CHAPTER II

THERMOHALINE VARIABILITY IN THE NORTHERN BAY OF BENGAL DURING THE MONSOON TROUGH BOUNDARY LAYER EXPERIMENT (MONTBLEX-90)
2.1. Introduction

The intra-seasonal variability of summer monsoon over India is characterised by the movement and intensity of the seasonal monsoon trough extending from northwest India to the head of the Bay of Bengal. This trough is known to have organised structure with both lateral and vertical gradients in pressure, winds and organised moist convection and oscillates north-south corresponding to inactive and active regimes of the monsoon respectively (Raghavan, 1973; Rao, 1976). Majority of the monsoon depressions form in the oceanic region of the trough (the northern Bay), travel in north-westerly to westerly direction along the trough, dissipate over the land and produce significant rainfall over the Indo-Gangetic plains. The composite of the monsoon depression tracks for a 16 year period (INDIA METEOROLOGICAL DEPARTMENT, 1979) provides ample evidence on the genesis of meteorological disturbances over the head of the Bay of Bengal and their propagation along the monsoon trough (Fig. 2.1). Although some scanty information on the thermodynamic characteristics of this trough is available (Aanjaneyulu, 1969; Raman et al., 1978; Anto et al., 1982; Holt and Sethuraman, 1986; Mooley and Shukla, 1989), systematic observations covering the land and oceanic regions of the trough are lacking. Moreover, the meteorological and oceanographic conditions favourable for the genesis of the monsoon lows/depressions at this region still remain unknown. The summer monsoon experiments conducted during 1977 and 1979 provided some preliminary insight on this problem (Rao et al., 1981; Rao and Rao, 1986; Rao, 1987; Rao et al., 1987). Long continuous time series measurements of meteorological and oceanographic variables are essential to understand this
Fig. 2.1. Typical tracks of meteorological lows/depressions during July (India Meteorological Department, 1979) with spatial transects (.........) and time series station (●) of ORV Sagar Kanya during MONTBLEX - 90. Stations 1 to 19 and 19 to 28 are designated as Leg1 and Leg2 respectively.

Fig. 2.2. Observed (---) and estimated (-----) radiation using Lumb's (1964) scheme at 18.5 N & 89 E during August 1977 and at 18 N & 89.5 E during July 1979.
problem in greater detail.

To partially augment the above situation under a national programme, MONsoon Trough Boundary Layer EXPERiment (MONTBLEX) was carried out during the summer monsoon seasons of 1989 and 1990 (Goel and Srivastava, 1990). Various leading institutions of India participated in this programme. The atmospheric boundary layer was monitored at various locations in the monsoon trough region utilizing the instrumented towers, air-craft flights and routine meteorological observations. The oceanographic component was conducted during August and September 1990 by deploying ORV Sagar Kanya for time series measurements of atmospheric and oceanographic variables at a stationary location in the head of the Bay of Bengal.

The oceanographic data set thus obtained is first of its kind as there was no attempt to collect systematic time series measurements of surface meteorology and vertical profiles of temperature and salinity for such a long duration for any location in the Bay of Bengal during any season. The observational period represents both active and inactive regimes of the monsoon. Thus, this oceanographic data set is ideally suitable for the evaluation of one dimensional numerical models which can resolve the short-term variability caused in the upper layer thermal structure of the ocean resulting from local meteorological forcing (Sanil Kumar et al., 1995a).

An attempt is made in this chapter to understand the observed variability in the surface meteorological forcing and thermohaline structure in the upper layers of the ocean during MONTBLEX at the oceanographic station. An attempt is made to explain the genesis of meteorological
lows/depressions in terms of local surface meteorological forcing and thermal conditions of the upper layers. Two 1-dimensional numerical models, KTDM (Kraus and Turner, 1967 - Deman, 1973 - Miller, 1976) and NK (Niiler and Kraus, 1977), were used to simulate the observed mixed layer characteristics.

2.2. Data and methodology

ORV Sagar Kanya covered two spatial legs for XBT (Sippican) survey (stations 1-19:15 to 17 August 1990 and stations 19-28:1 to 2 September 1990) in addition to time series measurements of surface marine meteorological elements at hourly interval and vertical profiles of temperature and salinity at three hourly interval. The temperature and salinity were collected using a MICOM STD (TSK Japan; accuracy: temperature 0.05°C, salinity 0.04 PSU, depth 0.3 %) at a stationary location 20°N and 82°E (Fig. 2.1) from 18 August to 1 September 1990 (phase-I) and from 8 to 19 September 1990 (phase-II). The wind speed recorded onboard at 22.5 m height was reduced to 10 m height following Wu (1980).

