# The Role of Surface Mixing in the Seasonal Variation of the Ocean Thermal Structure

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ABSTRACT

The role of surface-generated mixing in determining the seasonal variation of the ocean thermal structure is investigated using a one-dimensional numerical model. The model contains vertical eddy diffusion with a constant coefficient $K_H = 0.5$ cm$^2$ s$^{-1}$, an instantaneous convective adjustment mechanism as commonly used in oceanic general circulation models, and a simple parameterization of surface-generated wind and convective mixing based on recent mixed-layer theories. Forcing on the seasonal time scale is accomplished by prescribing the atmospheric solar radiation, longwave radiation, wind speed, temperature and dew point to vary sinusoidally with the annual period. Results of model integrations show that surface-generated wind and convective mixing are responsible for producing many features which are observed in the real ocean including the occurrence of two sea surface temperature maxima—one in summer and another in early fall.

## 1. Introduction

In a recent paper, Dorman *et al.* (1974) present an analysis of 20 years of meteorological data and 7 years of oceanographical data from Ocean Station November (30°N, 140°W). Fig. 1, taken from their paper, is a time-depth cross section of temperature showing the average seasonal variation. The data are from BT's averaged over the years 1964–70 and smoothed by a two-week running mean. The purpose of this paper is to show that many of the features in Fig. 1 can be explained by a simple model containing the effects of surface-generated wind and convective mixing.

## 2. The model

The model is a one-dimensional model which is designed to test a parameterization of wind mixing for use in an oceanic general circulation model (Haney, 1974). Since the main objective in the test model is to examine the effect of surface mixing on the seasonal variation of temperature, vertical advection is neglected. Horizontal advection and horizontal diffusion are also neglected so the first law of thermodynamics is

$$\frac{\partial T}{\partial t}=\frac{\partial}{\partial z}\left[S+K_H\frac{\partial T}{\partial z}-\overline{(w'T')}\right]+\delta_c(T), \quad (1)$$

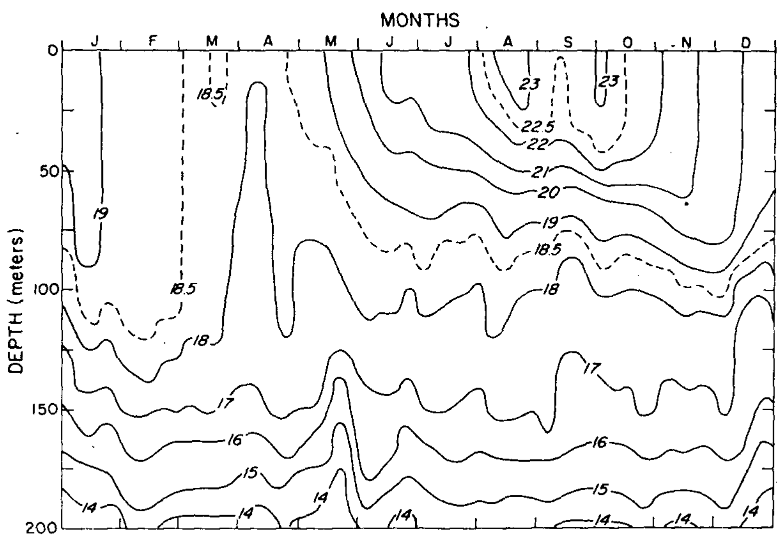


FIG. 1. Seasonal depth cross section of water temperature (°C) at Ocean Station November (from Dorman *et al.*, 1974).

where $t$ is time, $z$ height increasing upward, $T$ temperature, $\rho_0 cS$ the downward flux of solar radiation in the sea, $c$ the specific heat, $\rho_0$ the (constant) density, $K_H$ the (constant) vertical eddy diffusivity, and $\rho_0 c(\overline{w'T'})$ the upward flux of heat due to surface-generated wind and convective mixing. The last term in (1) models deep convective mixing and is a simple instantaneous adjustment of the lapse rate to isothermal which is made if, in the absence of any adjustment, the lapse rate were to become unstable. A model similar to (1), but with no surface-generated mixing, was used by Wetherald and Manabe (1972) to explain results in their coupled ocean-atmosphere general circulation model.

