Mixed-layer modelling at ocean weathership station Bravo

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Abstract

This paper is concerned with modelling the oceanic mixed layer at a high-latitude location in the Labrador Sea. Several simulations were carried out spanning the period 1964-1971 and will be presented during the presentation; however, this study will focus on one simulation covering the years 1969-1971. The key result advocated here is that the mixed-layer model cannot simulate salinity well and this can have important implications in climate modelling.

1 Introduction

The mixed layer refers to the upper portion of the ocean that is in direct contact with the atmosphere. It is usually observed to be well mixed and is modelled as a continuously stratified fluid. The mixed layer plays an important role in communicating and negotiating fluxes with the atmosphere. One-dimensional mixed-layer models offer a means of investigating the response of the upper ocean to atmospheric forcing. In mixed-layer models a set of conservation equations governing the mean horizontal velocity components \((\bar{U}, \bar{V})\), temperature \((\bar{T})\) and salinity \((\bar{S})\) are driven by wind stresses, solar radiation and fluxes of heat and salinity imposed by the atmosphere at the ocean surface. The mixed layer responds to this forcing through deepening or shallowing and heating or cooling.

Discussed in this paper is a one-dimensional, second-order, turbulence closure scheme for the mixed layer. The model used in this investigation is a variation of the second-order turbulence closure scheme of D’Alessio,
Abdella and McFarlane [1]. The success of the model is judged by comparing simulations with observations at ocean weathership station Bravo. Station Bravo was located near the centre of the Labrador Sea as shown in Figure 1. During the period 1946-1974 surface meteorological measurements were routinely observed and recorded along with subsurface temperature and salinity profiles. As recently reported by Lilly et al. [2], the Labrador Sea is known for its deep convection brought about by the harsh climate.

2 Governing equations and boundary conditions

Applying Reynolds averaging and in the spirit of the Boussinesq approximation the following set of conservation equations emerge

\[
\frac{\partial \overline{U}}{\partial t} = f\overline{V} - \frac{\partial (u'w')}{\partial z},
\]

\[
\frac{\partial \overline{V}}{\partial t} = -f\overline{U} - \frac{\partial (v'w')}{\partial z},
\]

\[
\frac{\partial \overline{T}}{\partial t} = \frac{1}{\rho_0 c_p} \frac{\partial I}{\partial z} - \frac{\partial (\overline{T'}w')}{\partial z},
\]

\[
\frac{\partial \overline{S}}{\partial t} = -\frac{\partial (s'w')}{\partial z}.
\]

In the above system \(\overline{U}\) and \(\overline{V}\) refer to the mean horizontal velocity components in the \(x\) and \(y\) directions respectively, \(\overline{T}\) is the mean temperature and \(\overline{S}\) is the mean salinity. The term \((1/\rho_0 c_p)\partial I/\partial z\) in equation (3) represents a non-turbulent source flux due to penetrating solar radiation. Typically, this is modelled by exponential terms to take into account the attenuation at different wavelengths. For example, Paulson and Simpson [3] suggest the two-term expression

\[
\frac{I(z)}{I_0} = Re^{z/l_1} + (1 - R)e^{z/l_2},
\]

where \(I_0\) is the transmitted radiation at the surface, and the parameters \(R, l_1, l_2\) depend on the optical properties at the particular location. For station Bravo, the parameter values that best fit the water type are: \(R = 0.4, l_1 = 5(m)\) and \(l_2 = 40(m)\). The vertical coordinate, \(z\), is taken to be zero at the surface and pointing upward; \(t\) is the time coordinate, \(f = 0.00012128(s^{-1})\) is the local Coriolis parameter, \(\rho_0 = 1027(kg/m^3)\) is a constant reference density and \(c_p = 4190(J/kg^\circ C)\) is the heat capacity at constant pressure. The mean turbulent kinetic energy (TKE), \(\overline{\epsilon}\), defined as \(\overline{\epsilon} = (\overline{w'^2} + \overline{v'^2} + \overline{w'^2})/2\), is also computed by solving