2.2.1. Surface heat budget estimates

Heat exchanges across the air-sea interface are estimated to assess the relative importance of local meteorological forcing in producing the observed changes in the surface layer. As no direct measurements of these fluxes are available, empirical relationships were utilised to estimate all the terms of heat budget equation

\[ Q = Q_I + Q_B + Q_S + Q_E \]  

(2.1)
where $Q$ is net surface heat flux reckoned positive when gained by sea, $Q$ is solar radiation at sea surface, $Q_B$ is net long wave radiation, $Q_S$ is sensible heat flux and $Q_E$ is latent heat flux. In the tropics, $Q_I$ is positive and all the other terms ($Q_B$, $Q_S$ and $Q_E$) generally are negative.

To estimate the solar radiation at sea surface, direct radiation at the top of the atmosphere (solar constant - $Q_o$), mean transmittivity of the atmosphere and the correction for cloudiness and reflectivity of the sea surface are required. Dobson and Smith (1988) evaluated a number of models for the estimation of insolation and found that the formulation of Lumb (1964) is more accurate and accordingly modified it for general use. This scheme which gained general acceptance (Dobson and Smith, 1988; Sanil Kumar et al., 1994; Harish Kumar and Mohankumar, 1995), is used in the present study. The amount of solar radiation incident on a unit horizontal area at the sea surface is

$$Q_I = Q_o \sin(\alpha) (A + B_1 \sin(\alpha))$$

where $Q_o$ is 1368 Wm$^{-2}$ (Frohlich and London, 1986), $\alpha$ is the elevation of sun, $A_I$ and $B_I$ are the regression coefficients for different cloud amounts (Table 2.1).

Table 2.1: Regression coefficients for different cloud amounts

<table>
<thead>
<tr>
<th>cloud (octa)</th>
<th>1</th>
<th>2</th>
<th>3</th>
<th>4</th>
<th>5</th>
<th>6</th>
<th>7</th>
<th>8</th>
</tr>
</thead>
<tbody>
<tr>
<td>$A$</td>
<td>0.517</td>
<td>0.474</td>
<td>0.421</td>
<td>0.380</td>
<td>0.350</td>
<td>0.304</td>
<td>0.230</td>
<td>0.106</td>
</tr>
<tr>
<td>$B$</td>
<td>0.317</td>
<td>0.381</td>
<td>0.413</td>
<td>0.486</td>
<td>0.457</td>
<td>0.438</td>
<td>0.384</td>
<td>0.285</td>
</tr>
</tbody>
</table>
\[
\sin \alpha = \sin \phi \sin \xi + \cos \phi \cos \xi \cos h \quad (2.3)
\]

\[
\xi = -23.45 \cos(t+11) \quad (\text{Oke}, 1987) \quad (2.4)
\]

\[
h = 15 \times (12 - t) \quad (\text{Oke}, 1987) \quad (2.5)
\]

where \( \xi \) is the declination, \( h \) is the hour angle of the sun and \( t \) is the time in hrs. This scheme produced realistic estimates of observed radiation at two closely stations during August 1977 and July 1979 (Fig. 2.2). The agreement between the observed and estimated radiation is very encouraging with exceptions during disturbed days when this scheme overestimated due to nonincorporation of cloud thickness.

The effective outgoing long wave radiation from ocean \( (Q_B) \) is the balance between the long wave radiation from the sea surface and from the atmosphere. The cloud cover, sea-air temperature difference, water vapour content of the atmosphere immediately above the sea surface are the important factors controlling \( Q_B \). Several semi-empirical expressions are available to estimate \( Q_B \). Following Fung et al (1984), the net long wave radiation \( (Q_B) \) was estimated employing the empirical expression developed by Berliand and Berliand (1952).

\[
Q_B = \varepsilon \sigma T_S^4 (0.39 - 0.05 \varepsilon^{0.5}) (1 - 0.59 c_1) + 4 \varepsilon \sigma T_S^3 (T_S - T_a) \quad (2.6)
\]

where \( \varepsilon \), emissivity of the sea surface (0.98); \( \sigma \), Stefan-Boltzmann constant (\(5.67 \times 10^{-8}\) W.m\(^{-2}\)K\(^{-4}\)); \( T_S \), sea surface temperature (K); \( T_a \), air temperature (K); \( c_1 \), vapour pressure...
(mb) at a height 10m; cl, fractional cloud cover (0≤cl≤1).