The parameterization of surface mixing is based upon the recent mixed layer theories of Kraus and Turner (1967), Turner (1969) and Denman (1973). Since the instantaneous convective adjustment parameterizes the convective mixing which occurs in the case of an unstable lapse rate, the primary objective is to parameterize the effects of surface-generated wind and convective mixing in the case of a stable lapse rate. Under the assumption that the temperature is independent of $z$ between the sea surface and the bottom of a well mixed layer of depth $h$, integrations of (1) give

$$\tfrac{1}{2}[(\overline{w'T'})_0+(\overline{w'T'})_{-h}]=\frac{1}{h}\int_{-h}^{0}(\overline{w'T'})dz +\frac{1}{2}\left[\left(S+K_H\frac{\partial T}{\partial z}-F_c\right)_0+\left(S+K_H\frac{\partial T}{\partial z}-F_c\right)_{-h}\right] -\frac{1}{h}\int_{-h}^{0}\left(S+K_H\frac{\partial T}{\partial z}-F_c\right)dz, \quad (2)$$

where $\rho_0 cF_c$ is the upward flux of heat due to the convective adjustment defined by $\delta_c(T)=-\partial F_c/\partial z$. Following Kraus and Turner (1967) a steady-state balance between the generation $G^*$, dissipation $D^*$, and the transformation of potential energy into kinetic energy is assumed for the turbulence in the mixed layer. Thus

$$\int_{-h}^{0}(\overline{w'T'})dz=D-G, \quad (3)$$

where $D=D^*/(\alpha\rho_0 g)$, $G=G^*/(\alpha\rho_0 g)$ and $\alpha=-\rho_0^{-1}d\rho/dT$ is the coefficient of thermal expansion. Solving (2) for $(\overline{w'T'})_{-h}$ and using (3) gives

$$(\overline{w'T'})_{-h}=-\frac{2}{h}(G-D)-(\overline{w'T'})_0+\left(S+K_H\frac{\partial T}{\partial z}-F_c\right)_0 +\left(S+K_H\frac{\partial T}{\partial z}-F_c\right)_{-h} -\frac{2}{h}\int_{-h}^{0}\left(S+K_H\frac{\partial T}{\partial z}-F_c\right)dz. \quad (4)$$

If $G$ and $D$ are given and $h$ is known, the surface-generated heat flux at the bottom of the mixed layer can be obtained from (4). The first term on the right-hand side of (4) gives the downward flux of heat due to an excess of surface wind mixing over dissipation while the second term gives that due to surface-generated convective mixing. The remaining three terms are due to the combined downward flux of heat by solar radiation, conduction and convective adjustment within the mixed layer. The sum of these terms is proportional to the departure of the combined flux from a linear $z$ dependence and is usually positive. Since this tends to reduce the magnitude of $(\overline{w'T'})_{-h}$, these terms together represent a stabilizing effect. The stabilizing effect due to the exponential absorption of solar radiation with depth is known to be significant in mixed-layer models (Denman, 1973), while the other effects may be small by comparison. As a simplification therefore, only the contribution due to solar radiation is retained in the last three terms in (4). Thus, the surface-generated upward flux of heat at $-h$ is given by

$$(\overline{w'T'})_{-h}=-\frac{2}{h}(G-D)-(\overline{w'T'})_0+(S_0+S_{-h}) -\frac{2}{h}\int_{-h}^{0}Sdz. \quad (5)$$

Between the sea surface and the depth $h$, $(\overline{w'T'})$ is determined in such a way that

$$\frac{\partial}{\partial z}[S-(\overline{w'T'})]=\text{constant}. \quad (6)$$

Thus,

$$(\overline{w'T'})=(\overline{w'T'})_0+\frac{z}{h}[(\overline{w'T'})_0-(\overline{w'T'})_{-h}] +S-S_0-\frac{z}{h}(S_0-S_{-h}). \quad (7)$$

