CHAPTER IV

THERMOHALINE AND CURRENT VARIABILITY OFF VISAKHAPATNAM DURING SUMMER MONSOON SEASON
4.1. Introduction

Western boundaries of the Bay of Bengal and Arabian Sea are unique because of the occurrence of seasonal upwelling and Boundary Currents associated with the seasonally reversing winds. All the major coastal upwelling zones are located at the eastern boundaries of the oceans except in the Bay of Bengal and Arabian Sea. In the western Arabian Sea, the upwelling processes are intense during summer monsoon and can not be explained in terms of local wind forcing alone (Schott, 1983). Remote forcing by offshore windstress curl and southern hemispheric trades are also the factors to be taken into consideration (McCreary and Kundu, 1985). In contrast to this, evidence is available for upwelling in the western Bay, driven by local winds alone as seen in the eastern boundaries of the oceans (LaFond, 1957; Murty and Varadachari, 1968; Unnikrishnan and Bahulayan, 1991; Shetye et al., 1991; Johns et al., 1992; Rao et al., 1993). In general, in the Bay of Bengal, upwelling takes place during February to August. With the onset and progress of the summer monsoon, the massive river discharge into the head of the Bay, dampens the upwelling in the northern Bay (Gopalakrishna and Sastry, 1985; Shetye et al., 1991; Johns et al., 1992; Rao et al., 1993). When the monsoon winds recede, sinking takes place all along the western boundary starting from the northern Bay and extending to the southern Bay. This process, continues from August to December (LaFond, 1958b; LaFond and LaFond, 1968).

Other than upwelling, the major seasonal feature in the western boundaries of the north Indian Ocean is the Western Boundary Currents (WBCs). Poleward WBCs (Somali
Current) appear in the Arabian Sea during the summer monsoon; whereas it occurs in the Bay of Bengal during January to May (Chapter – VI).

In addition, the western boundary of the northern Bay experiences tidal currents, high amplitude internal waves and fresher water inflow (LaFond and LaFond, 1968; LaFond and Moore, 1972; Shetye et al., 1991).

Under the influence of the above mentioned physical processes, significant seasonal and short-term variability in the oceanographic characteristics is to be expected in the western Bay of Bengal. Some studies were carried out to understand the seasonal and spatial variability in the western Bay (Ganapatı and Murthy, 1954; LaFond, 1954; 1958a; 1958b; Rao, 1956b; LaFond and Rao, 1954; Ramasastry and Balaramamurty, 1957; LaFond, 1972; LaFond and Moore, 1972; Shcherbinin et al., 1979; Gopalakrishna and Sastry, 1985; Rao et al., 1986; Rao et al., 1987; Raju et al., 1992; Rao and Sadhuram, 1992; Suryanarayana and Rao, 1992; Banse, 1990; Babu et al., 1991; Shetye et al., 1991; 1993; Kumar et al., 1992). However, information on the short-term variability in the current and thermohaline fields are sparse. Previously, few direct current measurements were made as follows: (i) for few hours off Godavari during September 1980 (Sarma and Rao, 1986), (ii) in the vicinity of Krishna river mouth during June/July 1982 (Antony et al., 1985), (iii) off the rivers Krishna, Godavari and Mahanadi and off New Moore island during June/July 1982 (Rao K. H. et al., 1987) and (iv) off Visakhapatnam for every month for a tidal cycle each (Rao P. B. et al., 1987). Apart from these studies, no information is available on the observed current structure from the
continental shelf region of the western Bay of Bengal either seasonally or synoptically. It is to be noted that all those measurements were made by hanging sensors from onboard platforms. Moreover, the informations on the short-term variability in the thermohaline field for longer than one full tidal cycle is practically nill.

To partially augment our knowledge on the short-term variability, a cruise was conducted off Visakhapatnam during the summer monsoon of 1986 (Sanil Kumar et al., 1989). The experimental site off Visakhapatnam was selected because of the reasons that it is almost at the middle of the east coast of India and the wind field with respect to the coastline is stronger and highly favourable for upwelling during summer (Murthy et al., 1988).

In this investigation, an attempt is made to study the short-term variability in the air-sea interaction and thermal structure at a deep and a shallow station off Visakhapatnam during the summer monsoon. The mixed layer response to the atmospheric forcing is simulated using KTDM and NK Models. At the shallow station, currents, thermohaline variability and the associated mixing characteristics in the water column are studied using moored vector averaging current meters at different levels.