The turbulent heat fluxes from the ocean are the heat loss/gain due to latent heat \(Q_E\) and sensible heat \(Q_S\) fluxes. In the tropical ocean, \(Q_E\) contributes to the heat loss due to evaporation while \(Q_S\) can be either gain or loss. The latent and sensible heat fluxes are parameterised with widely used bulk aero- dynamic method.

\[
Q_E = \rho_a C_L (q_e - q_{10}) V \\
Q_S = \rho_a C_C (T - T_{10}) V
\]  

where \(\rho_a\), density of air (Kg.m\(^{-3}\)); \(C_L\), latent heat of evaporation (J.Kg\(^{-1}\)); \(q_e\), specific humidity corresponding to \(T\) (Kg.Kg\(^{-1}\)); \(q_{10}\), specific humidity at 10m (Kg.Kg\(^{-1}\)); \(V\), wind speed (m.s\(^{-1}\)); \(C_C\), specific heat of air at constant pressure (J.Kg.\(^{\circ}\)K); and \(C_e\) and \(C_h\) are the non-dimensional transfer coefficients for moisture and heat respectively. The \(C_e\) and \(C_h\) are derived from a review of previous determinations following McCreary and Kundu (1989) and Friese and Schmitt (1976) respectively. These formulae take into account of the atmospheric stability.

\[
C_e = 0.0015 + 0.00033 (T_a - T) \\
C_h = 0.97 \times 10^{-3}; T_a - T < 0 \\
C_h = 0.88 \times 10^{-3}; T_a - T > 0
\]

2.2.2. Heat content of the water column

The net heat gain \(Q\) causes increase/decrease of heat content (HC) of the upper layers of the water column and sea surface temperature (SST). The increase in the heat storage
in the sea appears to trigger the formation of meteorological systems and thus resulting into the release of the excess heat to atmosphere (Gray, 1975). Therefore the temporal distribution of Q, HC, SST and the events of meteorological lows/depressions were compared to identify the relation between them. The HC is more representative to atmospheric processes when it is computed with respect to a shallow isotherm depth compared to any fixed depth (Stevenson and Niiler, 1983). So the HC was computed for the water column from the sea surface and the depth of the warmest isotherm appeared throughout the observational period using the following equation:

\[
HC = \rho C_p \sum_{i=1}^{r} (T_i - T_{ref}) \, dz \tag{2.11}
\]

where \(\rho\) is density and \(C_p\) is specific heat of the water column, \(dz\) is depth slabs of 1 m thickness, \(T_i\) is the observed temperature and \(T_{ref}\) is the reference temperature. The shallowest isotherm present throughout the observed record is taken to identify the near surface heat content variability that has occurred during this period.

2.2.3. One dimensional numerical models

It is well known that during surface heating regime under light wind conditions, the mixed layer shoals with an increase in temperature and during cooling regime mixed layer deepens due to convective overturning. Under strong winds, even if there is net heating, the surface layer deepens and cools. These changes are caused by due to turbulent entrainment of colder and denser water into the mixed layer from below layer.
2.2.3.1. Assumptions

The ocean is assumed to be horizontally homogeneous, incompressible and stably stratified fluid (Boussinesq approximation). The 'wave' like dynamical effects such as gravitational, inertial and Rossby waves are ignored. The surface mixed layer is considered to be vertically homogeneous down to a level 'h' with a density discontinuity at the base (Fig. 2.3). All the heat and momentum are uniformly distributed throughout this layer by turbulent diffusion. The time scales for the distribution of these properties are small compared to the times over which the processes of interest are assumed to occur. Below the surface layer, a stable density profile is specified at \( z = -h \).

The time dependent behaviour of the mixed layer at a stationary position is formulated from a set of differential equations based on equations of conservation heat, salinity and mechanical energy with appropriate boundary conditions. In the final equations, time derivatives of \( h \) (mixed layer depth), \( T \) (mixed layer temperature), \( S_m \) (mixed layer salinity), \( T_h \) (temperature immediately below the mixed layer) and \( S_h \) (salinity immediately below the mixed layer) are expressed in terms of the time-dependant boundary inputs, like surface wind stress, incoming solar radiation, net heat loss at the surface and the temperature and salinity (or density) gradients below the mixed layer.

