CHAPTER III

SALINITY AND CURRENTS IN THE NORTHERN BAY OF BENGAL DURING THE SUMMER MONSOON EXPERIMENTS
3.1 Introduction

The northern Bay of Bengal (north of 15°N) deserves special attention during summer monsoon season (June to September). More than 90% of the total river input into the Bay of Bengal, takes place at the boundaries of this water body (Table 1.3). This much water can raise the sea level of the Bay by 0.34 m (Varkey and Sastry, 1992). The whole water spreads over the southern Bay and dilutes the upper layers. As a result, a monotonous increase of salinity occurs towards the southern Bay and towards the subsurface layers. Sharp salinity fronts can also form at different locations of the Bay during different periods (Amos et al., 1972). In addition to the spatial variability, temporal variability on a synoptic scale is also expected to be very conspicuous due to the variability in river discharge resulting from energetic pulses of monsoon rainfall (Pisharoty, 1964; Krishnamurti and Balme, 1976). This variability in salinity can affect tropical ocean models (Cooper, 1988), mixed layer characteristics (Sprintall and Tomczak, 1990) and other surface layer dynamics (Toole and Raymer, 1985; Harenduprakash and Mitra, 1988; Shetye, 1993). So the understanding the salinity variability in the northern Bay of Bengal becomes very important. However, we had no field observations to study the variability of the salinity field on shorter temporal scales prior to 1977.

It is well known that the circulation in the northern Bay of Bengal is very much complicated due to the fresh water discharge (Shetye, 1993), seasonal reversal of winds (Shetye et al., 1993) and remote forcing (Potemra et al., 1991; Yu et al., 1991; McCreary et al., 1993). So the indirect methods
like geostrophic computations, wind driven models etc. may not be appropriate, especially during the summer monsoon season. Only the direct measurements using the moored instruments would provide an accurate picture on the prevailing circulation. However, no direct current measurements were made in the northern Bay of Bengal prior to 1977.

Though the northern Bay of Bengal possesses very complex oceanographic environment, no systematic field observations had been conducted to probe the synoptic scale oceanographic and atmospheric variability until 1977. During 1977 and 1979, as a part of the First Global GARP (Global Atmospheric Research Programme) Experiment, field observations were conducted in the north Indian Ocean known as MONSOON-77 and MONEX-79 (designated as M-77 and M-79 respectively in the following discussion). During both the expeditions former USSR deployed four-ship stationary polygons in the northern Bay of Bengal to collect time series measurements of temperature, salinity and currents. Utilizing these data sets, thermal structure and its response to atmospheric variability was studied in great detail earlier (Anto et al., 1982; 1985; Rao et al., 1981; 1983; 1985; 1987; Rao and Rao, 1986; Rao, 1987; Rao and Mathew, 1988). In this chapter, the focus is mainly laid to investigate the observed salinity and current variability.

3.2. Data and methodology

The stations occupied by the ships during both the experiments and the respective observational periods are given (Fig. 3.1 and Table 3.1). The approximate physical
Fig. 3.1. Stations occupied during the Summer Monsoon Experiments in the northern Bay of Bengal.

Fig. 3.2. Temperature (T), salinity (S) and sigma-t ($\sigma_t$) for polygons of M-77 and M-79.
separation between respective corners of each polygon was of
the order of about 60 nautical miles. Three or six hourly
time series of Nansen cast data at standard depths (during
M-79 data were reported at irregular depths), one hourly
standard surface meteorological elements, half hourly
currents at selected depths and daily totals of rainfall are
the main inputs to this study. Vertical profiles of salinity
at 5 m interval were interpolated (Borowski and Goulat, 1971)
and used in the vertical sections as the daily averages for
the top 200 m water column. The accuracy of salinity and
temperature measurements were 0.003 PSU and 0.02°C.

Current

TABLE 3.1: Station locations and the observational periods
during the summer monsoon experiments in the northern
Bay of Bengal.