The mechanical energy available for mixing, $G^*-D^*$, must be expressed in terms of known surface parameters; however, there is considerable uncertainty as to what the proper relationship is. As pointed out by Turner (1969), what is actually needed from careful observations is a direct relationship between the wind and the rate of change of potential energy in the mixed layer. Turner (1969), Kato and Phillips (1969) and Denman (1973) used an expression of the form

$$G^*-D^*=m\tau V_A, \quad (8)$$

where $\tau$ is the surface stress, $V_A$ the wind speed at 10 m, and $m$ a coefficient of proportionality whose value is still not accurately known. From observations in the open ocean, Turner (1969) obtained the value $m=0.01$, while in laboratory experiments Kato and Phillips (1969) obtained a smaller value $m=0.0015$. In the original Kraus and Turner model, a velocity scale equivalent to $m=0.00125$ was used while Denman's model gave realistic results only with this smaller value. In the present model we follow Kraus and Turner (1967) and define

$$G^*-D^*=\rho_0 w_*^3, \tag{9}$$

where $w_*$ is the friction velocity of the water.

The mixed-layer depth $h$ is obtained by means of a simplifying approximation. As noted above, the primary objective of this parameterization is to model the effects of wind mixing in the stable case. From physical considerations and from the Kraus and Turner mixed-layer theory, the appropriate scale depth in this case is the Monin–Obukhov length scale (Monin and Obukhov, 1953; Kitaigorodskii, 1960; Phillips, 1966). Therefore, as a first approximation to a more complete theory, the mixed-layer depth $h$ is assumed to be the smaller of the Monin–Obukhov length scale and the neutral planetary boundary layer length scale. Thus,

$$\left.\begin{aligned} h&=\min\left(L,\frac{w_*}{f}\right)\\ L&=|w_*^3/\kappa\alpha g[S_0-(\overline{w'T'})_0]\end{aligned}\right\}, \tag{10}$$

where $\kappa$ is the von Kármán constant and $f=10^{-4}$ s$^{-1}$ is the Coriolis parameter. In the unstable case, i.e., when $[S_0-(\overline{w'T'})_0]<0$, deep mixing by the convective adjustment mechanism is expected to dominate the vertical transfer of heat and therefore in this case the surface-generated heat flux $(\overline{w'T'})$ is completely neglected. Eqs. (5), (7), (9) and (10), along with surface boundary conditions, represent a closed parameterization of the surface generated heat flux $(\overline{w'T'})$. In order to prevent an upward flux of heat by wind mixing ("unmixing"), if (5) results in a positive value for $(\overline{w'T'})_{-h}$, it is then reset to zero.

The profile of solar radiation as a function of depth is taken from observations made by Paulson and Simpson (1974) during the NORPAX Pole experiment. They found that on the scale of the seasonal thermocline the flux can be closely approximated by an exponential with scale depth of 19 m. Thus $S$ is written

$$S=S_0\exp(\beta z),\quad \beta^{-1}=19\text{ m}. \tag{11}$$

In formulating the surface thermal boundary condition, the heat flux can either be prescribed or it can be calculated from the predicted sea surface temperature. Since the latter method is used in the oceanic general circulation model (Haney, 1974), it is also used in this test model. Results using the specified flux boundary condition and other evaluations of the model are given in Davies (1975). The surface thermal boundary conditions are thus

$$\left.\begin{aligned} K_H\frac{\partial T}{\partial z}&=0\\ F_c&=0\\ S_0&=S_0^*/\rho_0 C\\ (\overline{w'T'})_0&=(Q_B+Q_S+Q_E)/\rho_0 C\\ w_*&=(C_D\rho_a/\rho_0)^{\frac{1}{2}}V_A\end{aligned}\right\}. \tag{12}$$

In (12), $Q_B$ is the net upward longwave radiation, $Q_S$ the upward sensible heat flux, $Q_E$ the upward latent heat flux, $V_A$ the wind speed at 10 m, $\rho_a$ the (constant) air density, and $C_D$ the drag coefficient.

