

Dynamics of Circulation of the Waters around India: Recent Developments and Key Issues for the Future

Satish R Shetye

National Institute of Oceanography
Dona Paula, Goa 403 004

Abstract

The 1990s witnessed the development of a new theoretical framework to understand the dynamics of circulation in the north Indian Ocean. According to this framework, the seasonally varying wind stress acting at the surface of this small, tropical basin triggers Equatorial Kelvin/Rossby waves and coastal Kelvin waves. The superposition of these waves determines the seasonal, near-surface, wind-driven circulation in the basin. While this framework ignores the coupling between the ocean and the atmosphere, the year 1997 saw the discovery of the Indian Ocean Dipole (IOD) that appears to be a coupled ocean-atmosphere mode. The IOD did not have an impact on the Indian Summer Monsoon (ISM) of 1997. Its discovery, however, implies that a key issue in the next decade will be search for coupled atmosphere-ocean modes that influence the ISM. Recent datasets, new experiments, and an augmented ocean observing system are expected to play major roles in this search.



Introduction

If one divides the ocean basins of the earth into their northern and southern parts, with the equator as the dividing line, then the surface area of the north Indian Ocean is the smallest, being much smaller than that of the south Indian Ocean and the north/south Pacific/Atlantic Ocean. Another feature of the north Indian Ocean is that it is not connected to the polar seas. Both these features, the smallness and lack of access to polar seas, are the result of the presence of the Asian continent to the north of the Indian Ocean.

The presence of the continent adds yet another special feature to the north Indian Ocean. The winds over it exhibit a distinct seasonal cycle—the monsoons. Traditionally looked at as a large-scale land-sea breeze, the occurrence of the Indian Summer Monsoon during June-September marks the arrival of Inter-Tropical Convergence Zone (ITCZ) over the north Indian Ocean and the surrounding lands, aided by heating of the atmosphere by the Tibetan Plateau which has an average elevation of about 5 km. During the winter monsoon in November-February, the ITCZ is located over the south Indian Ocean, well to the south of the equator. The seasonal migration of the ITCZ, as seen in Figure 1, though observed everywhere over the tropics, is particularly large over the Indian Ocean. Also, the activity of the ITCZ during the Indian Summer Monsoon is particularly large over the Bay of Bengal and along the west coast of India, while the ITCZ disappears over the western Arabian Sea.

Accompanying the migration of the rain-bearing belt of the ITCZ are dramatic changes in the direction and magnitude of winds over the North Indian Ocean (Figure 1). These seasonal winds, traditionally called the monsoons, drive a circulation in the near-surface (approximately the top 500 m) north Indian Ocean that shows complete reversal in direction from season to season. The last decade saw major developments in the dynamical framework to understand this wind-driven circulation. The last few years also saw the discovery of a new phenomenon, the Indian Ocean Dipole (IOD), over the equatorial Indian Ocean. This article examines both these developments. The next section summarizes the

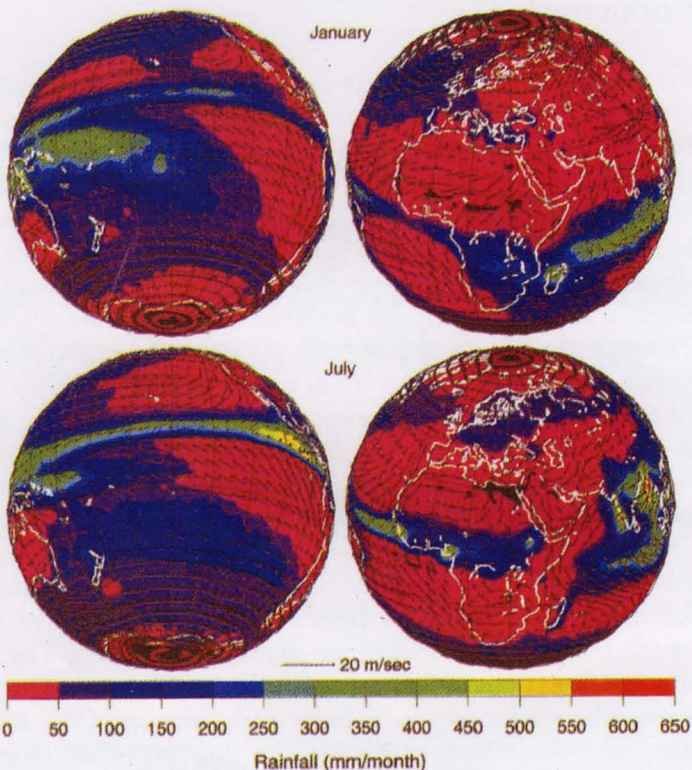



Figure 1 Monthly-mean Precipitation (mm/month) and Winds (m/s) during January and July.

dynamical framework. Section 3 examines the IOD and its implications. The concluding Section 4 looks at the promises that recent developments hold to further our understanding of dynamics of the north Indian Ocean.

