Indian Journal of Marine Sciences
Vol. 22, September 1993, pp. 194-197

Mixing in the Arabian Sea off Calicut during February 1985

P Madhusoodanan, P V Hareesh Kumar & N Mohan Kumar
Naval Physical & Oceanographic Laboratory, Thrikkakara, Cochin 682 021, Kerala, India

Received 2 December 1991, revised 13 April 1993

Short time variability in the temperature field at two stations off Calicut was analysed. Lower diurnal range of temperature was noticed in the surface layers at the shallow station (0.4°C) than at the deep station (0.8°C). This feature was explained based on surface heat exchanges, eddy diffusivity and Monin-Obukhov length. Monin-Obukhov length revealed the dominance of buoyant mixing over mechanical mixing at both the stations. Greater turbulence within the mixed layer at the shallow station (93 $cm^2 . sec^{-1}$) than at deep station (78 $cm^2 . sec^{-1}$) as revealed by the coefficient of eddy diffusivity reduced the amplitude of surface temperature at the shallow station. Two layers of thermal inversions were observed at the deep station while at the shallow station these two layers merged to form a single inversion layer. It is concluded that these thermal inversions are caused by the intrusion of the Arabian Sea High Salinity Watermass.

On a synoptic scale, variations of temperature and related parameters have been studied by several investigators for the Indian coasts. However, except the preliminary work of Hareesh Kumar and Rao[1], not much work has been carried out to understand the nature of mixing and eddy diffusion processes in the coastal waters. An attempt is made in this paper to document and explain the observed variability in the thermal structure and the nature of turbulence in the Arabian Sea off Calicut during February 1985.

Materials and Methods

R V Gaveshani occupied two time series stations off Calicut, from 10 (0400 hrs) to 12 (1000 hrs) February 1985 (shallow station) and from 12 (1400 hrs) to 14 (1200 hrs) February 1985 (deep station). The two stations are designated at S for shallow (11°N, 75°11′ E, depth < 100 m) and D for deep (11°, 1′ N, 74°56°E, depth > 1000 m) stations (Fig. 1). Ship was anchored as S whereas at D the observations were taken while drifting, but the position was reoccupied if the drift exceeded one nautical mile. Surface marine meteorological elements and vertical profiles of temperature (using MICOM BT) were collected at hourly intervals from these stations. Three hourly hydrocasts were also taken and salinity was estimated with the onboard salinometer (Autosal). Insolation Q_i was estimated following Lumb[2] with Atwater and Ball's[3] albedo correction (A):

$$Q_i = Q_0 \sin \theta T(1-A);$$
$$A = -0.0139 + 0.0467 \tan Z$$

where Q_0 is the solar constant (1353 W m^{-2}), θ is the solar altitude, T is the transmission factor estimated following Dobson and Smith[4] and Z is the zenith angle.

The net long wave flux (Q_b) was computed following Wyrtki[5]:

$$Q_b = \varepsilon \sigma T_s^4 (0.39 - 0.053\, q_s^{0.5})(1 - 0.53\, CL^2) + 4 \varepsilon \sigma T_s^3 (T_s - T_a)$$

where q_s is the specific humidity at the sea surface, T_s and T_a are the temperature of water and air respectively, ε is the emissivity of sea surface, σ is the Stefan-Boltzmann constant and CL is the cloud amount in fractions.

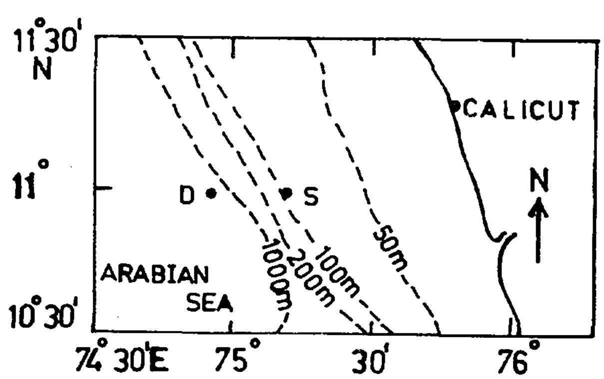

Fig. 1—Station location map

Turbulent fluxes of latent (Q_e) and sensible (Q_h) heat at the air-sea interface were estimated using bulk aerodynamic formulae:

