

INTRUSION OF BAY OF BENGAL WATER INTO ARABIAN SEA ALONG THE WEST COAST OF INDIA DURING NORTH EAST MONSOON

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ABSTRACT

It is known that the low salinity Bay of Bengal water intrudes into the Arabian Sea along the west coast of India during the north east monsoon. This paper examines a possible driving mechanism for this intrusion. Using the recent theory of McCreary et al¹. In the present study it is indicated that the meridional salinity gradient that occurs in the eastern Arabian Sea after the withdrawal of the south west monsoon, is capable of forcing a northward current along the west coast of India, in spite of a weak equatorward wind stress.

INTRODUCTION

Two different surface water masses are formed in the Northern Indian Ocean - the high salinity (~36 ppt) water in the Arabian Sea and the low salinity water (~33 ppt) in the Bay of Bengal. The former is formed by the excess of evaporation in the central and northern Arabian Sea; the latter due to excess of precipitation and large runoff into the Bay of Bengal. During the north east monsoon (November-January) the low salinity water moves south, and then enters the Arabian Sea along two branches. One moves west, and other north along the west coast of India². The surface salinity distribution in the eastern Arabian Sea, shown in Fig.1, brings out the salient features of the low salinity water intrusion.

A peculiarity of the northward intrusion along the coast is that it occurs when the coastal winds, though weak, are equatorward. This feature of the coastal circulation was earlier pointed out by Shetye³. The mechanism that drives the intrusion against the wind is not well understood. During the north east monsoon of 1985-86 a hydrographic section was made on the shelf off the west coast of India. Though the study of the low salinity intrusion was not the intent behind making the

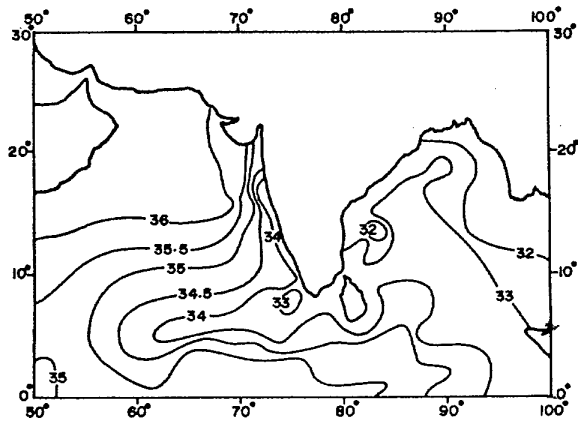


Fig.1. Surface salinity distribution in the North Indian Ocean during January-February (Data source: Wyrtky, 1974).

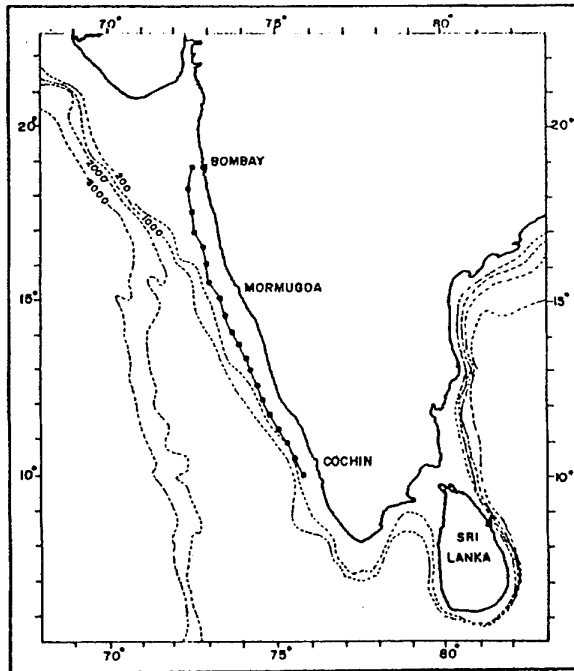


Fig.2. Station location for the hydrographic data shown in Fig.3.

section, it did provide an impetus to examine the driving mechanism for the intrusion. The next section describes the hydrographic data. In section III a recent theory of McCreary *et al*¹ has been used to identify a possible mechanism for the driving of the intrusion.

