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On the Upwelling off the Southern Tip and along the West Coast of India

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ABSTRACT

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The processes of upwelling off the southern tip (the Kanyakumari coast) and the west coast of India are highly localised features with different forcing mechanisms, and they cannot be treated as a uniform wind-driven upwelling system. Off the Kanyakumari coast, upwelling is due to the south-west monsoon winds that are tangential to the coast. Along the west coast, the area between 8° N and 9° N represents the shadow zone to the influence of the remote forcing on the upwelling; the latter is forced by the longshore wind stress. Moderate to relatively intense upwelling occurs along the Kollam to Mangalore coast (9° N to 13° N) due to the combined action of the longshore wind stress, the coastally trapped Kelvin waves, and the offshore propagating Rossby waves. North of this area (13° N to 15° N), upwelling is weak due to weak wind stress and is closely confined to the coastal belt. This results from the suppressive effect of the southwards-flowing Arabian Sea high saline water on the process of upwelling along this coastal stretch.

ADDITIONAL INDEX WORDS: *Wind stress, mass transport, local temperature anomaly, south-west monsoon, south-eastern Arabian Sea and Kanyakumari coast.*

INTRODUCTION

Upwelling is the slow and persistent rising of nutrient-rich subsurface water towards the ocean surface. Upwelling off the south-west coast of India, as indicated by rapid upwards movement of isotherms, surface cooling, and the associated fall in coastal sea level, occurs during the summer monsoon (south-west monsoon) months from May to September. Though upwelling signals are observed in sea level from February (SHENOI *et al.,* 2005) onwards, the chemical and biological indications of upwelling in the surface–subsurface waters are observed only in association with the commencement of the summer monsoon. With the onset of the south-west monsoon in May, weak-to-moderate upwelling occurs off the Kanyakumari coast and spreads northwards along the coast as the monsoon advances, reaching up to the Goa coast during peak monsoon season (July–August). Though several attempts have been made to explain the phenomenon, a clear picture on the formation and spread of the upwelling process off the Indian coast is still not available. The classical explanation of coastal upwelling describes wind-induced divergence caused by Ekman transport (SVERDRUP, JOHNSON, and FLEMING, 1942). JOHANESSEN, SUBHARAJU, and BLINDHEIM (1981) noted that wind is an important driving force from February onwards and upwelling is associated not only with local wind but also with larger-scale monsoonal (south-west) conditions, which drive the anticyclonic Arabian Sea monsoon gyre. Studies by SHETYE *et al.* (1985); MURALEEDHARAN and PRASANNAKUMAR (1996); NAIDU, RAMESH KUMAR, and RAMESH BABU (1999), and LUIS and KAWAMURA (2002a, 2002b) explain the phenomena as offshore divergence of the alongshore wind stress component. However, using a numerical model, MCCREARY, KUNDU, and MOLINARI (1993) found a large decrease in the shoaling and a decrease in the upper-layer thickness off the west coast of India when switching off the Bay of Bengal winds, which was also observed by BRUCE, JOHNSON, and KINDLE (1994) and SHANKER and SHETYE (1997). According to BAKUN, ROY, and LLUCH-COTA (1998), the strong westerly monsoon winds at the southern extremity of the Indian subcontinent are tangential to the landmass and drive a very strong offshore Ekman transport. This strong upwelling signal should tend to propagate northwards along the Indian coast via the coastally trapped wave mechanism.

In this paper, we have analysed the role of tangential winds off the Kanyakumari coast, the longshore component of the wind stress along the south-west coast, the influence of remote forcing in the form of the coastally trapped Kelvin waves and the Arabian Sea high saline water (ASHSW) in the formation, and the northwards extension of the upwelling process along the west coast of India. Comparisons are made on the basis of upwelling indices derived from sea surface temperature (SST) and wind (Figure 1), vertical velocities from wind and isothermal shift, surface salinity distribution,

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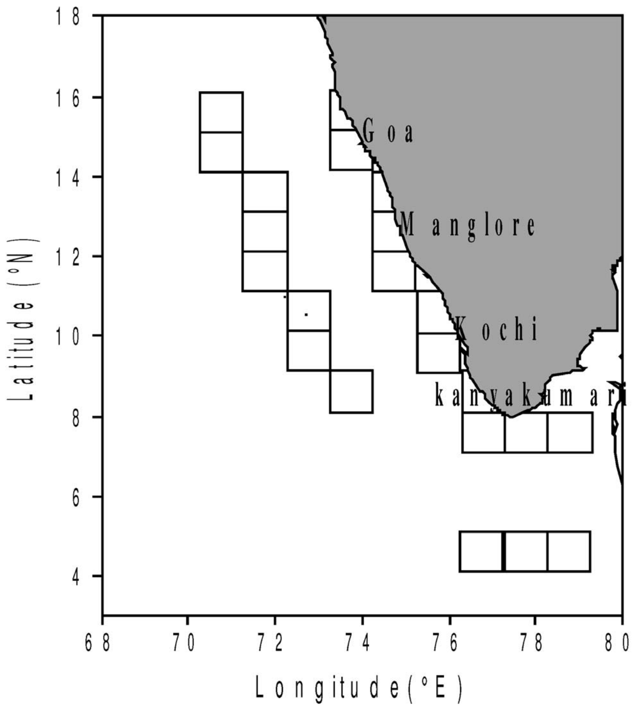