2.2.3.2. Boundary conditions

i) At the surface

At the top of the surface layer, the radiative and turbulent heat fluxes are proportional to the net heat
Fig. 2.3. Schematic diagram to show the mixed layer physics (Denman, 1973). The vertical temperature profile consists of a homogeneous mixed layer of thickness $h$ and temperature $T$, followed by a temperature discontinuity ($T - T_h$), and a region with temperature gradient ($\frac{\delta T}{\delta z}$). The surface fluxes are wind stress; incoming solar radiation; back radiation; latent heat and sensible heat flux together. The solar radiation decays in the mixed layer exponentially. A vertical velocity is prescribed below the mixed layer.
transfer through the ocean surface and fresh water flux is proportional to the difference between evaporation and precipitation.

ii) At the mixed layer base

The effect of diffusion across the stable interface at the bottom of the mixed layer is neglected and all the mixing processes are associated with the entrainment of the denser water into the mixed layer.

iii) Below the mixed layer base

In order to specify the system of equations completely, the temperature gradient below the mixed layer should be known. The assumptions made here are (a) no turbulent energy penetrates below \( z = -h \) and (b) the local influence of surface forces acting on the ocean at a depth \( z < -h \) is a small fraction of the forcing.

2.2.3.3. Integral equations of the Niiler and Kraus (1977) model

The final form of the major equations which determine the mixed layer characteristics of the model of Niiler and
Kraus (1977) are given below:

\[
\frac{gh}{2} \frac{\rho - \rho_m}{\rho} \frac{d\rho}{dt} + m_1 \frac{\rho U^2}{2} + m_2 S + \frac{\rho g}{C} \int_0^h q(z)dz - \frac{\rho g}{C} \int_{-h}^h p = m_3 \frac{h}{4} (Q - |Q|)
\]

\[
\frac{h(Q - q_h - r_h)}{2} - m_3 \frac{h}{4} (Q - |Q|)
\]

\[
\frac{dT}{dt} + (T - T_h) \frac{d\rho}{dt} = \frac{1}{\rho \rho C_p} (Q - q_h - r_h)
\]

\[
\frac{dS_m}{dt} + (S - S_h) = \frac{S_m}{(E - P)}
\]

\[
\frac{dT_h}{dt} = \gamma \rho C_p \left( \frac{dT}{dt} - L_T \frac{dh}{dt} \right)
\]

\[
\frac{dS_h}{dt} = -L_s \frac{dh}{dt}
\]

where \( U_* \), frictional velocity \( (U_* = \sqrt{\tau/\rho} \) where \( \tau \) is surface wind stress); \( \alpha \), coefficient of thermal expansion; \( g \), acceleration due to gravity; \( \rho \), mean density of the water column; \( \rho_m \), mean density of the mixed layer; \( \rho_h \), density just below the mixed layer; \( C_p \), specific heat of sea water; \( q_h \), heat flux into the thermocline; \( r_h \), radiative heat flux into the thermocline; \( L_T \), temperature gradient below the mixed layer; \( L_s \), salinity gradient below the mixed layer and
\( m_1, m_2 \) and \( m_3 \) are the coefficients which represent the fractions of wind, shear and convection energy available to change the potential energy of the water column. These unknown tuning coefficients of this model do not have universally acceptable values.

The term \( A \) is rate of energy that needs to agitate entrained water; \( B \), rate of work done by surface wind; \( C \), rate at which energy of mass velocity field is reduced by mixing across the layer base; \( D \), rate of potential energy change produced by penetrating solar radiation; and \( E \), rate of potential energy produced by fluxes across the sea surface. When the wind speed changes rapidly, inertial current in upper ocean produces a sharp velocity gradient across the layer base and then the term \( C \) determines the rate of shear mixing at the interface. Following Price et al. (1978) and Paulson and Simpson (1977), the terms \( S \) and \( q(z) \) in equation 2 may be written as

\[
S = \frac{\rho}{2} \left[ |V - V_h| \right]^2 \frac{dh}{dt} + \frac{1}{2} (V - V_h) \tau_h
\]

\[
q(z) = Q_i \left[ R e^{-\gamma_1 t} + (1-R) e^{-\gamma_2 t} \right]
\]

where \( V \) is velocity of current in mixed layer, \( V_h \) is velocity of current below the mixed layer base, \( R \) is a constant which varies with the water type, \( \tau_h \), momentum flux into the thermocline; and \( \gamma_1 \) and \( \gamma_2 \) are extinction coefficients of light attenuation for long and short wave length respectively. The terms involving \( q_h, r_h \) and \( \tau_h \) are small quantities and never exceed 10% of the principal terms. So they were significant only in the case of deep convective regime or for long term integration. Hence they were
neglected in the present study.