<table>
<thead>
<tr>
<th>Station designation</th>
<th>Location</th>
<th>Observational period</th>
</tr>
</thead>
<tbody>
<tr>
<td>N77</td>
<td>19.0 N &amp; 89.0 E</td>
<td></td>
</tr>
<tr>
<td>E77</td>
<td>17.0 N &amp; 91.0 E</td>
<td>11 - 18 August 1977</td>
</tr>
<tr>
<td>S77</td>
<td>15.0 N &amp; 89.0 E</td>
<td></td>
</tr>
<tr>
<td>W77</td>
<td>17.0 N &amp; 87.0 E</td>
<td></td>
</tr>
<tr>
<td>N79</td>
<td>18.0 N &amp; 89.5 E</td>
<td></td>
</tr>
<tr>
<td>E79</td>
<td>16.2 N &amp; 91.1 E</td>
<td>11 - 23 July 1979</td>
</tr>
<tr>
<td>S79</td>
<td>14.4 N &amp; 89.5 E</td>
<td></td>
</tr>
<tr>
<td>W79</td>
<td>16.2 N &amp; 87.7 E</td>
<td></td>
</tr>
</tbody>
</table>

meter data were collected at 25, 50, 100, 150 and 200m depths
at half hourly intervals for durations of the order of 1-2
weeks. These data were reported at a resolution of 1° in direction and 1 cm/s in speed. The corresponding surface wind data at 10 m height were also collected at one hourly intervals. In the following discussion the stations at northern, eastern, southern and western corners of the polygon are designated as N, E, S and W respectively.

3.2.1. Brunt-Vaisala frequency (N):-

N is a measure for static stability and is defined as

\[ N^2 = \frac{-g\Delta \rho}{\rho \Delta Z} \]  (3.1)

where \( g \) is the acceleration due to gravity, \( \Delta Z \) is the spacing between the current meters, \( \Delta \rho \) is the density difference and \( \rho \) is the average density of the layer between the sensors. The water column is considered neutral for zero values of \( N \), stable for positive values and unstable for negative values of \( N^2 \).

3.2.2. Rotary Spectra :-

Observed current records usually consist of harmonic oscillations of different frequencies. Spectral analysis is an effective tool to delineate the frequency components embedded in these data. This technique can be used for both scalar and vector series. But, for vector series, because of the earth's rotation, an asymmetry in the Fourier components of the horizontal plane can be expected. In the rotary spectral method, this asymmetry will be resolved as the difference in energy levels in clockwise (\( S_+ \)) and anticlockwise (\( S_- \)) spectra (Gonella, 1972). Therefore the
rotary spectral method is more appropriate for the analysis of current meter data.

Any discrete time series data, \( U(t) \) for a vector length 'j' varies from '1' to 'm' and sampled at 'T' time interval can be transformed into Fourier series as follows:

\[
U(t) = \sum_{j=1}^{m} (a_j \cos \omega_j t + i b_j \sin \omega_j t) \quad (3.2)
\]

where 't' is time, \( \omega_j = \frac{2\pi j}{mT} \), and \( a_j \) and \( b_j \) are the coefficients representing amplitude of the wave component of frequency \( \omega_j \). 'a_j' and 'b_j' can be found out using the formulae given below:

\[
a_j = \sum_{j=0}^{m-1} x(t) \cos \omega_j t \quad (3.3)
\]

\[
b_j = \sum_{j=0}^{m-1} x(t) \sin \omega_j t \quad (3.4)
\]