Figure 1. Location map and the 1° × 1° squares used to derive upwelling indices.

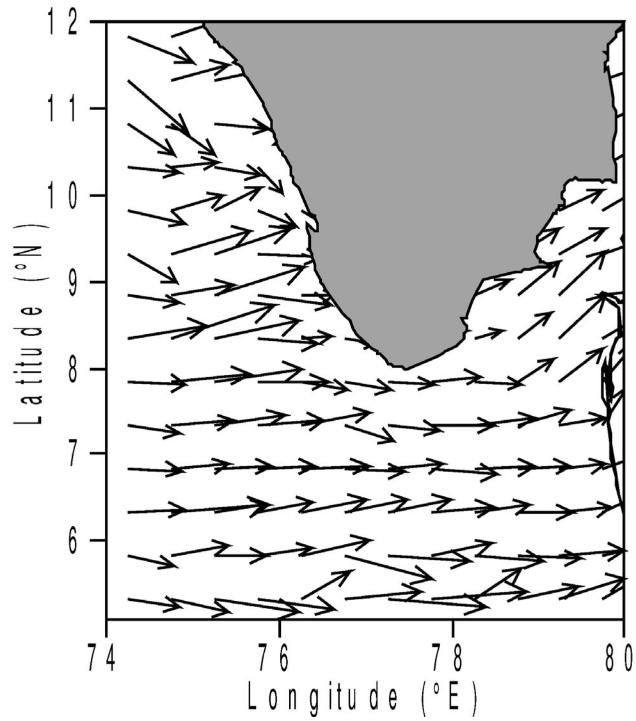

Figure 2. Wind pattern during July 2005 (from QuickScat Scatterometer) showing the tangential winds at the southern tip.

and monthly composite images of surface chlorophyll distribution.

METHODS AND MATERIALS

Upwelling Index from SST

Monthly averages of SST for 11 years (from January 1990 to December 2000) from "a Reynolds-reconstructed SST field" are examined in 1° × 1° latitude–longitude grids for the area 4°30′ to 14°30′ N and 70°30′ to 78°30′ E in the Arabian Sea (Figure 1). Before conducting the analysis, interannual variations in SST in the 11-year data were verified to detect any significant influence of Indian Ocean dipole events (RAO *et al.,* 2002) during 1994 and 1997 and El Nino and La Nina during 1997 and 1998, respectively. Considerable variations (~0.5°C) in SST were observed during 1997 and 1998 (the warm phase and the cold phase, respectively). However, the 11-year averaged data is not expected to have any bias on our analysis, since the anomalies are almost equal and opposite and will thus balance between them. Following the approach of WOOSTER, BAKUN, and MCLAIN (1976); PRELL and STREETER (1982), and NAIDU, RAMESH KUMAR, and RAMESH BABU (1999), the local temperature anomaly (LTA) is calculated as coastal upwelling indices by comparing the coastal and offshore SST. Since the influence of upwelling is known to extend 200–400 km from the coast in the south-west direction (ANTONY, NARAYANA SWAMY, and SOMAYAJULU, 2002), the offshore stations are chosen off 3 degrees to the coastal stations. This is cross-checked with thermal fronts (figure not included) observed during two FORV Sagar Sampada cruises (SS217 during September 2003 and SS237 during August 2005) and found matching except off Goa (15° N), where it is almost four times lesser than that recorded by ANTONY, NARAYANA SWAMY, and SOMAYAJULU (2002). LTA is calculated as

Along the south-west coast, $\mathrm{LTA}_{wc} = T_{\mathrm{lon\text{-}3}} - T_{\mathrm{lon}}$

Off Kanyakumari, $\mathrm{LTA}_{kk} = T_{\mathrm{lat\text{-}3}} - T_{\mathrm{lat}};$ (1)

where T_{lat} and T_{lon} represent the coastal stations between 8.5°–14.5° N and 76.5°–78.5° E, and $T_{\mathrm{lon\text{-}3}}$ and $T_{\mathrm{lat\text{-}3}}$ represent SST 333 km from the coast. LTA_{wc} and LTA_{kk}, respectively, refer to local temperature anomaly along the south-west coast and the Kanyakumari coast. The positive LTA values suggest coastal upwelling process.