2.2.3.4. Comparison between NK (Miller and Kraus, 1977) and KTDM (Kraus and Turner, 1967 - Denman, 1973 - Miller, 1976) model

The NK is analogous to KTDM, when the shear production term $m_2 S$ is ignored and convective efficiency parameter $m_3$ is chosen as unity. The remaining unknown coefficient in KTDM is only $m$ analogous to $m_1$. The important factor in $m_2 S$ is $\delta v^2$ i.e. $|V - V_h|^2$, square of the magnitude of velocity difference across the base of the mixed layer. An explicit expression for $\delta v^2$ is given by Price et al. (1978) as

$$|V - V_h|^2 = \frac{2V_s^4}{h^2 f^2} (1 - \cos (ft))$$  \hspace{1cm} (2.19)

where 'f' is coriolis force and 't' is time. In KTDM, $U_e$ alone is considered as the relevant scale velocity in determining the MLD. The inclusion of $\delta v^2$ produces more realistic results in the case of mixed layer deepening under strong wind events or long term integration. Following Paulson and Simpson (1977) double exponential term (eq. 18) is used in NK for the absorption of solar radiation at different depths instead of the single exponential term in KTDM (ie. $q(z) = Q e^{-\gamma h}$).

Equations 2.12 to 2.14 were integrated using Runge-Kutta numerical integration scheme with a time step of 5 min. When $dh/dt$ is $< 0$, terms involving below layer gradient lose the significance (no entrainment) and 'h' is found out using Newton's iterative technique. The MLD is taken as the deepest depth where a drop of 0.1°C occurs from the SST in the observed vertical temperature profiles (Davis
et al., 1981). Since the temperature values of MICOM STD close to surface are not reliable, the temperature at 5 m depth was considered to represent the mixed layer temperature.

2.3. Weather Summary of the Observational Period (15 August to 19 September 1990)

The time sequence of maps of surface pressure analysis (India Meteorological Department) over India and the Bay of Bengal corresponding to 0830 IST for the period 15 August to 19 September 1990 is shown in Fig. 2.4. The corresponding time sequence of cloud imageries viewed by INSAT 1D is also shown in Fig. 2.5. The genesis, intensification and landward propagation of meteorological disturbances characterised by close isobars and organised massive cloud cover are evident from Figs. 2.4 and 2.5. During this observational record, five meteorological lows formed over the head of the Bay of Bengal and traversed in west-northwesterly direction along the monsoon trough. Only one of these 5 systems intensified into a deep depression. The average life of these disturbances was about 5 days. Thus the observations collected during this experiment provided typical signatures of both active and inactive regimes of the monsoon. The time sequence of satellite imageries (Fig. 2.5) show the evolution of the cloud cover over the Bay of Bengal and the Indian subcontinent during the observational period. In general, massive cloud cover appeared southwest of the monsoon lows. The cloud cover shows a broad correspondence with the intensity of the monsoon trough and the strength of meteorological disturbances.
Fig. 2.4. Time series of surface pressure analysis - arranged columnwise in the chronological order from 15 August to 19 September 1990.
Fig. 2.5.
Time series of cloud imageries viewed by INSAT 1D - arranged columnwise in the chronological order from 15 August to 19 September 1990.
2.4. Surface meteorology and heat budget estimates

The hourly march of time series of the standard marine meteorological parameters as surface pressure (PR), wind direction (DD) and speed (FF), visually observed cloud cover (CL), dry (DB), wet bulb (WB) and sea surface temperatures (SST) are shown in Fig. 2.6. The perturbations caused by the meteorological disturbances are best captured in PR, DD, FF and SST distributions. The winds strengthened during disturbed weather conditions with sudden changes in direction. The DB and WB did not reflect any regular cyclic variations as noticed in other elements. However, violent fluctuations are noticed in DB and WB in association with rapid variations in wind direction. In general, SST showed an inverse relationship with surface pressure with an approximate time lag of 2 days. A broad correspondence between strengthening of winds and lowering of SST is also noticed during disturbed conditions.

The heat budget components were estimated with the hourly observed marine meteorological data and then daily averaged. The daily march of these components are shown in Fig. 2.7. In general, \( Q_i \) was mostly in excess of 150 W/m\(^2\) and showed synoptic scale fluctuations corresponding to variations in cloud cover due to disturbed and undisturbed regimes of the monsoon. The \( Q_B \) was almost invariant during both the phases with an average value of 35 W/m\(^2\). The most dominant heat loss term for the tropical oceans is \( Q_E \) which showed dramatic increase in association with deep depression (270 W/m\(^2\)) on 20 August 1990 and meteorological lows on 28 August, 11-13 and 18 September 1990. Occurrence of weak \( Q_S \) suggests near neutral conditions during most of the observational period. The net heat flux \( Q \) was positive
Fig. 2.6. Time series of surface meteorological parameters.
Fig. 2.7. Time series of heat budget parameters.
throughout with exceptions corresponding to the days of disturbed weather conditions (20, 27 August and 18 September 1990). In general the net heat gain fluctuated between 0 and 100 W/m².