$$Q_e = \rho_a C_e L_e (q_s - q_a)\, U$$

$$Q_h = \rho_a C_h C_p (T_s - T_a)\, U$$

where U = wind speed, q_a = specific humidity of the air, L_e = latent heat of evaporation, C_p = specific heat of seawater, ρ_a = density of air, C_e and C_h are the exchange coefficients for momentum and heat given by

$$C_e = C_h = (1 + 0.05\, U) \times 10^{-3} \quad \text{for } 0 < U < 20 \text{ m sec}^{-1}$$

All the heat flux terms are given in W m^{-2}.

Monin-Obukhov length[6] (L in meters) is given by

$$L = -U_*^3 / k B_0$$

where U_* (m sec^{-1}) is the frictional velocity $U_* = (\tau/\rho_w)^{1/2}$, τ is the surface wind stress ($\tau = C_d \rho_a U^2$), ρ_w is the density of seawater, k is the Von-Karman constant (0.4) and B_0 is the buoyancy flux given by

$$B_o = g/\rho_w (\alpha/C_p\, Q + F_s)$$

Here g is the gravitational accelaration, α is the coefficient of thermal expansion, C_p is the specific heat at constant pressure and F_s is the salt flux given by

$$F_s = \beta S / L_e (1 - S)\, Q$$

where β is the coefficient of haline contraction (0.8), S is the salinity and L_e is the latent heat of evaporation. Following Ostapoff and Worthem[7] K_z is given by

$$K_z = Z^2 \pi / P\, [\log_e (T_z/T_0)]$$

where T_0 and T_z are the amplitudes of temperature at the surface and at a depth Z m. For a diurnal heat wave, P = 24 h.

Results and Discussion

Time series of wind speed (FF), net air-sea heat flux (Q) and temperature at selected depths for shallow (S) and deep (D) stations are presented in Fig. 2. Wind field exhibited diurnal oscillation with an amplitude of 1 m sec^{-1} and 2 m sec^{-1} at S and D respectively. Net heat flux (Q) attained maximum (450 W m^{-2} at S and 600 W m^{-2} at D) at noon and minimum (-250 W m^{-2}) during night. Correspondingly, sea surface temperature (SST) exhibited diurnal oscillation at both S and D though with a different magnitude. SST shows larger diurnal range at D (0.8°C) than at S (0.4°C). Several short period oscillations are noticed in the thermal structure below 10 m at D and 15 m at S indicating the influence of internal waves. It can be inferred that when the heating penetrated to shallow depth (10 m at D and 15 m at S), diurnal range in temperature is maximum and *vice versa*.

To understand the role of mechanical and buoyant mixing processes, a non-dimensional length L known as Monin-Obukhov length was computed for 11 and 13 February (information on a full diurnal cycle was available only for 11 and 13 February). In a mixed layer of depth H turbulence can be generated by convection[8], if $H \gg L$ and by wind stress, if $H \ll L$. The average values of L at st. S is 1.02 (11 February) and at st. D it is 1.19 (13 February) and the corresponding mixed layer depths are 14 m and 9 m respectively. Since L is found to be much less than H at both the stations it can be concluded that buoyant mixing dominates over mechanical mixing.

The eddy diffusion coefficient (K_z) can be used to estimate the degree of turbulence in the water column[7]. Since the information on a full diurnal cycle was available only for 11 and 13 February, K_z was computed for these two days at different depth levels and finally averaged over the mixed layer. Higher values of K_z at st. S (93 cm^2 sec^{-1}) than at st. D (78 cm^2 sec^{-1}) clearly revealed grea-