Dynamics of Wind-driven Circulation of the North Indian Ocean

The annual cycle of winds over the north Indian Ocean (see January and July winds indicated in Figure 1) has the following



characteristics. First, the Indian Summer Monsoon winds are predominantly from the southwest, whereas during the winter monsoon the winds are mostly northeasterlies. Second, the strength of the winds during summer is much stronger than the northeasterlies of the winter. Third, during the summer monsoon, there is a distinct east-west gradient in the magnitude of winds over the ocean: winds over the western Arabian Sea are much stronger than those in the eastern Arabian Sea and those in the Bay of Bengal. The strength of the winds in the western Arabian Sea is due to the presence of the East African Mountains which serve as the western boundary to the low-level atmospheric flow, in the same way as the shelf and slope off the eastern United States of America serves as the western boundary to the oceanic flow over the North Atlantic. In each case, we have an intense return flow, the Gulf Stream in the Atlantic, and the Findlater Jet in the equatorial Indian Ocean (Anderson, 1976; Hart *et al.*, 1978).

The wind stress that the above annual cycle of winds imposes on the ocean can be represented as a superposition of a mean windstress and two harmonics, one with the period of a year and the other with six months (Rao, 1998). Of the three, the annual harmonic has the largest amplitude, about twice that of each of the other two, which are similar in magnitude (Figure 2). The amplitude of each of the three components is largest over the Arabian Sea; the three together drive the North Indian Ocean. To a good approximation, the resulting circulation can be looked at as superposition of circulation driven by each of the three components separately. This is a consequence of the linear character of the ocean. Except for the region off the coast of Somalia, that shows strong seasonal recirculations or eddies, the non-linear advective terms can be safely ignored in the momentum balance in the north Indian Ocean.

The linear character of the basin dynamics also implies that the seasonal circulation can be looked at as superposition of waves triggered by the annual and semi-annual wind-stress. What kind of waves are these? A tropical basin, like the north Indian Ocean, can support a variety of waves (see, for example, Gill, 1982). Below, we summarize the properties of low-frequency (periods of the order of few weeks or longer) and low-wavenumber (wavelengths long than a few

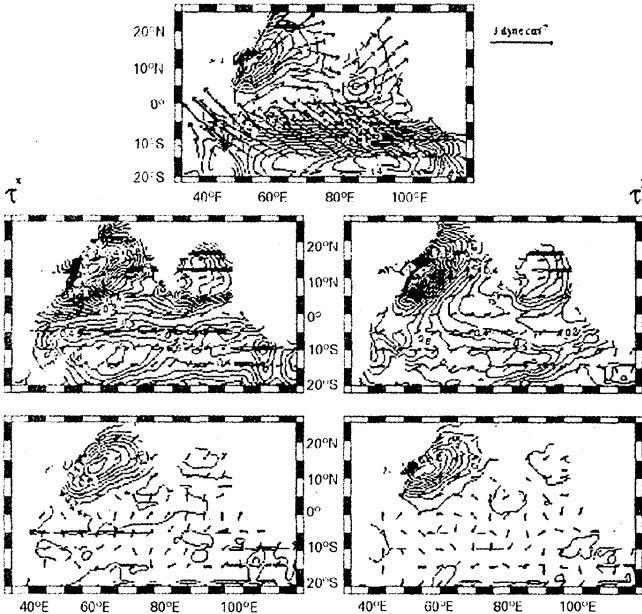


Figure 2 Wind-stress, as given by Helleman and Rosenstein (1983) is decomposed into a mean, an annual harmonic and a semi-annual harmonic. The arrows in the frame at the top gives the annual mean wind-stress. The middle left frame gives the amplitude (contours) of annual harmonic of the zonal component of the wind stress. The angle of the stick in the same frame gives the time of peak stress using the following convention. The angle is measured in the counterclockwise direction from the positive x-axis (end of March, June, September and December correspond to 90, 180, 270, and 360 degrees, respectively). The remaining three frames, middle right, bottom left and bottom right represent, respectively, the annual harmonic of the meridional component, the semi-annual harmonic of the zonal component, and the semi-annual component of the zonal component.

hundred kilometres) waves that are most important to understand the seasonal circulation of the north Indian Ocean.