Results will be shown below for two types of experiments which differ only in the method of calculating the sensible and latent heat flux. These two methods are

$$\text{Type I:}\quad Q_S+Q_E=K(T_S-T_A), \tag{13a}$$

$$\text{Type II:}\quad Q_S+Q_E=\rho_a C_D V_A[C_p(T_S-T_A)+L_q(q_S-q_A)], \tag{13b}$$

where $C_p$ is the specific heat of air at constant pressure, $L_q$ the latent heat of vaporization, $T_A$ the air temperature, $q_A$ the specific humidity of the air, $T_S$ the ocean surface temperature predicted by the model, and $q_S$ is the saturation specific humidity at the temperature $T_S$. In Type I, $K=35$ W m$^{-2}$ K$^{-1}$ is a constant atmosphere-ocean coupling coefficient (Haney, 1971) whose magnitude is consistent with the coefficients in (13b). With this type of boundary condition, the sensible and latent heat flux depends only on the computed air-sea temperature difference. The Type II condition is a bulk aerodynamic formulation in which the sensible and latent heat flux depends on both the air-sea temperature difference and the mean wind speed.

In all the experiments, the vertical thermal diffusivity $K_H=0.5$ cm$^2$ s$^{-1}$ and the drag coefficient $C_D=1.1\times10^{-3}$. The longwave radiation is taken to be a constant while the solar radiation, the temperature and specific humidity of the air, and the wind speed vary sinusoidally over the annual cycle. The initial temperature is isothermal at 18.5°C.

The one-dimensional numerical model is 4 km deep with the same vertical finite-difference structure and calculation of vertical diffusion and vertical convective adjustment as in Haney (1974) except the present model has 10 levels. The vertical fluxes of heat are calculated at the ocean surface, at the ocean bottom, and at the interfaces between the levels. These interfaces are located at the depths of 20, 45, 80, 125, 200, 325, 600, 1200 and 2200 m which is comparable to those used in the oceanic general circulation model of Bryan *et al.* (1975). The net downward heat flux at the surface, $\rho_0 C[S_0-(\overline{w'T'})_0]$, is obtained from (12) while the flux at the bottom is zero.

Table 1. Atmospheric data used in the four experiments.

| Experiment | Boundary condition | Solar radiation (W m$^{-2}$) $\bar{S}$ | $S'$ | $t_0$ | Air temperature (°C) $\bar{T}_A$ | $T_A'$ | $t_0$ | Wind speed (m s$^{-1}$) $\bar{V}_A$ | $V_A'$ | $t_0$ |
|---|---|---|---|---|---|---|---|---|---|---|
| 1 | I | 143 | 56 | $P/2$ | 18.5 | 3.0 | $P/2$ | 6.7 | 1.5 | 0 |
| 2 | I | 143 | 56 | $P/2$ | 18.5 | 3.0 | $3P/4$ | 6.7 | 1.5 | 0 |
| 3 | II | 143 | 56 | $P/2$ | 18.5 | 3.0 | $P/2$ | 6.7 | 1.5 | 0 |
| 4 | II | 143 | 56 | $P/2$ | 18.5 | 3.0 | $3P/4$ | 6.7 | 1.5 | 0 |

## 3. Results

Results will be shown for four experiments in which the only difference is the applied forcing. In all the experiments, the longwave radiation $Q_B = 48$ W m$^{-2}$. The sinusoidal variation of the atmospheric forcing is expressed by writing the solar radiation, air temperature and wind speed in the form