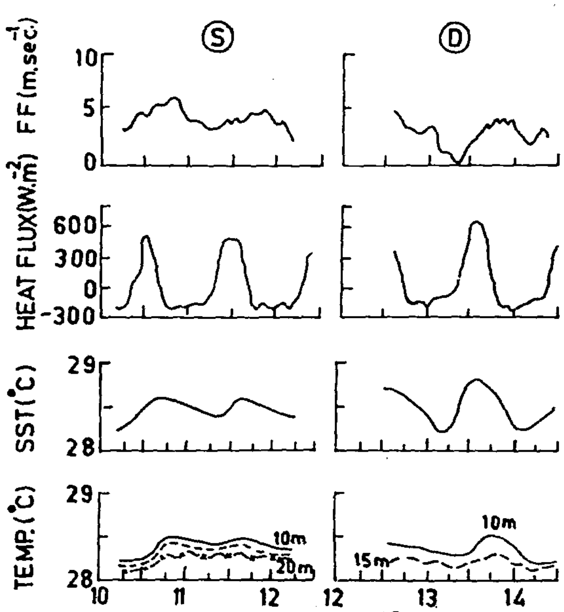

Fig. 2—Diurnal progress of wind speed (FF), net air-sea heat flux (Q), temperature at selected depths for shallow (S) and deep (D) stations

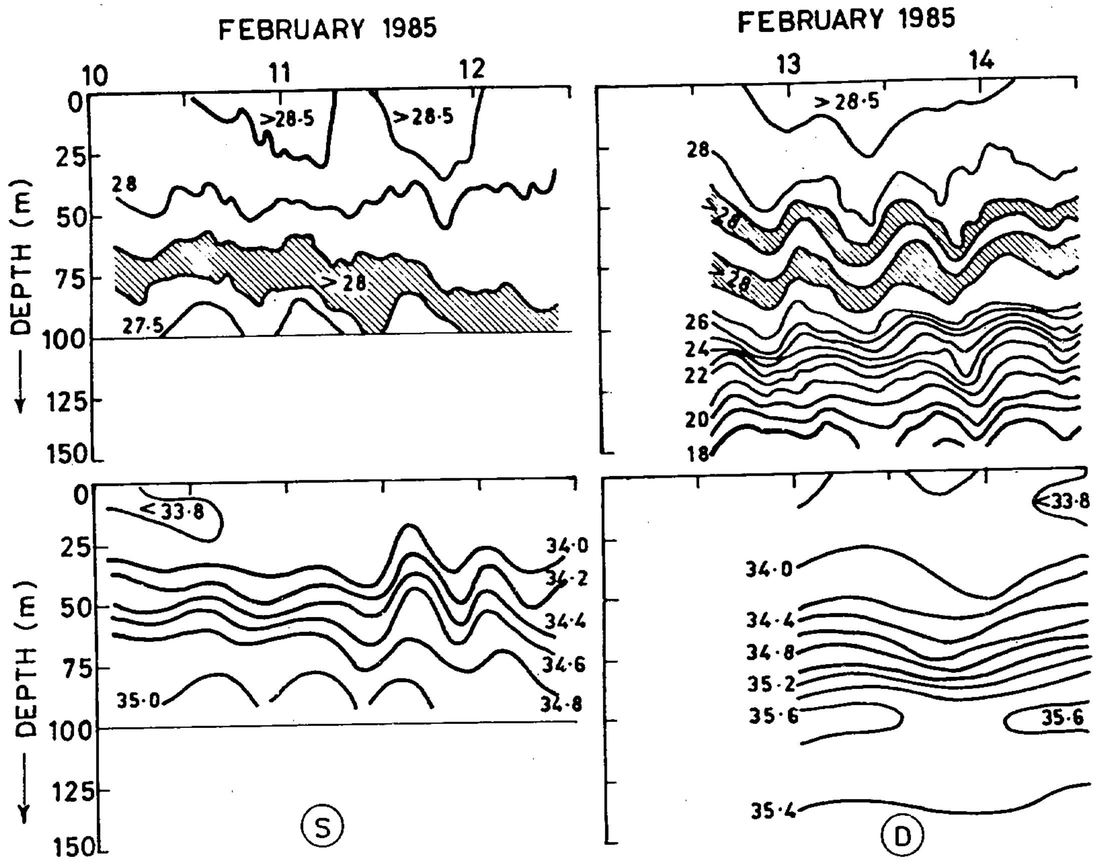

Fig. 3—Depth-time sections of temperature and salinity for shallow (S) and deep (D) stations

ter turbulence, hence more mixing at st. *S*. The accumulation/depletion of heat from a shallower depth along with the comparatively weak turbulent mixing resulted in larger diurnal range of temperature within the mixed layer at st. *D*.