Upwelling Index from Wind

A unique wind-forcing pattern occurs over the Indian Ocean, unlike the pattern over the other oceans. The winds blow strongly May–September, during the south-west monsoon, forming the Findlater jet over the Arabian Sea with maximum speeds of about 16 m/s. The winds during the season are generally south-west over most of the Arabian Sea, and they become northerly along the west coast of India. However, off the southern tip of peninsular India, they are from the west, stronger, and tangential to the coast (Figure 2). During the north-east monsoon (November–February) the winds are from the north-east and have maximum magnitude of about 6 m/s. During the transition months, March–April and October, the winds are very weak.

To evaluate the theoretical model's suggestions that a northerly wind along the west coast of the Indian continent drives an offshore Ekman drift together with a surface flow parallel to the coast, Ekman transport (M_x in kg/m/s) is calculated in 1° × 1° latitude–longitude grids for the west coast and the Kanyakumari coast (Figure 1). Monthly averaged wind data (both U and V components) from Comprehensive Ocean Atmosphere Data Sets (COADS Enhanced) for 11 years (in 1° × 1° latitude–longitude grids) from January 1990 to December 2000 are used to understand the role of local wind stress on the coastal upwelling processes. According to the classical square law formula (BAKUN, 1973), the alongshore component of the wind stress off the Kanyakumari and south-west coasts is computed as

$$\tau = \rho_a C_d |U| U \quad \text{off Kanyakumari} \quad \text{and}$$

$$\tau = \rho_a C_d |V| V \quad \text{off the south-west coast;} \qquad (2)$$

where $\rho_a = 1.29$ kg/m^3 is density of air; C_d is the nondimensional drag coefficient, which is a function of both wind speed and stability. In general, C_d increases with wind speed and stability of overlying atmosphere

$$1000C_d = 0.29 + 3.1/U_{10} + 7.7/U_{10}^2 \qquad (3 < U_{10} < 6 \text{ m/s}) \quad \text{and}$$

$$1000C_d = 0.60 + 0.07U_{10} \qquad (6 < U_{10} < 26 \text{ m/s}), \qquad (3)$$

where U_{10} is wind speed 10 m above the sea level.

U and V are the estimated wind vectors near the sea surface with magnitude $|U|$ and $|V|$. Along the west coast of India, the alongshore components are obtained with reference to the inclination of the coastline by adding the coastal inclination to the wind direction. Whereas along the Kanyakumari coast, as the winds are tangential to the coast (Figure 2), no inclination is considered in computing the alongshore wind stress component.

Under Ekman's assumptions of steady-state motion, uniform wind, and infinite homogeneous ocean, the mass transport per unit width of ocean surface is directed 90 degrees to the right (in the Northern Hemisphere) of the wind direction. This is related to the magnitude of the wind stress by $M = \tau/f$, where $f = 2\omega \sin \varphi$ is Coriolis force due to Earth's rotation at latitude φ, and ω = angular velocity of Earth's rotation and is equal to 7.292×10^{-5} radians per second.

In Situ Observations

Temperature profiles from Conductivity–Temperature–Depth (CTD) Profiler (SeaBird Electronics 911 series) and wind data from the automated weather system (AWS) collected onboard *FORV Sagar Sampada* in four cruises (FORV 234, 235, 236, and 237) from May to September 2005 are used to experience the vertical structure and to compute upwelling velocity during the study period.

Spatial variation in upwelling along the coastal region is established from vertical velocity computed from isothermal shift and wind at the six selected transects: Kanyakumari (8° N), Trivandrum (8.5° N), Kollam (9° N), Kochi (10° N), Manglore (13° N), and Goa (15° N). Vertical velocity in metres per day is calculated from the shift in 24°C isotherm between May 2005 and July 2005 for the southern transects up to Kochi. Data collected May–June 2005 and August–September 2005 are used for the Mangalore and Goa transects. Significance of taking average wind velocity during the two observations has been verified by analysing the day-to-day variations in the wind field and correspondingly the changes in the thermal structure. For this, time-series observations conducted along the Kollam (9° N) and Trivandrum (8.5° N) transects 24 June 2006–6 July 2006 were analysed and daily undulations (figure not included) in the thermal section were observed as shallowing and deepening of isotherms according to the variations in wind. Vertical velocity of the wind-driven upwelling is estimated following SARHAN *et al.* (2000)

$$\text{Velocity} = \tau/\rho_W f L \qquad (4)$$

where τ is wind stress, as shown in equation (2); ρ_W is the density of seawater, 1025 kg/m^3; f is Coriolis force, as shown in equation (2), and L is the cross-shore length scale of the upwelling.