- 1) *Equatorially-trapped Kelvin waves*: These waves are trapped in the vicinity of the equator with a trapping radius of about 300 km. The phase velocity of these waves is about 2 m/s and 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 500 km, as indicated in Figure 3.
- 2) *Long equatorially-trapped Rossby waves*: These waves propagate westward non-dispersively. At the equator, their wave speeds are slower than the Kelvin waves by a factor of 3, 5, 7, 9, etc. Their phase velocity at locations away from the equator are

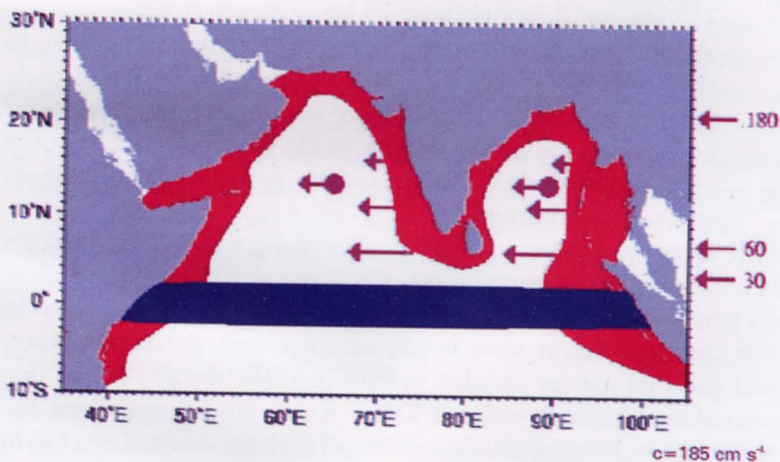



Figure 3 Equatorial and Coastal Waveguides of the north Indian Ocean as described in Section 2. The blue region shows the waveguide for the equatorial Kelvin wave, and red for the coastal Kelvin wave. The trapping radius for equatorial Rossby waves with periods of 30, 60, and 180 days are shown with pink arrows to the right of the frame. The number next to an arrow gives the period of the wave. Pink arrows inside the frame give the magnitude of the westward phase velocity of Rossby waves at different latitudes.




slower and is inversely proportional to the square of the distance from the equator. The trapping radius of these waves is dependent on the period: the longer the period, the larger is the trapping radius. For a wave with the period of about 60 days, the trapping radius is about 1000 km; for six-monthly waves it is about 3,000 km, which is of the same order as the meridional extent of the north Indian Ocean. In Figure 3, the trapping radius of a wave is indicated with the help of a pink arrow with the period of the wave written next to it. The particle velocity of Rossby waves can have both eastward and northward components.

- 3) *Coastally-trapped Kelvin waves*: The coastally-trapped Kelvin waves propagate with the coast to their right in the northern hemisphere. They are non-dispersive, and are trapped at the coast, with a trapping scale that is inversely proportional to the Coriolis parameter. At 5°N, the trapping scale is about 150 km, at 10°N it is about 75 km. The approximate trapping scales of the waves, or the extent of the coastal waveguide, in the north Indian Ocean are shown in Figure 3. The particle velocity associated with this wave is along the coast and it can be either in the same direction as the phase velocity or in the opposite direction. The cross-shore particle velocity due to this wave is always zero.

If a coastal Kelvin wave propagates along a meridional eastern boundary which is equatorward of a "critical latitude" (dependent on period of the wave), then it radiates westward Rossby waves. In this case, the Kelvin wave is not separable from the Rossby wave, and the two remain coupled to form a single 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. The coastal wave-guide thus turns "leaky" and loses some of its energy as it propagates.

Figures 2 and 3, in essence, provide a framework to understand the wind-driven circulation of the north Indian Ocean. Figure 2 provides details of the forcing that is exerted on the ocean and Figure 3 gives the expected response in the form of the three trapped waves. In order to find out the details of how the velocity field will evolve at a location, it is necessary to compute the details



regarding generation of the waves, their propagation, etc. However, the overall characteristics of the circulation can be understood using Figure 3.