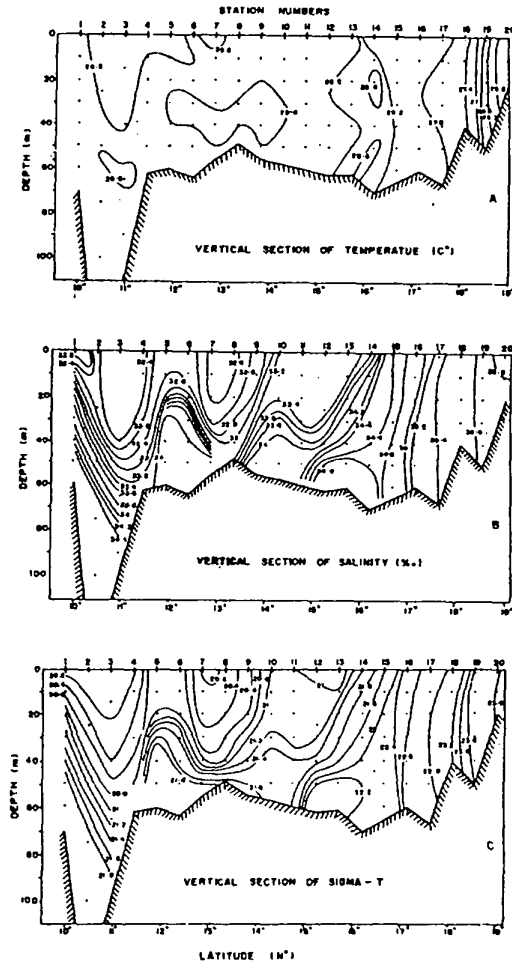


Fig.3. Hydrographic data. (See figure 2 for station locations) (A) Temperature ($^{\circ}\text{C}$), (B) Salinity (ppt) and (C) Sigma-t (g/cm^3).

DATA

During 25 December 1985 to January 1986 a cruise was carried out along the west coast of India. The ship, *R V Gaveshani*, moved northward approximately along the 50m isobath from Cochin to Bombay as shown in Fig. 2. data were collected at 20 stations at a spacing of half a degree. At each station salinity and temperature were measured at standard depths. Temperatures were measured using Nansen bottles and analysed for salinity using a ship board salinometer.

Fig. 3 A, B and C shows temperature, salinity and sigma-t distributions. As seen from the figure, the temperature field does not show any systematic variation with depth, nor with latitude between stations 1 and 17 the temperatures stay between 28.8°C and 29.0°C. It is only near the northern edge of the section that the temperature shows a small decrease with latitude. In contrast, the salinity field in Fig.3 B shows a well marked increase towards the north. The water column near the northern edge is vertically well-mixed and has salinity well in excess of 35.5 ppt. Vertical stratification increases towards the south. The southern portion of the section shows fresher 33 ppt water overlying saltier water. The only source of this 33 ppt water is the Bay of Bengal water. The sigma-t variation shown in Fig.3 C matches very well with salinity variation, implying that it is the salinity that controls the density field. It is also noted that the surface density increases monotonically from south to north. The increase is particularly sharp around the 18°N latitude.

The winds recorded at Goa and Mangalore at 0000 and 1200 IST hours during the cruise showed the land-sea breeze cycles to be the main signal. As seen in Fig.4, the longshore component of these winds was weak, of the order of 0.5m/s, and directed towards the Equator. This behaviour is consistent with the climatic monthly-mean wind field values⁴.

DISCUSSION

Because of the monsoons, the waters off the west coast of India experience a wind stress which is strongly time dependent. Data on monthly-mean winds show that the longshore component of the wind stress, which is equatorward throughout the year, is weak (0.1 dyne/cm²) in March. It begins to increase in April, reaches a peak value of around 0.5 dyne/cm² in July, then tapers off to about 0.1 dyne/cm² by November⁵. Climatic ship drift data indicate that from April to October the longshore surface current is equatorward. There is also evidence of upwelling along the coast. This behaviour of the coastal circulation

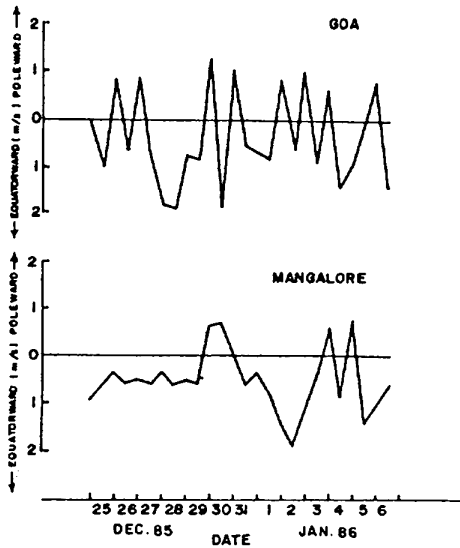


Fig.4. Longshore component of winds (m/sec) recorded at Goa and Mangalore.

is typical of the eastern boundary currents such as the Benguela, California and Canary currents. From November to January, during the north east monsoon, the longshore wind stress though equatorward is weak (0.1 dyne/cm^2), there is no evidence of upwelling. The northward coastal intrusion of the Bay of Bengal water occurs during this time.