$$q = \bar{q} + q' \cos\left[\frac{2\pi}{P}(t - t_0)\right], \tag{14}$$

where $\bar{q}$ is the annual mean, $q'$ the amplitude of the annual variation, $t_0$ the time when $q$ is a maximum, and $P = 365$ days is the annual period. The data used in the four experiments are shown in Table 1. The solar radiation and the wind speed are the same in all the experiments. The solar radiation varies sinusoidally in time with a maximum of 199 W m$^{-2}$ on 1 July and a minimum of 87 W m$^{-2}$ on 1 January, while the wind speed is a maximum on 1 January and a minimum on 1 July. In experiment 1, the air temperature has the same phase as the solar radiation while in experiment 2 it is one-fourth of a period ahead with the maximum air temperature occurring on 1 October. Experiments 3 and 4 are the same as experiments 1 and 2, respectively, except that Type II boundary conditions are used. For these two experiments, the dew point is specified and the specific humidity of the air, $q_A$, is calculated from it using the Clausius–Clapeyron equation (Fleagle and Businger, 1963). In experiments 3 and 4 the dew point was specified to have the same sinusoidal variation as the air temperature $T_A$ but with an annual mean value of 14.5°C. In general, the atmospheric forcing specified in experiments 2 and 4 represents as closely as possible the average seasonal atmospheric data at Ocean Station November as given by Dorman *et al.* (1974).

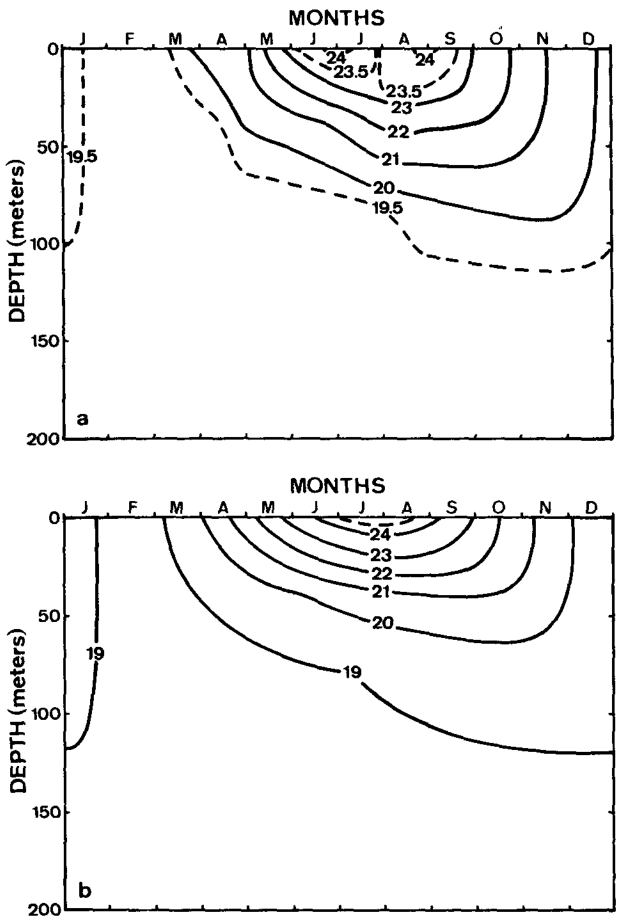


Fig. 2. Seasonal depth cross section of water temperature (°C) taken from the last year of experiment 1. Parameterization of surface-generated wind and convective mixing $(\overline{w'T'})$ is included in (a) and excluded in (b).