The tropical nature of the north Indian Ocean, its small size, and the presence of strongly seasonal forcing makes the classical Sverdrup Equation (see Gill, 1982, p 465) unsuitable to understand the wind-driven circulation of this ocean. The alternative framework that we have discussed here, i.e. to understand the circulation in terms of locally- and remotely-forced tropical and coastal waves, began to take shape with the insights provided by numerical experiments described in McCreary *et al.* (1993). The importance of these waves in generating observed features, such as the Lakshadweep High/Low that are unique to this region, was highlighted in analytic solutions and numerical experiments in Shankar and Shetye (1997). New observations, too, played an important role in identifying these features. The 10-day ocean surface topography, such as that provided by the Topex-Poseidon altimeter, made it possible to appreciate the structures of the large-scale wind-driven circulation and their evolution during the course of the seasonal cycle. This evolution was much too rapid to be properly appreciated in the earlier data that were mainly based on hydrographic surveys. The new data acquired with new technology centred around satellite-based sensors have thus played an important role in providing new insights into the working of the north Indian Ocean and in evolving a framework to understand it better.

The Indian Ocean Dipole

The new data also played a major role in the discovery of a phenomenon that had remained undetected earlier. The occurrence of a coupled ocean-atmosphere phenomenon in the equatorial Indian Ocean, starting in summer 1997 and lasting till early 1998, was reported by Saji *et al.* (1999) and Webster *et al.* (1999). Now often called the Indian Ocean Dipole (IOD), this event begins with an anomalous upwelling and cooling of sea surface temperature (SST) in the eastern equatorial ocean and ends with a warming in the western equatorial ocean (Figure 4). The cooling in the east causes

draughts over Indonesia; the warming in the west leads to severe rainfall over eastern Africa. During the IOD event there are spacio-temporal links between SST and winds. It has therefore been suggested that air-sea coupling, through precipitation and ocean dynamics, dominates the evolution of this event. The ocean waves associated with this phenomenon are the same as those discussed in the last section. What is new is that the coupling between the atmosphere and the ocean plays a role: anomalous winds trigger ocean dynamics that generate SST anomalies at a different location which, in turn, generates wind anomalies, and so on. This is the first time that such a coupling has been invoked to explain the events occurring in and over the Indian Ocean.

For India, the most important climate element is the Indian Summer Monsoon. The bay is an important site for the monsoon activity. It forms a location for the most intensive deep convection and precipitation in the tropical oceans during the months of July

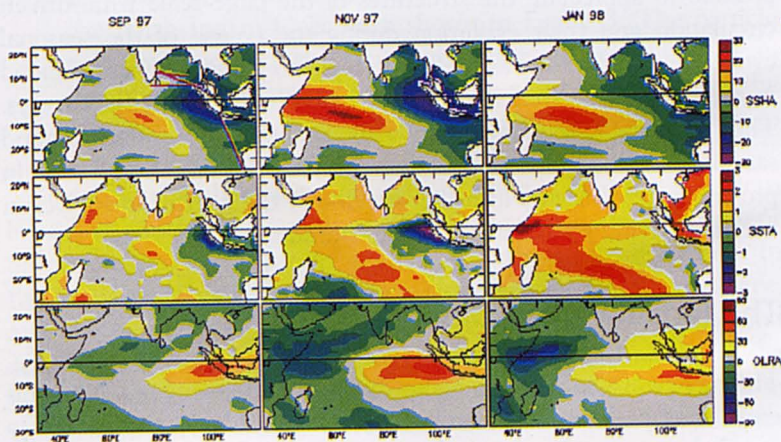
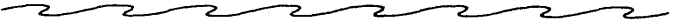


Figure 4 Taken from Rao et al. (2002), the figure shows the evolution of the Topex–Poseidon sea surface height anomalies (cm) (top row), the Reynolds sea surface temperature anomalies ($^{\circ}\text{C}$) (middle row) and NCEP outgoing longwave radiation anomalies (watts/m^2) (bottom row) during September 1997 (first column), November 1997 (middle column), and January 1998 (right column).

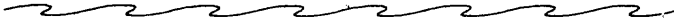


and August, which are the two most active months of the summer monsoon. It is known that to sustain such convection SST has to exceed a critical threshold of about 28°C (Gadgil *et al.*, 1984; Graham and Barnett, 1987). Any phenomenon, such as the IOD, with the potential to alter the SST in the Bay of Bengal is therefore important.