4.2.1. Data and methodology

Twelve stations were occupied in the coastal zone off Visakhapatnam during cruise No. 170 of RV Gaveshani from 23 to 30 June 1986 (Fig. 4.1). At each station subsurface temperature profiles were taken using MICOM BT (TSK, Japan. Accuracy: temperature - 0.05°C; depth - 3 m for 100 m depth)
Fig. 4.1. Stations covered during June/July 1986 off Visakhapatnam.
while Nansen casts provided data on temperature and salinity at discrete depths. In addition to this, hourly subsurface temperature and standard marine meteorological observations were made at 17°20'N and 84°20' (depth >2000 m; deep station - designated as ST1) from 24 to 27 June 1986 and at 17°45'N and 83°50'E (depth ~90 m; shallow station - designated as ST2) from 27 to 30 June 1986. At the shallow station, 4 Aanderaa current meters (accuracy: speed - 4 cm/s; direction - 5°; temperature - 0.05°C and salinity - 0.03 PSU) were moored for recording temperature, salinity and current speed and direction at every 10 minutes interval. The current meters were moored in such a way that 2 were positioned in the mixed layer (10 to 25 m), one in the pycnocline (50 m) and the last one close to the bottom (85 m). Stations perpendicular to the coast on either side of the mooring location are referred as section A and B.

4.2.2. Power Spectra: -

A Fast Fourier Transform (FFT) algorithm following Ahmed and Natarajan (1983) was used to compute power spectral energies embedded in different bands of offshore (U), alongshore (V), temperature (T) and salinity (S) fields. Due to the restriction on the number of data points to be used in an FFT (fast fourier transform) algorithm, data length was limited to 64 h (384 points at 10 min interval). The data were smoothed by a 3 point moving average to remove the frequencies greater than 1/30 cycles/min. The procedure adapted for computing power spectra is given in chapter 3 (3.2.2). First set of spectral values was computed from the data points 1 to 256 and the second set from 127 to 384.
points by allowing 50% overlapping. The resultant two sets were averaged over the same frequencies and presented in this chapter as logarithmic plots of frequency versus spectral density.

4.2.3. Vertical shear (S):—

S represents the dynamic stability and is given by the equation

\[
S^2 = \left( \frac{\Delta U}{\Delta Z} \right)^2 + \left( \frac{\Delta V}{\Delta Z} \right)^2
\]  

(4.1)

where \( \Delta U \) and \( \Delta V \) are the differences in the offshore and alongshore components of the flow.

4.2.4. Richardson Number (Ri):—

A dimensionless quantity, Richardson number

\[
Ri = \frac{N^2}{S^2}
\]  

(4.2)

measures the relative importance of both density (static stability) and mechanical (dynamic stability) effects in mixing. The transition from waves to turbulence occurs (Munk, 1966; Phillips, 1966) when \( Ri \) is less than 0.25.

Mixed layer simulation was carried out using the methodology described in Chapter II and the Brunt Vaisala frequency was computed using Eq. 3.1.

4.3. Results and Discussion

4.3.1. Thermal structure on either side of the stationary locations:—

Vertical sections of temperature along sections A and
The surface temperature was low (~26°C) near the coast along both sections and increased to over 28°C offshore. Throughout the upper 100 m water column upsloping of isotherms towards the coast indicated upwelling. Upwelling starts off Visakhapatnam by February and continues up to June/July under the favourable winds (LaFond, 1954; 1957; LaFond and LaFond, 1968; Murty and Varadachari, 1968; Rao et al., 1988; Shetye et al., 1991). Observational evidence is reported on the occurrence of spatial gradient of > 3°C in temperature within 30 to 40 km wide upwelling band due to upwelling off Visakhapatnam during summer monsoon (Murty and Varadachari, 1968; Rao et al., 1988; Shetye et al., 1991). By July/August, the massive fresh water discharge at the head of the Bay inhibits the upwelling processes (Gopalakrishna and Sastry, 1985; Johns et al., 1992; Rao et al., 1993).