But a vector series, \( U(t) \) can be split up into two scalar series of \( u_1(t) \) and \( u_2(t) \) in a horizontal rectangular cartesian system. These scalar series are converted into two Fourier series. Thus we get

\[
u_1(t) = \sum_{j=1}^{m} (a_{1j} \cos \omega_j t + i b_{1j} \sin \omega_j t) \quad (3.5)
\]

\[
u_2(t) = \sum_{j=1}^{m} (a_{2j} \cos \omega_j t + i b_{2j} \sin \omega_j t) \quad (3.6)
\]
Then
\[ P_{u_1u_1}(f_j) = \langle a^2 + b^2 \rangle \text{ autospectra for } u_{1j} \] (3.7)

\[ P_{u_2u_2}(f_j) = \langle a^2 + b^2 \rangle \text{ autospectra for } u_{2j} \] (3.8)

\[ P_{u_1u_2}(f_j) = \langle a_{1j} a_{2j} + b_{1j} b_{2j} \rangle \text{ cross spectra} \] (3.9)

\[ Q_{u_1u_2}(f_j) = \langle a_{1j} b_{2j} - a_{2j} b_{1j} \rangle \text{ quadrature spectra} \] (3.10)

where the symbol \( \langle \rangle \) represents the average over all the segments of vector length 'm' and \( f_j \) is the frequency \( f_j = j/mT, \ 0 \leq j \leq m-1 \).

Gonella (1972) derived the following relationship utilizing the above spectra (equations 3.5 to 3.8):

\[ S_-(f_j) = \frac{1}{8} \{ P_{u_1u_1}(f_j) + P_{u_2u_2}(f_j) - 2Q_{u_1u_2}(f_j) \} \] (3.11)

\[ S_+(f_j) = \frac{1}{8} \{ P_{u_1u_1}(f_j) + P_{u_2u_2}(f_j) + 2Q_{u_1u_2}(f_j) \} \] (3.12)

\[ S_t(f_j) = S_-(f_j) + S_+(f_j) = \frac{1}{4} \{ P_{u_1u_1}(f_j) + P_{u_2u_2}(f_j) \} \] (3.13)

where \( S_, S_+, \) and \( S_t \) are the clockwise, anticlockwise and total spectra respectively.

3.2.2.1. Procedure adapted in the spectral analysis:

The procedure adapted in the spectral analysis in this study is based on Ahmed and Natarajan (1983). The data set was centered to a zero mean by subtracting the mean of the record from each data point. For example
\[ \bar{u}_1(j) = u_1(j) - \frac{1}{m} \sum_{j=1}^{m} u_1(j) \]  

(3.14)

where \( \bar{u}_1(j) \) is the zero mean series of \( u_1(j) \). This new series was low pass filtered with a cut off frequency of 0.08 cph (12.5 h) to remove frequencies greater than that of semi-diurnal tide. Then each \( \bar{u}_1(j) \) was multiplied by a window sequence \( w(j) \) to combat a phenomenon known as 'leakage'. In this analysis Hanning window is used.

\[ w(j) = 0.5 \left[ 1 - \cos \left( \frac{2\pi j}{m-1} \right) \right] \]  

(3.15)

These processed series were used for the Rotary spectral analysis. The results \( \bar{U}_1(f_j) \) were affected by power loss due to the windowing processes. To account for the power loss, \( \bar{U}_1(f_j) \) was divided with the average power in the window sequence \( W \). Here

\[ W = \frac{1}{m} \sum_{j=0}^{m-1} w^2(j) \]  

(3.16)

The desired power density spectrum was then computed as

\[ p_{\bar{u}_1u_1} (f_j) = \frac{1}{mW} \bar{U}_1' (f_j) \]  

(3.17)

Similarly \( p_{\bar{u}_2u_2} \), \( p_{u_1u_2} \) and \( Q_{u_1u_2} \) were also computed.

3.3. Results and discussion

The behaviour of the monsoon during 1977 and 1979 was contrasting. During 1977, the monsoon behaviour was above normal (Anon. 1978) while in 1979 it was below normal (Awade et al. 1986). Consequently, the freshwater inputs through
rainfall and river discharges into the Bay are expected to differ during the monsoon seasons of both the years. Added to these differences, the observations during M-79 were collected one month earlier to those of M-77.