Surface salinity values derived from CTD-SeaBird 911 (calibrated against salinity derived from water samples collected simultaneously with a rosette sampler and analysed with guideline 8400 Autosal) collected onboard *FORV Sagar Sampada* during May, June, and September (1997–2003) in different cruises (SS142, SS173, SS203, and SS217) were used to understand the surface salinity distribution during different monsoon months.

Observations from Satellite

Climatology of the monthly (June–September) composite maps of the chlorophyll a, Ocean Colour Level 3-binned product for 4 years (2003–2006) derived from MODIS AQUA at 9-km spatial resolution from Goddard Space Flight Centre (GSFC) National Aeronautics and Space Administration (NASA) are used to describe the productivity patterns in the area.

RESULTS AND DISCUSSION

Upwelling off the Southern Tip (Kanyakumari)

Seasonal variations in LTA from the 11-year average data gives positive LTA values greater than 0.25°C for the Kanyakumari coast (Figure 3a), except for the intermonsoon season (February–April). Peak values are recorded during July and August with considerable horizontal gradient in temperature (LTA = 0.85°C). After August, LTA decreases gradually, reaching near-zero values by February. Ekman mass transport from the mean monthly climatology for the 11 years indicates strengthening of offshore mass transport off Kanyakumari from May to October, with a maximum (8000 kg/m/s) in August (Figure 3b) corresponding to the strong westerly winds that are tangential to the coast during the south-west monsoon.

The influence of wind in driving the upwelling process off the Kanyakumari coast is ascertained by comparing the ob-

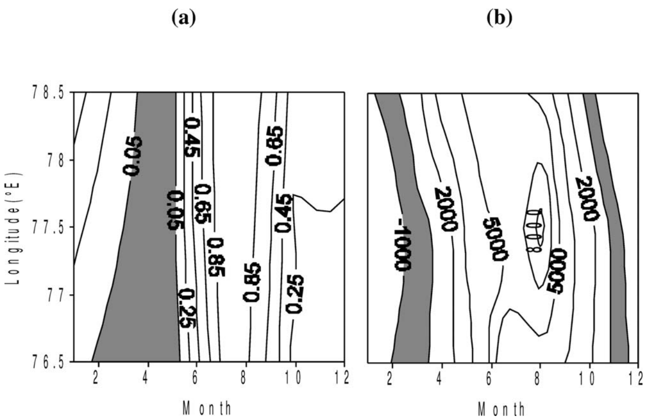

Figure 3. Variation of (a) LTA and (b) Ekman mass transport (kilograms per metre per second) as a function of time (month) and space (longitude) at Kanyakumari. Shaded part indicates negative values.

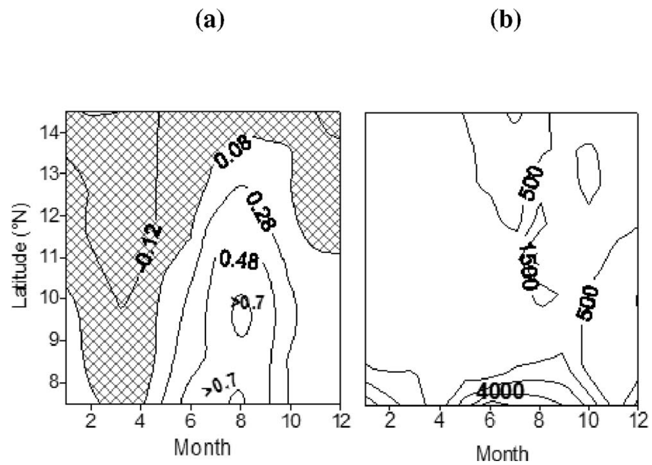

Figure 4. Variation of (a) LTA and (b) Ekman mass transport (kilograms per metre per second) as a function of time (month) and space (latitude) along the south-west coast of India. Shaded part indicates negative values.

served vertical shift of the 24°C isotherm with the vertical velocity estimated from wind. The upslope of the 24°C isotherm at the rate of 1.17 m/d is less than the estimated vertical Ekman velocity of 1.62 m/d. The latter highlights the influence of tangential wind stress.

Upwelling off the South-West Coast

Along the south-west coast of India, LTA values are positive May to October up to 13.5° N (Figure 4a), with the rest of the area north of 13.5° N recording negative values. This indicates the absence of upwelling features in the surface temperature distribution of the area. Figure 4a also depicts local variation in upwelling June–August, with LTA higher than 0.7°C between 9° N and 10° N (between Kollam and Kochi). LTAs weaken northwards, reaching near-zero values at 13.5° N.