4.3.2.1. Observations at stationary locations:

Surface meteorological elements at deep (24 to 27 June 1986) and shallow (27 to 30 June 1986) stations for the observational periods are shown in Fig. 4.3. In general, oscillations with varying amplitudes and periods were prominent in all the elements at both the stations. The surface pressure (PR) fluctuations were comparatively more periodic (at semi-diurnal periodicity) at ST2. The maximum drop of 8 mb occurred in the pressure field at ST2 on 24 June 1986. The speed (FF) of the southwesterly wind was higher than 10 m/s most of the time and had crossed 15 m/s occasionally. Cloudy (>5 octa) to overcast (>7 octa) sky was seen during the entire observational period except on 30 June.
Fig. 4.2. Vertical sections of temperature along sections A and B.
Fig. 4.3. Time series of meteorological elements at the deep (ST1) and shallow (ST2) station locations.
1986 at ST2. On 30 June, the sky became clear and the air (DB), wet bulb (WB) and sea surface temperature (SST) increased considerably. DB, WB and SST were lower at ST2 compared to those of ST1. The low temperature (~ 26°C) at the shallow station was due to the existing upwelling as evidenced from Fig. 4.2.

Time series of hourly heat budget components ($Q_I$, $Q_E$, $Q_S$ and $Q_B$) and net heat flux ($Q$) are shown in Fig. 4.4. The day time peaks of $Q$ were less than 800 W/m$^2$ due to the massive cloud cover except on 30 June 1986 when the peak reached up to 1200 W/m$^2$. The minimum of the peak was ~500 W/m$^2$, observed on 26 June 1986. Thus the $Q$ showed wide range in day to day variation with the diurnal maxima between 500 and 1200 W/m$^2$. It can be seen that the $Q$ followed mainly $Q_I$ as expected for the tropical oceans. Maximum differences between $Q$ and $Q_I$ were observed on 24 and 26 June 1986 when the heat loss due to $Q_E$ was maximum. The other two terms ($Q_S$ and $Q_B$) had not influenced much on $Q$.

Temporal variability in the thermal field at TS1 and TS2 is inferred from Fig. 4.5. An important observation at TS1 was the appearance of warmer water (>28°C) in the surface layers with an inversion (29°C) at 20 m. Since the heating due to surface heat flux starts from the surface, the inversion layer appeared at subsurface has to be viewed as the manifestation of the advected warmer water. Also it has to be noted that even though $Q$ was more on 24 June 1986, the corresponding warmer layer was absent. These observations suggest the one dimensional vertical heat flux was not the only major controlling factor of the thermal variability at the deep station. The isotherms at shallow station was found
Fig. 4.4. Time series of heat budget components at the deep (ST1) and shallow (ST2) station locations.
Fig. 4.5. Temperature-time sections at the deep (ST1) and shallow (ST2) station locations.
to be pushed upwards by 25 to 30 m when compared with those of at deep station. For example, 25°C isotherm was around 80 m at deep station; whereas it was around 35 m at the shallow station. This uplifting of the isotherms can be attributed to the prevailing upwelling (Fig. 4.3). The wavy isotherms with amplitudes of 15 to 20 m indicated the influence of internal tides.

The residual surface wind and subsurface currents at 10, 25, 50 and 85 m for the entire observational period at the shallow station are presented as the vector averaged plots (Fig. 4.6). The wind direction was southwesterly with an average speed of ~12 m/s. The current at 10 m was southeasterly and was stronger (~41 cm/s) with respect to the currents at deeper levels. The direction of residual flow changed to southwesterly with a sizable reduction in speed (~10 cm/s) in the pycnocline (50 m depth). This reduction in speed might have resulted from the weak vertical transport of wind induced momentum into the stratified pycnocline. In the present study the clockwise rotation and diminishing magnitude of current vectors (when looked from top) clearly suggest the influence of wind induced Ekman type circulation. As expected the currents at 85 m were weak due to bottom frictional effects. The direction of surface currents fairly agreed with the derived currents from wind field utilizing a barotropic model (Bahulayan and Varadachari, 1986). Those currents showed northeasterly flow during June with a speed of 30 cm/s. During the summer monsoon of 1978, the surface currents derived off Visakhapatnam from density distribution also showed a northeasterly component. But, south of Visakhapatnam no well-defined field of motion could be seen.
Fig. 4.6. Vectorially averaged surface wind and subsurface currents at shallow station (ST2). Both the wind and current vectors start from the origin.

Fig. 4.7. Stick plots of surface winds and subsurface currents at ST2. Both the wind and current vectors start from the origin.
(Gopalakrishna and Sastry, 1985). However, the geostrophic currents during July/August 1989 (Shetye et al., 1991) revealed a northerly current very near to the coast and a southeasterly current offshore. The southeasterly current was observed as a part of the cyclonic eddy in the offshore region. However, the direct reading current meter data off Visakhapatnam during June 1980 (Rao et al., 1987) showed a varying current between northwesterly to southwesterly. The ship drift data (Cutler and Swallow, 1984) also show a northeasterly flow. In short, the residual current observed in the surface layers in the present study agreed with the derived flow pattern of the earlier studies.