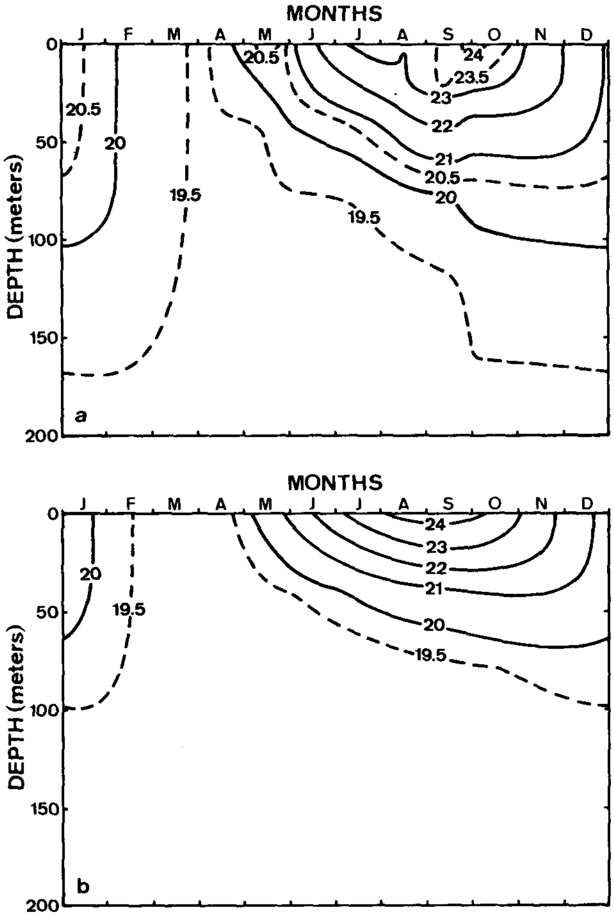


Fig. 3. As in Fig. 2 except for experiment 2.

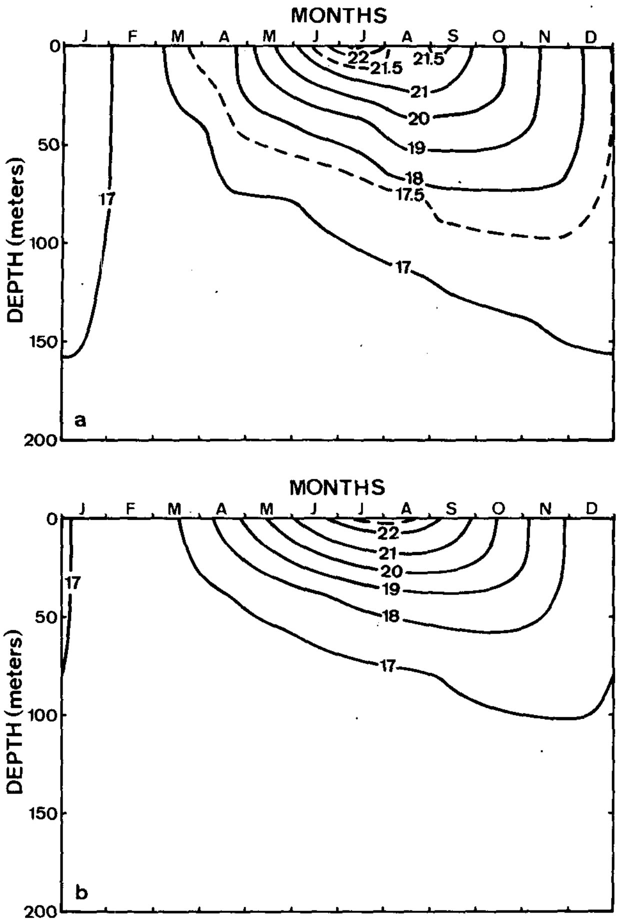


FIG. 4. As in Fig. 2 except for experiment 3.

With the above atmospheric parameters varying sinusoidally over the annual cycle, the model was integrated for 100 years at which time the seasonal variation of the ocean thermal structure was essentially repeating itself. The results from the last year of each experiment are shown in Figs. 2–5. Perhaps the most predominant feature which appears in every experiment is the marked asymmetry between the heating and cooling seasons of the year. The heating in summer penetrates downward comparatively slowly while the cooling in winter tends to occur at all depths simultaneously. This asymmetry, which is characteristic of the observed seasonal variation at Ocean Station November (Fig. 1), was also obtained in the oceanic general circulation model of Wetherald and Manabe (1972) and is due to the parameterized convective adjustment mechanism. Another prominant feature in all the experiments is the absence of a permanent thermocline; however, this is to be expected because of the neglect of vertical advection and other dynamical features of the steady thermohaline circulation.