The forcing for a typical coastal wind driven upwelling system arises from the offshore transport of water near the surface. Consequently, there is an onshore flow in the deeper layers. The strength of such a forcing mechanism can be estimated by the term $(T\ell/f)$ which is the magnitude of the surface Ekman transport; T and f are longshore wind stress and the Coriolis parameter, respectively. As pointed out by McCreary et al¹, thermohaline forcing of eastern boundary currents also arises due to similar onshore/offshore water movement. The difference is that the driving is not due to longshore wind but due to longshore density variation in the ocean interior. In the Arabian Sea, salinity increases towards the north. A consequence of the resulting density change is that the dynamic height of the sea surface drops towards the north. The resulting pressure field can drive a geostrophic eastward current, and westward current in the deeper layers. Such a circulation pattern is similar to the one that could be induced by a poleward wind stress. The main difference in the thermohaline driven case

is that the surface flow instead of being confined to a shallow Ekman layer, is spread upto the thermocline.

It is worthwhile to express the magnitude of the above thermohaline-driven surface flow as a function of the longshore density gradient. A comparison of such an expression which can then serve as a useful tool to determine the relative magnitude of thermohaline versus wind-driving. To derive such an expression we consider a north-south coast which forms the eastern boundary of a geostrophic, hydrostatic ocean interior. Let the depth of the ocean be D . The thermal wind relation then gives

$$\frac{\partial u}{\partial z} = -\frac{g}{\rho_0 f} \left(\frac{\partial \rho}{\partial y} \right)$$

where g , ρ_0 and f are acceleration due to gravity, mean density, and density perturbation respectively, u is the eastward velocity, y and z are the northward and upward co-ordinates respectively.

We assume that the meridional density gradient in the ocean interior varies exponentially with depth with an e-folding depth $(1/k)$, i.e. $\left(\frac{\partial \rho}{\partial y} \right) = \Delta \rho_0 e^{-kz}$ being the meridional density

gradient at the surface. Imposing the restriction $\int_0^D u dz = 0$

we get
$$u = \frac{g \Delta}{\rho_0 f k^2 D} \left[k D e^{-kz} - (1 - e^{-kD}) \right]$$

In the above, the onshore velocity is restricted to a depth S where

$$S = -1/k \ln \left(\frac{1 - e^{-kD}}{kD} \right)$$

The mass transport, T_s from surface to depth S is given by

$$T_s = \frac{g \Delta}{f k^2} \left[1 - \frac{1}{kD} (1 - e^{-kD}) - \frac{S}{D} (1 - e^{-kD}) \right]$$

When T_s exceeds (τ_i/f) , thermohaline-driving is expected to dominate over wind-driving. When $1/k$ and S are much smaller than D , T_s can be approximated by $\frac{g \Delta}{f k^2}$

Off the west coast of India, Δ during the southwest monsoon is virtually zero². The climatic monthly-mean wind-stress during the same season is around 0.3 dyne/cm². This makes τ_i/f approximately 8.0×10^3 g/(cm. sec).

During the north eastmonsoon τ_i is less than 0.1 dyne/cm², giving a value of 2.5×10^3 gm/cm sec.) for τ_i/f . Assuming

that the influence of surface density variation along the coast decreases by a factor of ten at a depth of 100 m, we get K equal to $2 \times 10^{-4} \text{ cm}^{-1}$. Taking Δ to be $1.5 \times 10^{-11} \text{ g/cm}^3$, we get T_s equal to $9 \times 10^3 \text{ g/(cm sec)}$. This over three times larger than the value of τ_1/f during this season, implying that the thermohaline driving would overwhelm wind driving.

The preceding discussion suggests the following with regard to the coastal intrusion of the Bay of Bengal water. During the southwest monsoon, the strong winds dominate the circulation over the Arabian Sea. When the southwest monsoon withdraws in October, the magnitude of wind driving decreases substantially.

Thermohaline - driving arising due to density contrasts then attains significance. The salinity in the northern reaches of the eastern Arabian Sea is higher than that near the southeast Arabian Sea. The resulting pressure field drives an eastward surface current and a westward deep current. The consequence of this deep-sea circulation is northward intrusion along the coast.

The magnitude of this intrusion is expected to be determined by the kind of meridional density contrast that occurs in the eastern Arabian Sea during the north east monsoon. This contrast in turn would depend on the behaviour of the preceding south west monsoon. Abnormally high precipitation/runoff in the Bay of Bengal or anomalous evaporation in the Arabian Sea will enhance the meridional salinity gradient, and hence the magnitude of intrusion. We thus have an interesting possibility of a delayed effect on the coastal circulation of an anomaly in the behaviour of the preceding southwest monsoon. At the present, this scenario is based on theoretical consideration alone. Observations are needed to check their validity. Observations are also needed to examine other implications of the theory of McCreary et al. For example, their theory predicts an equatorward undercurrent. It is not known whether this appears along the west coast at the time of the northward surface intrusion or not.

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