2.5 Thermodaline variability

The annual cycle of the SST as revealed by global MCSST analysis (100 km x 30 days) for 1990 (Legeckis, 1991) is characterised by strong heating from February to May followed by a moderate cooling during June and July (Fig. 2.8). Another phase of moderate warming is noticed from August to October followed by an intense cooling during November to January. The near surface water column is also characterised by a variation of -5°C in the annual cycle. The near surface waters are warmer than 28°C from April to November and showed only little variation during summer monsoon season contrary to the Arabian Sea. The daily averaged SST is presented for the observational periods for comparison with the satellite measurements in the inset of Fig. 2.8. The agreement is quite encouraging with a root mean square difference of 0.25°C.

The annual cycle of the observed temperature and salinity in the top most 250m water column at the time series station constructed from the data sets derived from Levitus (1982) is shown in Fig. 2.9. It must be cautioned that very few data are available for this region and hence these representations may not be very robust. However, the features are quite discernable. The thermal structure is characterised by a near surface mixed layer which is variable in depth and temperature. Maximum MLD (∼50 m) is found during winter. The salinity field shows the presence of a very sharp halocline
Fig. 2.8. Global MCSST analysis at 20°N and 90°E for 1990 (Legeckis, 1991). The figure in the inset represents the comparison between the MCSST analysis and sea truth data.
Fig. 2.9. Annual cycle of temperature and salinity in the top most 250m water column at the time series station (based on Levitus, 1982).

Fig. 2.10. Thermal structure along Leg1 and Leg2 based on XBT data.
in the near surface water column during the summer monsoon season due to massive fresh water discharges from the Ganges and Brahmaputra rivers. While the salinity below 100m depth varied only marginally throughout the year. Thus the station covered during MONTBLEX represents a very typical oceanographic setting where the salinity effects are expected to play prominent role on the near surface layer characteristics.

The horizontal variability of the thermal structure in the top 750m water column constructed with XBT data corresponding to two transects of MONTBLEX (Fig. 2.1) is depicted in Fig. 2.10. The thermal structure is distinctly characterised by a strong stratification in the upper 200m water column beneath the near surface mixed layer. In addition, the wavy nature of isotherms in the thermocline also suggests the presence of mesoscale eddy circulation or of propagating waves with horizontal scales of the order of 300 kms which is typical of mesoscale eddies in the Bay of Bengal (Rao, 1974; Swallow, 1983; Legeckis, 1987). The vertical extent of these perturbations is limited to about 250m of water column.

The mean vertical profiles of temperature, salinity and $\alpha$ based on 7 CTD casts representing both the phases are shown in Fig. 2.11a. The temperature profile is characterised by a shallow near surface isothermal layer capped over a steep thermocline extending up to 200 m depth. A very sharp halocline is also present between surface and 175 m depth below which salinity remained invariant with depth. The corresponding density profile reflects strong pycnocline in the top 200m water column.

The synoptic scale evolution of thermal structure in
Fig. 2.11. (a) Mean vertical profiles of temperature, salinity and $\sigma_t$ at the time series station. (b) Temperature and salinity profiles at 3 hourly intervals in chronological order. (c) Depth-time section of temperature. (d) Depth-time section of salinity.
the top 100 m water column at the stationary position is shown in Fig. 2.11b and c. The analysis is based on time series measurements of vertical temperature profiles recorded by MICOM STD system. Isotherms were drawn at 1°C interval. The thermocline showed a descent during phase-I and an ascent during phase-II. These trends in the vertical motion of isotherm topography within the thermocline suggest a possible influence of either a moving clockwise eddy or a propagating wave. Eddy of dimension of more than 440 km in the northern Bay of Bengal during the summer monsoon period of 1979 (Chapter III). Hence it is probable that a similar eddy might have been presented in this case too. However, the available data are inadequate to address this problem in greater detail. The embedded short period fluctuations in the isotherm topographies suggest the influence of internal tides and inertial oscillations. The corresponding variability noticed in the salinity field is depicted in Fig. 2.11b and d. The isohaline contours were drawn at 0.5 PSU interval for values greater than 32 PSU and at 1 PSU interval for values less than 32 PSU to avoid congestion in the diagram. A great deal of small scale structure is evident in the topmost 20m water column throughout the observational period and this is more so during phase II. This type of small scale variability is caused by local rainfall, variations in river discharges and associated complex circulation regimes. The halocline also showed a descent during phase-I and an ascent during phase-II in accordance with thermal regime. The high frequency oscillations show a great deal of correspondence with the isotherm fluctuations.