The important effect of surface-generated wind and convective mixing is seen by comparing the upper part of each figure with the lower part of the same figure. In the experiments without surface-generated mixing (lower half of each figure) the thermal structure varies primarily on the annual period with a rather constant rate of warming in summer and cooling in winter. In the experiments with the surface-generated mixing included (upper half of each figure) the thermal structure is much closer to the observed structure. For example, in all the experiments with surface mixing there is a rather abrupt deepening of the isotherms in spring which is also characteristic of the observed pattern. Most significant, however, is the occurrence of two sea surface temperature maxima and minima which agree most favorably with the observed pattern. The occurrence of a double sea surface temperature maximum in the model, which is also an observed characteristic of many other ocean stations (Ramage, 1974), is due to the rapid mixing of cold water up into the surface layers which is produced when the wind begins to increase during the heating cycle in early fall. In experiment 4 (Fig. 5) this rapid mixing produces a correspondingly rapid drop in the surface temperature during August. Partly because of the cooler surface temperatures but largely because the air temperature in this experiment is still increasing the surface layer soon

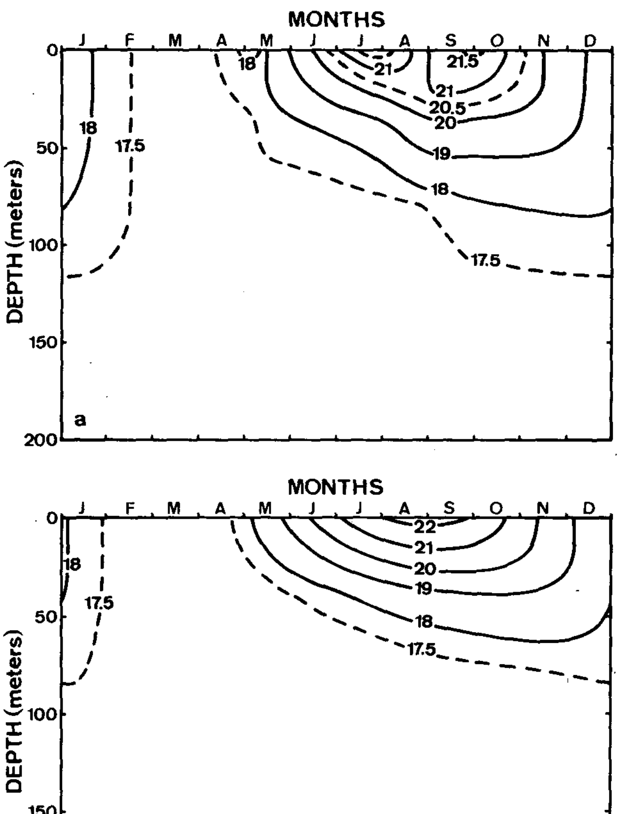


FIG. 5. As in Fig. 2 except for experiment 4.

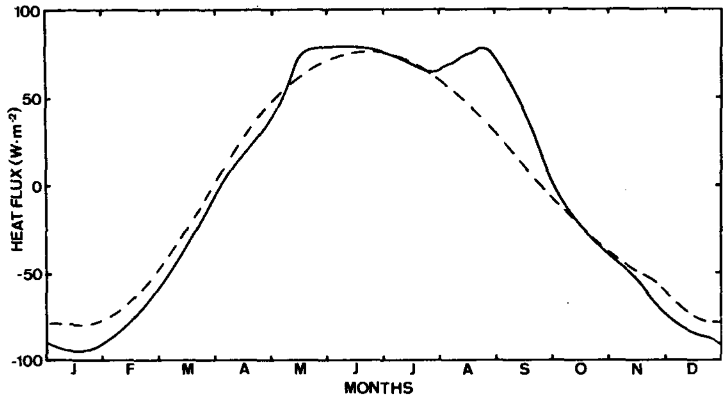


Fig. 6. Downward surface heat flux taken from the last year of experiment 4. Solid curve is from the experiment with surface mixing included and the dashed curve is from the experiment with surface mixing omitted.

begins to warm again thus producing a secondary maximum in the surface temperature near the end of September. In this model, therefore, the surface-generated wind and convective mixing play a critical role in the evolution of the ocean thermal structure.