An interesting feature observed during the 1997 IOD was that, even though coastal Kelvin waves propagated into the bay, there was no cooling of SST. This can be seen in Figure 4. This is in striking contrast to the cooling observed south of the equator off Sumatra. Why the difference? This question has been addressed by Rao *et al.* (2002). They argue that there was indeed a cooling of the subsurface waters associated with the propagation of the coastal Kelvin waves in the bay. However, the shallowing of the thermocline associated with the upwelling favourable Kelvin wave did not result in lowering of the SST for two reasons. First, the near-surface stratification in the bay due to salinity variation is high; second, the winds in the bay are comparatively weak (much weaker than those in the Arabian Sea, as discussed in Section 2). The combined result of the two is that the subsurface cooling is not carried to the surface because the winds are too weak to break down the near-surface stratification. The near-surface stratification is high for two reasons. First, the Bay of Bengal is a region of high precipitation. Second, the bay receives large quantities of freshwater in the form of runoff from the rivers that are located on the lands that surround the bay. In essence, the low-salinity layer that results from high precipitation and runoff helps to sustain the SST in the bay above the threshold of 28°C which, in turn, maintains the high activity of atmospheric convection.

Concluding Comments

Though the IOD of 1997 had no impact on the SST in the bay, the discovery of IOD marks an important development in understanding coupling between the atmosphere and the north Indian Ocean. The discovery was made possible because new data sets became available. The improvements in our understanding of dynamics of the



north Indian Ocean that were described in Section 2 have also benefited from the new observations. Most of these use satellite-based sensors, and are useful to monitor the surface of the ocean. However, we still need to have further improvements in the subsurface monitoring of the waters of the north Indian Ocean. This ocean, especially the Bay of Bengal, remains one of the data-poor areas of the world oceans, particularly with respect to subsurface data.

The near future promises to help in this regard. The ARGO system, that is now under implementation, is expected to provide subsurface data with a temporal resolution of about 10 days. Such data, together with satellite-based sensors that provide similar temporal resolution, would be of great help in improving our understanding of the north Indian Ocean whose peculiarity is high temporal variability. In the near future, historical data, new observations platforms, and new experiments aimed at specific issues related to the coupled atmosphere-ocean system over the north Indian Ocean, are expected to provide new insights into the dynamics of this system.

References

1. Anderson, D L T (1976) The low-level jet as a western boundary current. *Monthly Weather Review*, **104**: 907–921.
2. Gadgil S, Joseph, P V & Joshi, N V (1984) Ocean-atmosphere coupling over monsoon regions, *Nature*, **312**: 141–143.
3. Graham, N E & Barnett, T P (1987) Sea surface temperature, surface wind divergence and convection over the tropical oceans, *Science*, **238**: 657–659.
4. Gill, A. (1982) *Atmosphere-Ocean Dynamics*, Academic Press, San Diego, USA.
5. Hart J E, Rao J E, van de Boogaard H, Young J A & Findlater, J (1978) Aerial observations of the East African low-level jet stream, *Monthly Weather Review*, **106**: 1714–1724.
6. Hellerman, S & Rosenstiel, M (1983) Normal monthly wind stress over the world ocean with error estimates, *Journal of Physical Oceanography*, **13**: 1093–1104.

7. McCreary J P, Kundu, P K & Molinari, R L (1993) A numerical investigation of dynamics, thermodynamics and mixed-layer processes in the Indian Ocean, *Progress in Oceanography*, 31: 181–224.
8. Rao, A S (1998) Rossby waves in the Bay of Bengal: Observations and simulations. Ph D thesis, Andhra University, Visakapatnam.
9. Rao S A, Gopalkrishna V V, Shetye S R & Yamagata, T (2002) Why were cool SST anomalies absent in the Bay of Bengal during the 1997 Indian Ocean Dipole Event? *Geophysical Research Letters* (in press).
10. Saji, N H, Goswami B N, Vinayachandran P N & Yamagata, T (1999) A dipole mode in the tropical Indian Ocean, *Nature*, 401: 360–363.
11. Shankar, S & Shetye, S R (1997) On the dynamics of the Lakshadweep High and Low in the Southeastern Arabian Sea, *Journal of Geophysical Research*, 102: 12,551–12,562.
12. Webster P J, Moore A M, Loschnigg J P & Leben, R R (1999) Coupled Ocean-Atmosphere Dynamics in the Indian Ocean during 1997–98, *Nature*, 401: 356–360.