The vertical stability regime represented by the Brunt-Vaisala frequency was computed utilising the vertical
profiles of temperature and salinity and the corresponding depth-time section is shown in Fig. 2.12. The contours were drawn at 3 cycles/hour interval. The pycnocline is characterised by a strong stability regime with values in excess of 10 cycles/hour. During phase-II, strong stratification occurred very close to the surface on account of strong gradients and rapid fluctuations in the salinity field. On the whole, the stratification in the pycnocline did not differ significantly between phase-I and phase-II.

2.6. Genesis of meteorological lows

As the thermal conditions of the near-surface water column are known to play a key role in the convective activity of the overlying atmosphere (Gray, 1975; Gadgil et al., 1984; Graham and Barnet, 1987), focus is laid on small scale thermal fluctuations within the isothermal layer in relation to disturbed and undisturbed weather conditions (Fig. 2.15). The 29°C isotherm appeared on 5 occasions (18-20 and 23-25 August, 31 August-1 September, 11 and 15-18 September 1990) during the entire observational record, each lasting approximately for 1 to 3 days. The vertical extent and duration of these mini-warm pockets also varied from one event to another. A very clear correspondence is noticed between the occurrence of these mini-warm pockets and subsequent genesis of meteorological disturbances (19-22, 27-28 August, 1-2, 11-13 and 18-19 September 1990). The genesis and sway of these meteorological disturbances resulted in surface cooling (SST around 28.7°C) by extraction of available surplus thermal energy. Warming of water beyond 29°C appears to trigger instability in the overlying atmosphere towards the formation of disturbed weather. Higher
Fig. 2.12. Depth-time section of Brunt-vaisala frequency (cycles/hour).

Fig. 2.13. Depth-time section of temperature for the upper 40 m water column. 'Dep' for depression and 'low' for meteorological low pressure systems.

Fig. 2.14. Time series of heat flux and heat content with respect to 28.7°C isotherm.
SST is already known as a favourable condition for the genesis of tropical cyclones (Gray, 1975; Graham and Barnet, 1987). In the monsoonal seas, SST of 28°C has been found as the threshold value for the generation of organised convection in the atmosphere (Gadgil et al., 1984). However, due to the averaging of the daily cloudiness (area considered – 2.5° x 2.5° square) and monthly SST (area – 5° x 5° square) for larger area, Gadgil et al. (1984) could not establish the exact threshold value of SST for the genesis of the atmospheric disturbances. The MONTBLEX experiment has provided an unique opportunity to collect time series of thermal profiles using high accuracy instrument (MICOM STD) at closer time intervals during a monsoon regime represented by both disturbed and undisturbed conditions. When considering the fact that the SST fluctuation even within 0.3°C (28.7°C to 29°C) appears to control the atmospheric stability. The accuracy of instrument being 0.05°C with a vertical resolution of 1 m amply meets the requirement.

The accumulation of net heat flux at the surface is expected to increase the heat content of the near-surface water column. The heat content with respect to 28.7°C isotherm (the warmest isotherm present throughout the observational period) was estimated (Fig. 2.14) to identify correspondence with local surface heat fluxes and the formation of the meteorological disturbances. Whenever the heat content of the water column exceeded ~0.4 x 10^8 J/m^2, the ocean released energy to the atmosphere, leading to cooling of the surface waters. These events remarkably coincided with the formation of the meteorological disturbances except for the low pressure system formed during 11-13 September 1990. Rao and Rao (1986) and Rao et al. (1987) reported such a
correspondence between the genesis of a depression and an increase in the heat content of the upper layers in the head of the Bay of Bengal. A close agreement between the distributions of $Q$ and $HC$, with a time lag of 1-2 days ($HC$ lagging $Q$) during phase I, suggests the importance of local heat exchange in regulating the thermal variability of the upper layers. The weak agreement during phase II is probably due to circulation patterns causing a complex thermohaline structure (Fig. 2.11b).

2.7. Numerical simulation of mixed layer characteristics

Among the coefficients, $m_1$ and $m_2$ represent the fraction of turbulent energy production due to (i) interaction of the surface stress and near surface current shear and (ii) interaction of entrainment stress and shear at the mixed layer base. The coefficient $m_1$ depends on the atmospheric and oceanic stability, sea state and surface roughness. In addition to these factors, $m_2$ depends on the stratification of water column. The correct dependence of these variables on the above mentioned factors is not known exactly (Bush, 1971). Therefore, it is a common practice (Davis et al., 1981; Sanil Kumar et al., 1994) to assign constant values by trial and error with bounds determined by laboratory and field experiments. The value for $m_3$ (0.83) (convective efficiency parameter) has been determined fairly well by the experiments of Deardorff et al. (1969). Considering the uncertainty in $m_1$ and $m_2$, in this study, NK model was run with different combinations of values drawn from Davis et al. (1981).