However, Ekman mass transport derived from the longshore component of the wind stress gives weak positive values throughout the year between 8.5° N and 14.5° N (Figure 4b). The mass transport is maximum at the south June–July (4000 kg/m/s) and reduces progressively northwards, which is consistent to the observations of Shetye *et al.* (1991); Pankajakshan, Patnaik, and Gosh (1997), and Bakun, Roy, and Lluch-Cota (1998) on the upwelling patterns off the south-west coast.

With a view to understand the observed anomaly between the LTA and the Ekman transport, we compared the vertical shifts in the 24°C isotherm to the vertical Ekman transport from wind stress across selected transects of the west coast, namely, Trivandrum (8.5° N), Kollam (9° N), and Kochi (10° N) for the period from 14 May 2005 to 5 July 2005 and Manglore (13° N) and Goa (15° N) for the period from 4 June 2005 to 3 September 2005. Along the transect off Trivandrum, the upslope of 24°C isotherm occurs at a rate of 0.857 m/d, which is closer to the vertical Ekman transport (0.745 m/d) from the wind. We further applied the best-fit regression equation of LTA and Ekman transport (LTA $= 0.2164 + 0.0002x$) on the wind stress off Trivandrum and found that the estimated LTA (0.552°C) corresponds approximately to the observed LTA (0.561°C). The preceding observations indicate that upwelling along the Kanyakumari–Kollam coast (8° N to 9° N) is driven predominantly by the alongshore component of the wind stress.

North of Kollam (9° N), weak Ekman transport is persistent throughout the year (Figure 4b), with peak values of 1500 kg/m/s from June to July up to 13° N, further north of which it falls to around 500 kg/m/s. However, LTA from the *in situ* SST observations is positive only up to 13.5° N (0.175°C for July), north of which the values are negative. The general trend is that LTA progressively increases towards the south (with a peak value of 0.721°C at 9.5° N in August) and decreases progressively towards north (−0.084°C at 14.5° N in June). All along the coast, LTA is high during July and August, which corresponds to the peak south-west monsoon period. The role of wind stress in maintaining LTA was tested by applying the regression equation (LTA $= 0.2164 + 0.0002x$) for Trivandrum (8.5° N) to the local wind stress for each degree latitude north of 8.5° N. For the area 9° N to 13° N, the derived LTA is significantly lower than the observed LTA. This could explain how only part of the upwelling process is due to wind stress and further implies the role of remote forcing in the sustenance of the observed LTA. We further tested the preceding by comparing the vertical shift of the 24°C isotherm off Kollam and Kochi for May–July with the vertical Ekman transport from wind stress. At both Kollam and Kochi, the isothermal shift was at the rate of 1.1 m/d, whereas the estimated vertical transport was only 0.38 m/d for Kollam and 0.55 m/d for Kochi. These observations suggest that the upwelling process along the area 9° N to 13° N is influenced by the combined effect of local winds and remote forcing.

North of 13° N, the LTAs progressively decreased, reaching negative values at 14° N. The observed LTAs are lower than LTAs derived from wind stress by applying the regression

