0.25 v_o \cos(\omega t)|x - 91.5^\circ|, \text{ where } 87.5^\circ \le x \le 91.5^\circ. \quad (10)$$

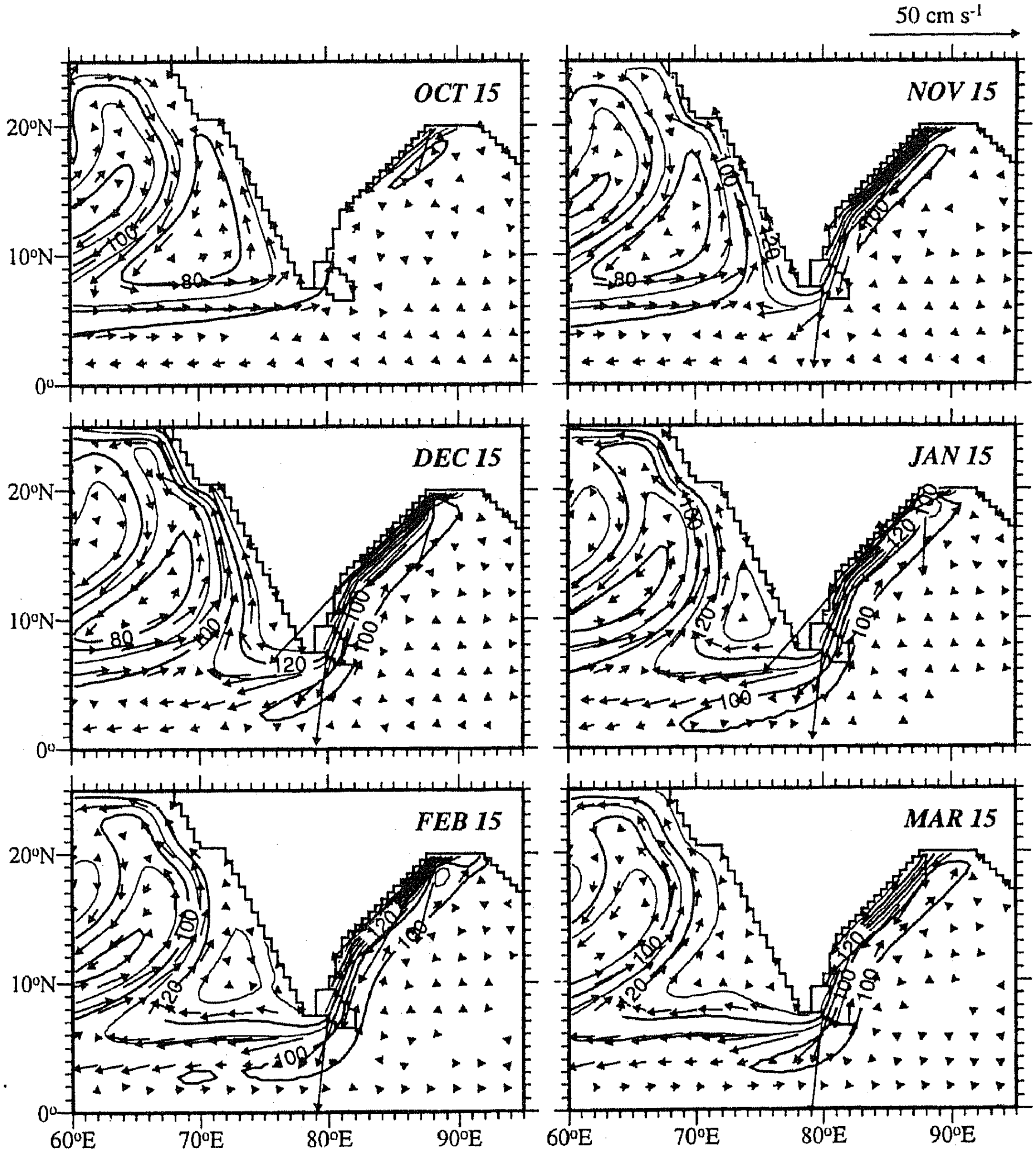


**Figure 7.** Upper layer thickness $h$ (m) and velocity $(u, v)$ $(\mathrm{cm\,s^{-1}})$ for the simulation based linear $1\frac{1}{2}$-layer reduced gravity model. Contour interval is 10 m. Taken from Shankar & Shetye (1997).