The hourly surface wind and current at 20 min interval at 10, 25, 50 and 85 m depths are shown as sticks (Fig. 4.7) for delineating the short-term variability in the subsurface current and the surface wind fields. The above data were smoothed with a 3 h moving average to explain clearly the well organized low frequency oscillations. The wind field was mostly southwesterly with speeds between 4 and 16 m/s. The currents in the mixed layer were southeasterly with fluctuations of speed from 15 to 45 cm/s. Current speed in the pycnocline was reduced and superposed with oscillations around 12 h periodicity. This 12 h periodicity indicated the importance of the internal tides. In the mixed layer the oscillations with 12 h periodicity were not conspicuous under the influence of mixing. The flow near the bottom was very weak due to friction.

Time series plots of temperature, salinity, U and V components of currents (all obtained from the Aanderaa current meters) are shown in Fig. 4.8 to explain the

62
Fig. 4.8. Time series of temperature (T), salinity (S) and offshore (U) and alongshore (V) components of currents
Fig. 4.9. Spectra of temperature (T), salinity (S), and offshore (U) and alongshore (V) components of currents.
short-term variability and the interrelationship between these parameters. Under the influence of upwelling, the temperature and salinity showed a decrease and increase respectively in the entire water column during all the 3 days. The temperature fluctuated between $25.3^\circ C$ and $26.3^\circ C$ with a decreasing trend in the surface layer. A marked decrease of $2^\circ C$ in temperature was observed within the observational period at 85 m. The maximum amplitude of temperature and salinity variability was observed at 50 m due to strong gradients in the pycnocline. Temperature varied between $23.5^\circ C$ and $28.5^\circ C$ and salinity between 33.8 and 34.2 PSU at this depth. At all these depths an inverse relationship between temperature and salinity was observed. Positive values of $U$ at 10, 25 and 85 m indicated offshore flow as a consequence of upwelling. But at 50 m, $U$ was negative during most of the period. Southeasterly components ($V$ negative) were seen during most of the observational period at all depths with exceptional appearance of positive $V$ at 50 m occasionally. The occurrence of offshore flow at 10 and 25 m and onshore flow at 50 m with decreasing trend in temperature and increasing trend in salinity confirmed the upwelling process in the water column. At 50m (pycnocline), the internal waves also contributed to variation.

4.3.2.2 Spectral analysis:

The prominent frequencies embedded in the offshore, alongshore, temperature and salinity fields were identified (Fig. 4.9). In all the elements at 10, 25, and 50 m, maximum energy was found to be concentrated between the frequencies 0.00117 cpm (14.2 h) and 0.00156 (10.7 h). These frequencies
are nearer to semi-diurnal tidal frequencies ($S_2 = 12$ and $M_2 = 12.4$ h). Because of the low resolution of the spectra in this analysis, the frequencies correspond to 12 and 12.4 h reflect in the frequency band 0.00117 to 0.00156 cpm. It is well known that the semi-diurnal tide is one of the major generating mechanisms to explain these periodic oscillations in the water column, especially in the coastal waters off Andhra coast (LaFond and Moore, 1972). Even in the north central Bay of Bengal (upto ~550 km away from the coast) tidal dominance was observed in the current structure (Chapter III). It is therefore concluded that the oscillations found in the frequency band, 0.00117 to 0.00156 cycles/min were generated due to tide. The wide spacing of the spectra of temperature and salinity between 25 and 50 m was due to the fact that the energy level was ~100 times greater at 50 m compared to 10 and 25 m. This is attributed to the greater amplitude of the internal waves in the temperature and salinity fields in the pycnocline. The different shape of spectra at 85 m implied the different forcing function at this level compared to those of the upper layers. In general, the energy peaks in the semi-diurnal tidal frequency were weak at 85 m due to frictional effects which damp most of these natural oscillations.