The evolution of thermohaline structure at the two stations (Fig. 3) revealed the existence of thermal inversions. Thermal structure at st. *D* is characterised by two inversion layers (10 m thickness and 0.2°C) whereas only a single inversion layer (20 m thickness and 0.18°C) is noticed at st. *S*. Eventhough these inversion layers were noticed around 75 m, their relative positions vary with time under the influence of propogating internal waves. The instability caused by the thermal inversion is offset by the increase in the salinity (0.4×10^{-3} at st. *S* and 0.8×10^{-3} at st. *D*) which makes the stratification stable. Since the effect of diurnal heating was confined to top 15 m, the possibility of air-sea exchange processes in the formation of these inversions can be ruled out.

The T-S diagram (Fig. 4) revealed the existence of two important watermasses. In the upper few meters, a watermass having a temperature around 28°C and salinity $< 35.5 \times 10^{-3}$ is noticed. Darbyshire[9] attributed this water to the Equatorial Indian Ocean/Bay of Bengal watermass. The salinity maxima noticed around 100 m at st. *D* ($\sigma_t = 23.4$; temperature = 26°C; salinity = 35.65×10^{-3}) suggests the presence of Arab-

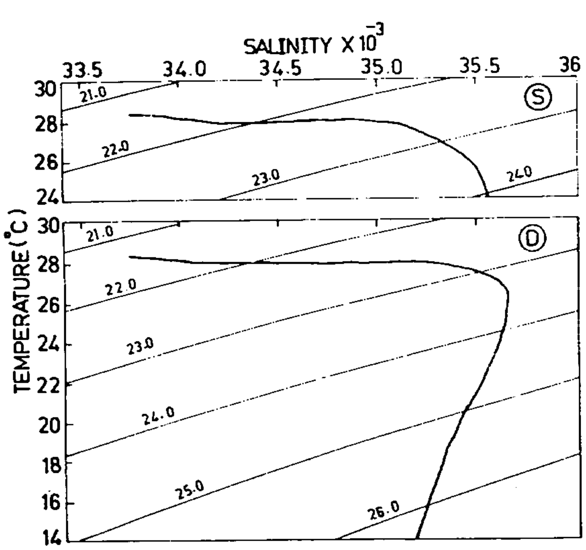

Fig. 4—T-S diagram for shallow (*S*) and deep (*D*) stations

ian Sea High Salinity Watermass. The depth of occurrence of the thermal inversions (Fig. 3) approximately coincides with the core of Arabian Sea High Salinity Watermass (Fig. 4). Hence, it is inferred that the thermal inversion noticed around 75 m is caused by the intrusion of this high salinity water to the continental shelf.

References

1 Hareesh Kumar P V & Rao R R, *Indian J Mar Sci*, 16 (1987) 71.
2 Lumb F E, *Quart J Royal Met Soc*, 90 (1964) 43.
3 Atwater M A & Ball J T, *Monthly Weather Review*, 109 (1981) 880.
4 Dobson F W & Smith S D, *Quart J Roy Met Soc*, 114 (1988) 165.
5 Wyrtki K, *Hawaii Ins Geophy Rep HIG*-66-3, (University of Hawaii, Honolulu) 1966, pp. 8.
6 Monin A S & Obukhov A M, *Trans Geophys Inst Akad*, 151 (1954) 163.
7 Ostapoff F & Worthem S, *J Phys Oceanogr*, 4 (1974) 601.
8 Phillips O M, *The dynamics of the upper ocean*, (Cambridge University Press, Cambridge) 1977, pp. 336.
9 Darbyshire M, *Deep-Sea Res*, 14 (1967) 295.