3.3.1. Observed mean hydrography: The mean distributions (corresponding to the observational period) of the vertical profiles of observed temperature, salinity and \( \sigma_t \) for the polygons of M-77 and M-79 are shown in Fig. 3.2. The most notable feature is that the \( \sigma_t \) profiles of all locations followed salinity profiles but not the temperature profiles in the upper \( \approx 50 \) m which is quite uncommon to the other world oceanic regions. This shows the importance of salinity distribution influencing the dynamics of the northern Bay of Bengal. During M-77 the salinity differences within the polygon area were relatively larger in the upper layers (30m) compared to those of M-79. The salinity was also lower during M-77 compared to M-79. Higher river discharges resulting from the better monsoon might have probably produced these differences. The lowest surface salinity (<22 PSU) and strong halocline occurred at N station during 1977 because of its close proximity to the river mouths of Ganges and Brahmaputra. All the other locations showed relatively higher salinity with isohaline layers extending from surface to about 20 to 30m depth. These spatial differences in the vertical salinity distribution within the polygon are mainly attributed to the relative distance of the stations from the river mouth and local mesoscale circulation patterns causing differential advection of salinity. Very large spatial variability in the observed salinity field at the Head of the
Bay during M-77 and M-79 was also reported by Sarma et al. (1988). During both the experiments the salinity profiles below 50m depth resembled each other with diminishing spatial differences. This result suggest that the year to year variability in the salinity field is probably limited to the upper 50 m.

Brunt-Vaisala frequency (BVF) was calculated utilizing the mean temperature and salinity profiles for each observational period for all the locations (Fig. 3.3). Relatively large values of BVF are noticed during M-77 at N and E locations in the near surface layers due to the development of strong halocline caused by freshwater discharges at the Head of the Bay. The observed spatial variability of BVF was also large during M-77 in the upper layers within the observational array. The variability of BVF was insignificant below 100m. During M-77, the large values of BVF and a sharp decrease in the topmost 50m water column is prominently seen only at N location (closest station to the river mouth). In the near surface layers the BVF values decreased in a clockwise manner suggesting the reduction of dilution from N to W locations. However, the situation during M-79 was more or less similar to that of open ocean conditions with the only exception at N location due to its proximity to the river mouth as inferred from low salinity values there.

3.3.2. Short-term variability in the salinity field:

The day to day variability observed in the top 100m water column at all the four corners of the stationary polygons during M-77 and M-79 is depicted in Fig. 3.4. Daily averaged salinity profiles are utilized to construct
Fig. 3.3. Brunt-Väisälä Frequency (N) for polygons of M-77 and M-79.
depth-time sections. The isopleths were drawn at interval of 2 PSU for salinities < 33 PSU and at interval of 0.25 PSU for salinities > 33 PSU for M-77. But they were drawn at constant interval of 0.25 PSU for M-79 as the vertical salinity gradients were weak during M-79.

During the observational period, the surface salinity at N during M-77 showed a dramatic reduction by as much as 8 PSU within a week. Such variability is quite unusual in a deep oceanic region. Contrary to this, during the same period the surface salinity at E registered an increase by 4 PSU.