The results (simulated hourly time series of MLD and MLT) of these combinations represent minimum and maximum
values of the parameter range are presented Figs. 2.15 and 2.16. The disagreement between the simulated and observed series of MLD progressively reduced with progressive increase of \( m_1 \) and \( m_2 \) (Fig. 2.15). This is also equally true for MLT (Fig. 2.16). Better simulation with larger coefficients imply that mixing both at near-surface and mixed layer base was important in mixed layer variability at the northern Bay under the alternate rough and fair weather conditions. However, during phase II, the disagreement between the observed and simulated values were higher. Prominent small scale variability in both temperature and salinity profiles was evident during phase II (Fig. 2.11b). This variability showed the dominance of internal dynamic processes rather than the one dimensional processes and thus explain the reasons for larger departures between the observed and predicted values.

The \( m \) in KTDM model, is analogous to \( m_1 \) in NK model and hence its dependence on environmental factors is also similar. A wide range of values for \( m_1 \) are available in the literature varying from 0.009 to 0.01 (Kato and Phillips, 1969; Turner, 1969; Denman, 1973). However, in the present study a value of 0.009 was chosen after making a few trial runs. The corresponding simulated values of MLD and MLT are given in Fig. 2.17a and b. In general, KTDM model predicted deeper MLD while NK model predicted lower MLD. Also the shoaling events were comparatively well captured in NK model. Another notable point is that during the period of deep depression, the deepening rate of MLD was simulated reasonably well by NK; whereas it was poorly done by KTDM. For obtaining an overall picture of the variations between the observed and simulated values of MLD and MLT, the root
Fig. 2.15. Time series of observed and simulated (using NK model) mixed layer depth.
Fig. 2.16. Time series of observed and simulated (using NK model) mixed layer temperature.
Fig. 2.17a. Time series of observed and simulated (using KTDM model) mixed layer depth.

Fig. 2.17b. Time series of observed and simulated (using KIDM model) mixed layer temperature.
mean square (RMS) errors were computed (Table 2.2). In general, NK is found to perform better than KTDM for the northern Bay during summer monsoon period. Since the differences between NK and KTDM are mainly in the parameterization of the terms of shear production at the layer base ($m_zS$), the convective efficiency ($m_3$) and the absorption of solar radiation at different depths [$q(z)$], the relatively better performance of NK can be explored in terms of these terms. The $\delta v^2$ in $m_zS$ is important in the case of strong wind events. In the present case, the strongest wind event occurred on 20 August 1990 in association with the deep depression and the NK model performed better. So it is quite evident that the inclusion of $\delta v^2$ in NK model improved the performance of the model. The other possible factors are $m_3$ and $q(z)$. In NK, $m_3$ is restricted to 0.83 instead of 1 in KTDM. This implies that in KTDM, 100% of the surface energy flux is utilized for the mixed layer cooling while in NK, only 83% is used. Since the overcooling was not noticed in MLT with KTDM, it is concluded that the difference in the parameterization of $m_3$ is not very significant for the present data set. But, under different environmental conditions (when heating was weak), unrealistic cooling was observed with KTDM (Alexander and Kim, 1976; Sanil Kumar et al., 1993). The differences between the simulated values occurred using these two models during shoaling events of the mixed layer, can be explained in terms of the parameterization of $q(z)$. In KTDM, the solar radiation absorption at different levels is parameterized based on single exponential decay of the insolation, while in NK, the absorption is based on double exponential decay. It seems that the parameterization of $q(z)$ in NK appears to improve
Table 2.2. Kraus-Turner-Denman-Miller

<table>
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<th>Sech disc depth (m)</th>
<th>$\gamma$ (cm$^{-1}$)</th>
<th>rms deviation</th>
<th>mld (m)</th>
<th>mlt ($^\circ$C)</th>
<th>phase 1</th>
<th>phase 2</th>
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Table 2.2. Miller-Kraus

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<th>mld (m)</th>
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the performance of the model.

The lack of exact coincidence between the observed and simulated mixed layer characteristics with these two models, suggests that one dimensional forcing can not explain the entire observed variance as this area is characterised with rich eddy fields and associated complex mesoscale circulation (Legeckis, 1987; Chapter III) with overwhelming small scale structure evident from the high resolution vertical temperature and salinity profiles (Fig. 2.11b).