The differences caused by the Type I and Type II boundary conditions are seen by comparing Fig. 2 with Fig. 4 and Fig. 3 with Fig. 5. One simple effect of the Type II condition is to reduce the overall temperature about 2°C because of the additional upward latent heat flux. The Type II condition essentially acts to couple the ocean to an atmospheric equilibrium temperature which is determined by both the air temperature and the (colder) dew point. A more interesting effect of the Type II condition, however, is to slightly amplify the sea surface temperature maximum which occurs in summer relative to that which occurs in early fall.

The differences caused by changing the phase of the prescribed air temperature so that its maximum occurs on 1 October (as observed at Ocean Station November) instead of 1 July are seen by comparing Fig. 2 with Fig. 3 and Fig. 4 with Fig. 5. The main effect is to shift the ocean thermal structure ahead about one month producing results (Fig. 3 and Fig. 5) which correspond more closely with the observed pattern.

The total downward flux of heat at the ocean surface calculated in experiment 4 is shown in Fig. 6. Since the area under the curves is zero, the total heat flux for the year is zero and therefore the annual mean temperature is constant from one year to the next. The most interesting feature in the heat flux curves is the double maximum in the experiment with surface mixing (solid curve) and no such maximum in the experiment without mixing (dashed curve). The maxima in the heat flux curve occur in May and late August respectively and are clearly a result of surface mixing. Such double maxima are also characteristic of the observed heat balance at Ocean Station November (Dorman *et al.*,

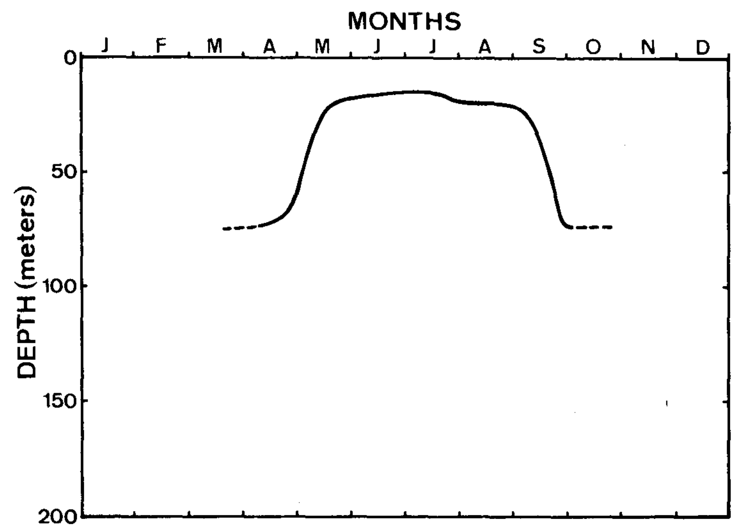


Fig. 7. Depth of surface-generated wind and convective mixing calculated from (10) for experiment 4.

1974; their Fig. 5); however it is felt that more careful observations are needed in order to determine whether or not the ocean thermal structure at the Ocean Station, including the surface heat balance, actually evolves in the same way and for precisely the same reasons as in the model.

Fig. 7 shows the depth $h$ of surface-generated wind and convective mixing as a function of time calculated from (10) for experiment 4. Since this mixing is neglected when the surface heat flux is upward, $h$ is not defined at those times. The mixing depth is largest when the heat flux changes sign in the spring and fall and clearly shows the initial deepening in August which accompanies the sea surface temperature minimum.

In conclusion it should be pointed out that the model contains several empirical parameters such as the coefficient of vertical eddy diffusion and a wind mixing parameter $m$ whose true values in the real ocean are not well known. In addition, the model is based upon a simplified diagnostic expression for the mixing depth $h$. However, the true values of the empirical parameters are perhaps not much different from the values used in the model and the expression for $h$ should be quite accurate during the stable time of the year when surface-generated mixing is most effective in influencing the thermal structure. Accordingly, at least the general qualitative features of the above results should be applicable to the real ocean and these clearly show that surface-generated wind and convective mixing play an important role in determining the detailed evolution of the seasonal temperature structure, especially that at the surface.

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