4.3.2.3.Mixing characteristics of the water column:-

The mixing characteristics of the water column are determined by density stratification and vertical momentum exchange through shear instability. Since the current meter data were available at greater depth intervals, values of $R_i$ can differ much for individual levels. So in this Chapter, $R_i$
is used mainly to explain the relative importance of density stratification and shear flow in the mixing characteristics. Time series of $R_i$, $N$, and $S$ for 3 different layers viz. 10-25, 25-50 and 50-85 m representing surface, middle and bottom layers respectively, are given in Fig. 4.10. Since the energy in the high frequency band was negligibly small, the data were subjected to a 3 h moving average to bring out the major features. On an average $N$ was increasing towards the subsurface layers due to increasing density gradient. In the surface layer, $N$ varied between 0.6 and 2.6 cycles/h. But in the bottom layer $N$ ranged from 4.7 to 8.3 cycles/h. Since the strength and variability in the current field were stronger in the surface and middle layers, $S$ was also higher (0 to 0.02 sec$^{-1}$) in those layers with larger amplitude of variability. At the bottom layer, $S$ ranged between 0.02 and 0.06 sec$^{-1}$. In general, the $R_i$ was very low (<0.25, unstable layer) in the surface layer during the observational period. In the middle layer $R_i$ dropped occasionally below 0.25 when the shear increased. The spells of instability occurred on 27 June 1986 night, between 28 June 1986 afternoon and 30 June 1986 and towards the end of observational period. During these periods, the current vectors at 50 m was almost opposing the current vectors at 25 m (Fig. 4.7). The observations showed that when the periodically oscillating currents (tidal periodicity) at thermocline opposed the stronger surface layer flow, sufficiently larger current shear was produced to bring down $R_i$ below 0.25 across the mixed layer base. Presumably the vertical mixing between the pycnocline and mixed layer was enhanced by internal tides in a transient manner depending upon the phase of near-surface.
Fig. 4.10. Time series of Richardson number (Ri), Brunt Vaisala Frequency (N) and vertical current shear (SR) at TS2.
offshore flow and subsurface onshore flow. This unstable water column (in the vertical direction) might have caused to decrease the temperature and to increase the salinity in the surface mixed layer (Fig. 4.8) between 28 June afternoon and 30 June 1986. The Ri in the bottom layer indicates that the water column was dynamically stable.

4.3.2.4. Simulation of mixed layer characteristics:

Following the methodology described in chapter II, a number of simulations were carried out with different combinations of $m_1$ and $m_2$ with NK and $'\gamma'$ with KTDM models. The most suitable combination of $m_1$ and $m_2$ for both the stations (ST1 and ST2) was found when $m_1 = 0.39$ and $m_2 = 0.48$. This combination was selected from Davis et al. (1981). The most representative value of $'\gamma'$ in KTDM was found when it was 0.0012 equivalent to Secchi disc depth 14 m. The results of these simulations along with the observed MLD and MLT are shown in Fig. 4.11. Observed MLD fluctuated within a wide range between 5 and 50 m. The shoaling/deepening of MLD corresponding to diurnal heating/cooling reflected at shallow station, but was not evident at deep station. The observed MLT also exhibited greater variability in the range 27.5°C and 28.8°C at TS1 compared to TS2. The diurnal heating/cooling was also not clearly resolved at both the stations. The results of the simulation showed that both NK and KTDM performed well in simulating MLD at the TS2 but poorly performed at TS1. The excursions in the simulated MLD were larger in the case of NK compared to KTDM because of the inclusion of the shear production term in NK. The oscillations in the observed temperature field were also not
Fig. 4.11. Time series of mixed layer depth (MLD) and mixed layer temperature (MLT),
simulated well with both the models. The reasons for the wide departure between the observed and simulated MLD and MLT have to be viewed in terms of likely influence of offshore advection of upwelled cold water. If the surface heat exchange processes (buoyancy effect) are dominant in the variation of the thermal structure, an increase/decrease is expected in SST and thermal field with respect to the values of Q only. But from the above discussions, it was clear that such correlation was not revealed in the present case. Also the influence of the internal tide on the instability of the water column in the mixed layer was proved (spells of instability occurred on 27 June 1986 night, between 28 June 1986 afternoon and 30 June 1986 and towards the end of observational period) from Fig. 4.10 (section 4.3.2.3). Therefore, the greater departure between the observed and simulated MLD occurred during the 'spells of instability' at the shallow station using NK is obvious. This observation prompted to accept the better results of KTDM at shallow station with caution. The above observations suggest the physical processes other than the one dimensional processes (viz. upwelling, internal tide etc.) were dominant off Visakhapatnam during the observational period. At present those processes are not included in the 1D models. Thus the above result, necessitates a comprehensive study on the impact of oceanic internal processes in mixed layer dynamics.