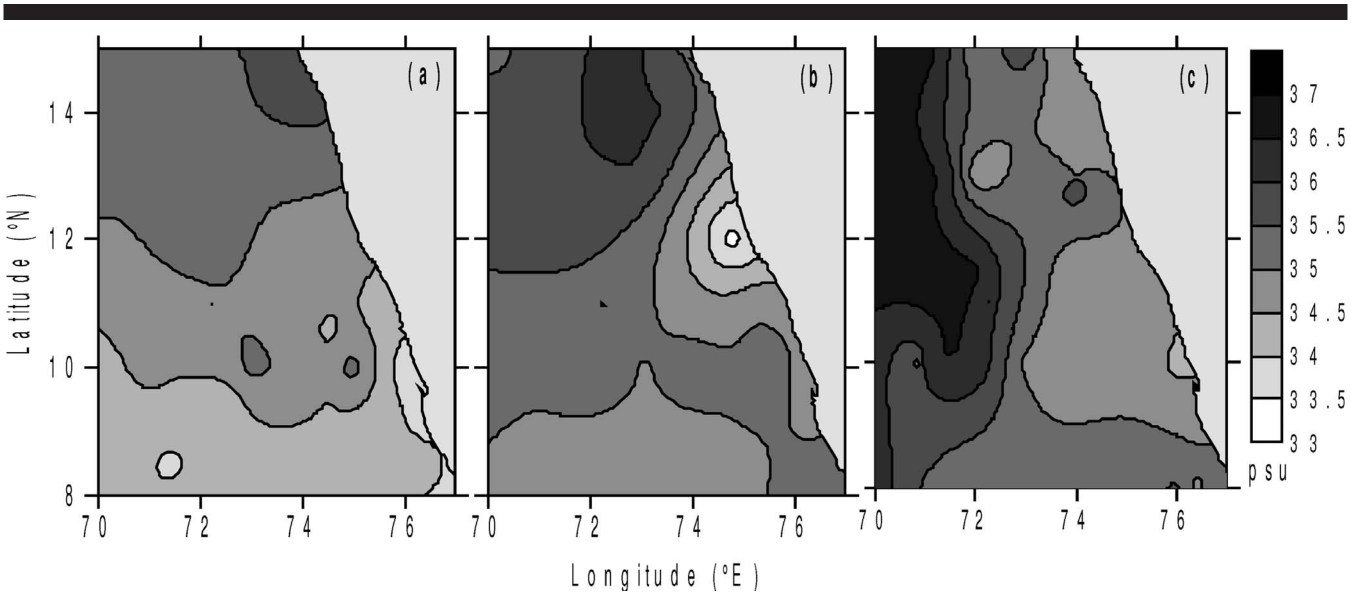

Figure 5. Sea surface salinity for (a) May, (b) June, and (c) September show the southwards propagation of the ASHSW.

equation (LTA = 0.2164 + 0.0002x) for Trivandrum on the local wind stress. The seasonal averages of the observed and derived LTAs are 0.086 and 0.363, respectively, for 13–14° N and 0.052 and 0.288, respectively, for 14–15° N. Similarly, the observed rates of the vertical transport (figure not included) at 13° N (0.35) and 15° N (0.067) are lower than the values derived from wind stress (0.365 for 13° N and 0.24 for 15° N). From this, it is expected that the process of upwelling along this coastal strip is suppressed somehow, despite having winds favourable for upwelling.

Surface Salinity Distribution

Surface salinity plots (Figure 5) for May, June, and September reveal the formation and fate of the upwelled waters along the coast. During May, the south-west monsoon winds are weak and upwelling is confined to a narrow zone between 9° N and 10° N. During June, the south-west monsoon winds are stronger and the upwelling process extends along the coast from 9° N to 13° N. The upwelled waters of the area are characterised (in temperature, salinity, and density values) by the presence of the Bay of Bengal waters on the surface. These waters are transported offshore to a distance of approximately 200 km by the combined effect of Ekman transport and Rossby waves radiated from coastally trapped Kelvin waves (Shanker and Shetye, 1997; Shenoi *et al.,* 2005). Further north, the upwelling is weak and confined close to the coast, possibly due to the suppressive effect (cold, dense, upwelled waters are pushed to the coast by warm, high-saline waters, and its propagation occurs to the south-east) of the southwards-extending ASHSW (Muraleedharan and Prasannakumar, 1996). On the basis of water mass studies, Stramma, Fischer, and Schott (1996) had indicated that water at the shelf edge was a mixture of low-salinity water advected out of the Bay of Bengal around Sri Lanka and higher-salinity Arabian Sea water. Shanker and Shetye (1997) examined the dynamics of the coastal circulation of the region using a one-and-a-half-layer dynamically reduced gravity model and observed that the upwelling Rossby waves radiated by the upwelling, coastally trapped Kelvin waves, with a peak in July, propagating polewards along the Indian west coast, are instrumental in generating the fall in sea level during the active monsoon season. Prasannakumar and Prasad (1999), while describing the ASHSW, had noted that, as the south-west monsoon winds become progressively well established from June to September, the open-ocean circulation reverses its direction with a zonal current towards east that become stronger. They also noted that, under the influence of the south-west monsoon circulation, the high-salinity core is pushed north-eastwards and slowly spreads towards south along the west coast of India. The salinity plots for May, June, and September are consistent to the preceding observations.

Inferences from upwelling indices, vertical velocities, and the surface salinity pattern strongly indicate the absence of remote forcing in the region between 8° N and 9° N. This leads to the conclusion that the Kelvin waves reach the west coast of India, become coastally trapped about 9° N, and propagate polewards as the coastally trapped waves, thereby demarcating a narrow strip of the coast between Kanyakumari and Kollam (8° N to 9° N) as the “shadow zone” (Figure 6). This could be explained in terms of the 2000-m isobath (Figure 6), which may act like a cliff in the path of the wave, deviating it to its optimal energy path. The upwelling process along the shadow zone is thus purely driven by the longshore component of the wind stress, as indicated by the rate of the vertical Ekman pumping. The absence of Bay of Bengal water on the surface layer of the shadow zone further substantiates the preceding view. Upwelling along the area 9° N to 13° N is under the combined influence of wind stress and the upwelling mode of the Kelvin and Rossby waves (Shanker and