Here $x$ is in degrees longitude, $\omega = 2\pi$ year$^{-1}$, and $v_o = 40\,\mathrm{cm}^{-1}$. Thus there is flow into (out of) the Bay during October-March (April–September), generating an annual coastally-trapped Kelvin wave along the east coast of India. Figure 7 summarizes the results of this numerical experiment with monthly plots of $h$ and $(u, v)$ for October–March, thus showing one half of an annual cycle of events that results in the formation of a high off southwest India. Inflow into the northern bay starts on October 1, beginning the downwelling phase of the Kelvin wave at the east coast; by mid-October, there is equatorward flow in the northern half of the east coast but the current is still poleward in the south. This is the continuation of a current along the western edge of a low in the Arabian Sea, which formed off southwest India in July and moved offshore. This current bifurcates at the southern tip of the Indian subcontinent, part of the flow being deflected poleward along the west coast and the rest bending around Sri Lanka to flow poleward along the east coast. By mid-November, the downwelling phase of the Kelvin wave manifests all along the coastline and the equatorward EICC bends around Sri Lanka to flow on as the poleward WICC. Radiation of Rossby waves weakens the WICC off southwestern India, and it eventually reverses even as the equatorward EICC peaks by January 1. By mid-January, there is a weak equatorward, upwelling-favourable current off southwestern India, closing the circulation around a high. This cycle of events comes to a close by March, when the high drifts westward together with the Rossby waves that constitute it, spreading across the southern Arabian Sea; the Rossby wave signature now appears in the form of a westward current in the vicinity of 5° N and northeastward velocities to the north of this latitude. By the end of March, the inflow in the northern bay vanishes; outward flow starts on April 1, beginning the second half of the annual cycle. This leads to a low off southwestern India in July.

The pattern seen off the west coast of India is the result of a group of coastally-trapped Kelvin waves propagating along the coast and the Rossby wave radiated by the Kelvin waves. It can further be shown that instead of only an annual cycle, if we had superposition of annual and semi-annual waves, we can get the asymmetry between the magnitude of the LH and that of the LL that has been seen in the observations from this area. The magnitude of the sea level high during the northeast monsoon is generally higher than of the low during the southwest monsoon.

Shankar and Shetye (1997) further showed that the pattern of evolution of the LH/LL can be reproduced analytically by superposition of solutions consisting of Rossby waves and coastally-trapped Kelvin waves. In this solution it was pointed out that were the period associated with Kelvin waves less than about 40 days, then they would not be able to radiate Rossby waves and hence a high/low could not form. We can understand this result in terms of $y_c$ given in (9). For the period of 40 days $y_c$ is approximately 500 km. Hence Rossby waves with this period can exist only within this distance of the equator. The southern tip of Sri Lanka is at about 6°N, or about 660 km from the equator. Hence no Rossby waves with period of 40 or fewer days can exist off the west coast of Sri Lanka. The coastally-trapped Kelvin wave will then exit without getting coupled with a Rossby wave. As the period increases, so does the northern limit of the Rossby waveguide.

## 5. Discussion

What then are the elements that lead to formation of the West India Coastal Current and Lakshadweep High/Low? In its simplest form, the WICC is a coastally-trapped Kelvin wave, and the Lakshadweep High/Low the result of Rossby waves radiated by the Kelvin wave. Thus, to have the WICC and LH/LL we need two basic elements: (1) coastally-trapped Kelvin waves along the east coast of India; (2) the periods of the Kelvin waves must be large enough so as to have the $y_c$ associated with them significantly to the north of the region where the LH/LL form.

As seen earlier, the monsoon winds have associated with them, annual and semi-annual periods. It has been shown earlier that the EICC is forced by the winds on the equator, the winds along the periphery of the Bay, and those in the open Bay (Shankar *et al* 1996; McCreary *et al* 1996). Each of these forcing functions has annual and semi-annual waves associated with them. It is the superposition of these forcing functions that forms the EICC which can be looked on as a sum of Kelvin waves with annual and semi-annual periods. When the Kelvin waves reach the west coast of India after turning around Sri Lanka, and make their way northward along the coast, they radiate annual and semi-annual Rossby waves. The Kelvin waves along the coast form the WICC, and the Rossby and Kelvin waves form the LH/LL. As the Rossby waves propagate westward, the LH/LL too stretch westward.

The special features of the basin which make this symphony of waves possible are: the tropical character of the basin which makes the existence of the waves possible; and the occurrence of the monsoon which provides the forcing mechanism to generate the waves with the required periods.

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