Rainfall was measured onboard three ships only during M-79. The daily march of estimated evaporation E (following Rao and Basil Mathew, 1988), observed rainfall P, E-P, cumulative E-P and observed surface salinity at three stations of M-79 are shown in Fig. 3.5. As E-P was mostly positive with occasional exception on rainy days, the surface salinity is expected to show an increase. However, the rainfall occurred at S during 18-22 July 1979 seemed to have lowered the surface salinity. But the impact of the heavy rainfall event of 18 July at W is not noticed in the daily-averaged surface salinity. However, a closer examination of the diurnal variation of salinity in the topmost 50 m water column on 18 July at W (Fig. 3.6) revealed the influence of rainfall event limiting to the top 20 to 30 m water column. In the surface layer a mild increase in salinity by about 0.08 PSU for 0000 hr to 1200 hr may be attributed to the eastward advection of saline waters (observed eastward flow at 25 m was 55 cm/s and the mean salinity of the topmost 20 m water column was higher at W than that at E). The sudden drop of salinity from 1200 hr to
Fig. 3.4. Depth-time sections of salinity during M-77 and M-79.
Fig. 3.5. Evaporation (E), precipitation (P), E-P, \( \Sigma E-P \)
and salinity (S) at three stations of M-79.
Fig. 3.6. Salinity for the upper 50 m, surface winds and currents on 18 July 1979 at W79.
1800 hr by about 0.09 PSU in the upper 20 m water column could be attributed to rainfall (4.2 cm of rainfall from 1316 hr to 1455 hr) at W. After the cessation of rainfall, the salinity again showed an increase from 1800 hr to 0000 hr. Before the occurrence of the rainfall event a well-marked halocline is evident between 20 and 30 m depths at 1800 hr. During the following 6 hours this stratification weakened probably under the influence of surface wind (~8 m/s) and wave mixing. The dilution caused in the upper 20 m water column due to this rainfall event is estimated utilizing the following equation.

\[ \Delta S = \left( \frac{E - P}{h} \right) S \] (3.18)

where \( \Delta S \) is salinity change; \( S \) the average salinity of the 20 m water column before the occurrence of rainfall (33.39 PSU) and \( h \), the isohaline layer depth (20 m). The computed value of \( \Delta S \) 0.067 PSU against the observed change of 0.09 PSU is in reasonable agreement.

3.3.3. Short term variability in the current field:

Direct measurements of currents at half hour intervals were utilized to describe the short term variability in the flow regime at selected depths in the top 200 m water column. These data provide a good description of the vertical structure of velocity field in the upper 200 m water column. The current meter data were filtered to remove the variance with periods less than semidiurnal (M2) tide i.e. 12.5 hrs (0.08cph). The smoothed data are shown as sticks for all the 4 locations in Figs. 3.7 and 3.8 for M-77 and M-71 respectively. The observed surface wind data were also
subjected to similar processing and the wind sticks are shown in the topmost panels of Figs. 3.7 and 3.8. During M-77, the surface winds were predominantly southwesterly at all locations with an average speed of 9 m/s, implying a nearly steady wind forcing over the observational array (Fig. 3.7). However, the subsurface current field showed some interesting variations in space and time. The overall flow was towards southeast at N and S locations, and towards northwest at E and W locations suggesting convergence in the northeastern sector and divergence in the southwestern sector. The flow weakened very rapidly with depth only at N and E locations while the weakening was moderate at S and W locations. This rapid downward decay of current strength only at N and E locations may be attributed to weak downward transport of surface wind stress due to strong stratification in the pycnocline. During M-79 the surface winds were also from southwest at all locations with an average speed of 7 m/s (Fig. 3.8). The flow at N and W was towards northeast while at E and S locations it was towards southwest. A close examination of the current sticks indicates the presence of a clockwise eddy circulation extending from 25m to 200m depth. Swallow (1983) inferred a weak clockwise eddy of 400-500 km diameter at this area from the temperature field. This eddy was centered near S location where the 20°C isotherm was depressed by 30m. The reduction in the current speed with depth at N and E locations is smaller in magnitude compared to that noticed during M-77. It is to be noted that the density stratification was also less during M-79 compared to that of M-77. This perhaps clearly demonstrates the importance of local stratification in the downward transport.
Fig. 3.7. Stick plots of surface winds and subsurface currents during M-77. Both the wind and current vectors start from the origin.
Fig. 3.8. Stick plots of surface winds and subsurface currents during M-79. Both the wind and current vectors start from the origin.
of surface momentum. Flow was strong and steady throughout the 200m water column only at the S location. During both the experiments the flow regime exhibited a well defined energetic oscillatory nature. These synoptic scale fluctuations show an excellent correspondence with the local inertial periods. Pollard and Millard (1970) successfully demonstrated the importance of local wind forcing on the amplitude of inertial oscillations beneath the mixed layer. The amplitude of these inertial oscillations also appeared to be related to the local stratification.