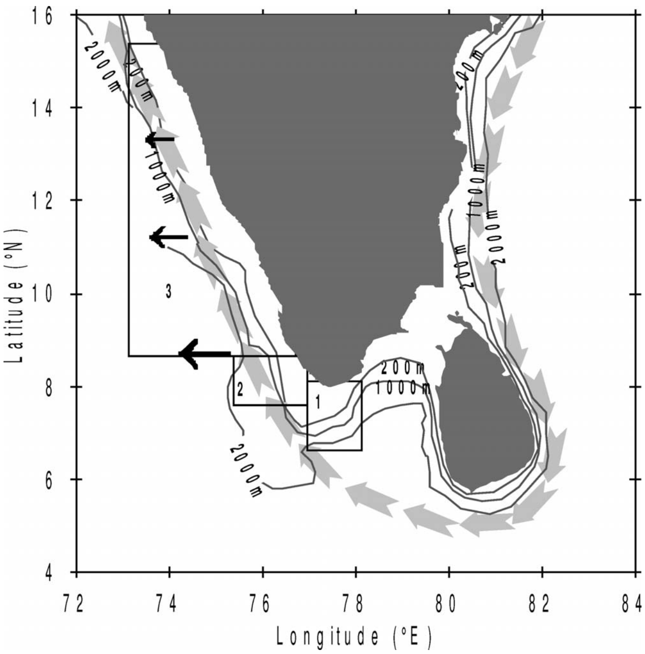

Figure 6. Schematic representation of different upwelling zones classified according to the formation mechanism, as well as intensity. Arrows along the coast represent coastal Kelvin waves. Westwards-directed black arrows depict Rossby waves, the phase velocity of which decreases moving from the equator; upwelling at area 1 is strongly wind driven; area 2 is a shadow zone with weak wind-driven upwelling; upwelling at area 3 is the result of remote forcing, as well as wind stress.

SHETYE, 1997). As a result, moderate-to-intense upwelling occurs along the coastal belt from 9° N to 13° N. The vertical velocity isothermal shift in the area 9° N to 11° N far exceeds the vertical shift computed from wind stress. The upwelled waters have distinct signatures of the Bay of Bengal waters and are transported approximately 200 km offshore by the Rossby waves. Further north (north of 13° N), the intensity and extend of upwelling weakens due to the suppressive effect of the ASHSW and the weakening of the Rossby waves (BRANDT *et al.,* 2002; SHETYE, 2005). However with the advancement of the monsoon, the wind becomes stronger and pushes the ASHSW offshore, resulting in weak upwelling confined close to the coast. At the extreme north (14° N to 15° N), upwelled waters are confined very close to the coast and are thus not reflected in the LTA estimates, perhaps due to the high resolution (1° × 1°) of the data. This explains the occurrence of negative LTA at the extreme north of the study area. Longitudinal variations in SST off the Goa coast (15° N) clearly show that upwelling is confined very close to the coast (Figure 7).

Surface Chlorophyll Distribution

Observations on the upwelling patterns of the south-west coast are further verified using the 4-year (2003–2006) climatology monthly composite images of surface chlorophyll distribution along the study area for the south-west monsoon (Figure 8). During July, surface chlorophyll maxima are seen

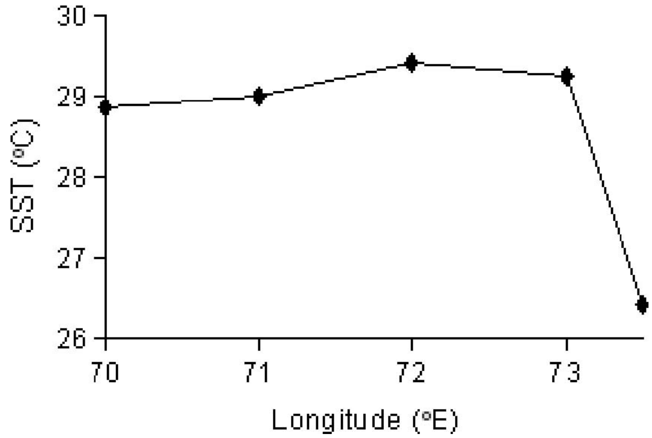

Figure 7. SST (°C) variation along the 15° N transect during September 2003, showing the limited offshore extension of upwelling.

as isolated patches south of Kanyakumari and west off Trivandrum (8.5° N). A narrow stretch of low primary productivity demarcates the shadow zone between 8° N and 8.5° N. With the advancement of the south-west monsoon, the chlorophyll patches extend northwards along the west coast, corresponding to the geomorphology of the coast and the northwards propagation of the monsoon. The offshore extent of the surface chlorophyll spread is maximum between 8.5° N and 9° N, which is south of the peak upwelling area (9° N to 13° N), as inferred from the surface salinity distribution (Figure 5b). The discrepancy should be due to the lag phase between the availability of nutrients and the actual primary production. The West India Coastal Currents (WICCs) transport the surface waters of the west coast southwards, and at the extreme tip of peninsular India, the surface chlorophyll patch is transported equatorwards by the combined action of the offshore Ekman transport and the WICCs. The formation and the fate of the chlorophyll maxima south of the Kanyakumari coast and along the west coast of India are consistent to our observations on the upwelling patterns for the coast.

CONCLUSIONS

Upwelling along the south-west coast and southern tip (off the Kanyakumari coast) are localised phenomenon under the influence of varying forcing mechanisms. South of the Kanyakumari coast, upwelling is strong and is influenced by the south-west monsoon winds that are tangential to the coast. Chlorophyll maxima are recorded in June when the south-west monsoon winds are strongest at this point. Surface chlorophyll is transported equatorwards due to the combined effect of offshore Ekman mass transport and WICCs. Along the west coast, the area between 8° N and 9° N represents the shadow zone, where the role of remote forcing is minimum and the upwelling occurs exclusively by Ekman transport brought out by the longshore component of the wind stress. The area north of this (9° N to 13° N), shows moderate-to-intense upwelling under the combined influence of the wind stress and remote forcing in the form of coastally trapped

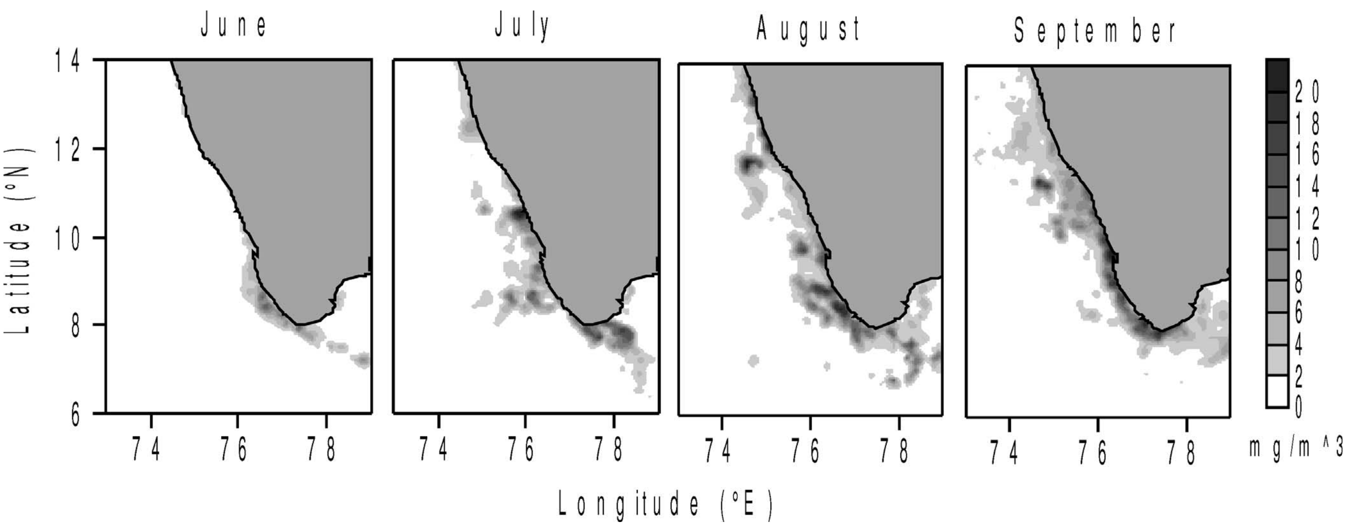

Figure 8. Four-year climatology (2003–2006) of monthly composite surface chlorophyll from MODIS AQUA for June–September.

Kelvin waves and the offshore radiating Rossby waves. Within this area, the offshore propagation of the upwelled waters is maximum at the southern part and progressively weakens northwards, perhaps due to the weakening of the offshore propagating Rossby waves. Further north (13° N to 15° N), the process of upwelling is weak and confined very close to the coast, despite winds favourable for upwelling. This may be due to the presence of the southwards-flowing tongue of the ASHSW, which suppresses the upwelling along the coast. During July and August, the south-west monsoon winds are strong over the area, which may push the ASHSW slightly offshore, leading to weak upwelling confined close to the coast.

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