3.3.4. Mean wind and current patterns:-

The surface wind and subsurface current data were vectorially averaged for the total observational period and the mean vectors are presented for M-77 and M-79 in Fig. 3.9. During both M-77 and M-79 the surface winds were predominantly from southwest with average speeds of 9m/s and 7m/s respectively. But the observed current vectors during both the experiments do not resemble each other. A well defined clockwise circulation is evident only during M-79 with no significant variation either in direction or in speed with depth. Apparently no Ekman type of balance is noticed during either of the experiments with the only exception at S location during M-77. The forcings produced by thermohaline gradient and massive river discharges might have been significant in producing the observed flow patterns in the upper 200m water column. The significant weakening of the flow from the mixed layer to thermocline noticed at N and E locations during M-77 was not noticed during M-79 when the pycnocline at N and E locations was relatively weaker. On the
Fig. 3.9. Vectorially averaged surface winds and subsurface currents during M-77 and M-79.

Fig. 3.10. Rotary spectra during M-77.

Fig. 3.11. Rotary spectra during M-79.
other hand, during M-79 the flow was uniformly stronger throughout the 200m water column at all the four locations. This feature suggests the importance of the strength of pycnocline in determining the vertical structure of the flow beneath the mixed layer base. The clockwise circulation can also be inferred from the mean flow field only during M-79. The stronger flow regime noticed at S location also suggests that the centre of the clockwise eddy was located towards south within the observational array which is in accordance with an earlier inference of Swallow (1983). This eddy circulation is noticed throughout the 200m water column.

3.3.5. Rotary spectral analysis:-

Rotary spectra were computed to examine the nature of embedded periodic oscillations in the current field over a range of frequencies of positive (corresponding to velocity vectors that rotate anticlockwise with time) and negative (corresponding to velocity vectors that rotate clockwise with time) components of the flow. These clockwise (continuous line) and anticlockwise (dashed line) spectral estimates for all the locations and depths for M-77 and M-79 are presented in logarithmic scale in Figs. 3.10 and 3.11 respectively. The most striking feature of the rotary spectra is the dominance of the clockwise component over anticlockwise component during both M-77 and M-79 especially in the low frequency band. In general the spectral energy is about an order of magnitude higher in the clockwise spectra compared to the anticlockwise spectra. The peaks in the clockwise spectra mostly correspond to inertial, diurnal and semidiurnal periodicities. However, the inertial peak is insignificant in
the anticlockwise spectra as the inertial flow is clockwise in the northern hemisphere (Pollard and Millard, 1970). In general, a spectral peak of 128 hr period is dominant in the clockwise spectra while another peak of 85 hr period is noticed in the anticlockwise spectra at N, E and S locations during M-79. These peaks correspond to approximately 3-5 day oscillations in the current field. Such periodicities in the observed outgoing longwave radiation representing active weather over this region during summer monsoon season was reported earlier (Lau and Chan, 1988). However, these peaks could not be resolved for W location during M-79 and for all locations during M-77 due to short data lengths. During M-79, the inertial peak in the clockwise spectra was prominent at E and S locations compared to other locations. During M-77 and M-79 the peaks corresponding to diurnal and semidiurnal periods are well resolved in both clockwise and anticlockwise spectra revealing the importance of tidal forcing in producing these oscillations. However these peaks were less prominent at E and S locations. At N location where the stratification was strongest, multiple peaks in the low frequency band are noticed both in the clockwise and anticlockwise spectra only during M-79. The absence of such a feature at N location during M-77 may be attributed to low resolution of spectra due to short data length (6 days).