\[
\frac{\partial \overline{\epsilon}}{\partial t} = -\left(\overline{u'w'}\frac{\partial \overline{U}}{\partial z} + \overline{v'w'}\frac{\partial \overline{V}}{\partial z}\right) + \overline{u'w'} - \frac{\partial}{\partial z}\left(e\overline{w'} + \frac{1}{\rho_0}P'\overline{w'}\right) - \overline{\epsilon}.
\]
Primed quantities reflect deviations from the mean and $u'w'$, $v'w'$, $T'w'$, $s'w'$, $b'w'$, $e\bar{w}'$, and $P'w'$ represent turbulent fluxes. Buoyancy is defined as

$$B = -g(\rho - \rho_0)/\rho_0 = g(\alpha \Delta T - \beta \Delta S) ,$$

where $\Delta T$ and $\Delta S$ refer to changes in temperature and salinity with respect to constant reference values.

Closure is achieved through the following expressions which are similar in form to a Mellor-Yamada [4,5] level 2.5 turbulence closure scheme

$$u'w' = -K_m \frac{\partial \bar{U}}{\partial z} , \quad v'w' = -K_m \frac{\partial \bar{V}}{\partial z} , \quad K_m = \sqrt{2} S_m l \sqrt{\bar{e}} ,$$

$$v'w' = g(\alpha T'w' - \beta s'w') ,$$

$$T'w' = -K_h \frac{\partial T}{\partial z} + \frac{1}{2} \alpha g \tau_t T'^2 , \quad K_h = \tau_t w'^2 ,$$

$$s'w' = -K_s \frac{\partial S}{\partial z} , \quad K_s = K_h ,$$

$$e\bar{w}' + \frac{1}{\rho_0} P'w' = -2a_1 \frac{\bar{e}}{\varepsilon} \cdot \frac{\partial \bar{e}}{\partial z} ,$$

$$\frac{T'^2}{\overline{\varepsilon}} = -\frac{a_2}{\sqrt{2} \sqrt{\overline{\varepsilon}}} \frac{l}{\overline{T'w'}} \frac{\partial T'}{\partial z} ,$$

$$\frac{w'^2}{\overline{\varepsilon}} = \frac{1}{2} \frac{a_4}{4\sqrt{2} \sqrt{\overline{\varepsilon}}} \frac{l}{\overline{b'w'}} .$$

Here, $K_m, K_h$ and $K_s$ represent the eddy viscosities for momentum, heat and salt respectively, $g$ the acceleration due to gravity, $\alpha, \beta$ the thermal expansion and haline contraction coefficients respectively, $\tau_t$ the turbulent time scale and $l$ the turbulent length scale. The adopted length scale is the Blackadar [6] formula

$$l = \frac{k(|z| + z_0)}{1 + k|z|/l_0} ,$$

with

$$\frac{1}{l_0} = \frac{1}{l_{MY}} + \frac{1}{l_b} ,$$

for locally stable conditions and $l_0 = l_{MY}$ for locally unstable conditions. The Mellor-Yamada asymptotic length scale, $l_{MY}$, is given by

$$l_{MY} = 0.2 \int_{-\infty}^{0} \sqrt{\overline{\varepsilon}}|z|dz/ \int_{-\infty}^{0} \sqrt{\overline{\varepsilon}}dz ,$$

and $l_b$ is the buoyancy length scale

$$l_b = \frac{\sqrt{\overline{\varepsilon}}}{N} ,$$
imposed by the underlying stable stratification having a Brunt-Väisälä frequency, \( N \), where

\[
N^2 = \frac{\partial B}{\partial z} .
\]  

(19)

The parameter \( z_0 = 0.0001 \text{(m)} \) is the roughness length. From the length scale, the time scale is easily obtained through

\[
\tau_t = \frac{a_3 l}{\sqrt{2\varepsilon}} .
\]  

(20)

The dissipation of TKE, \( \varepsilon \), is expressed according to Kolmogorov [7]

\[
\varepsilon = \frac{2\sqrt{2\varepsilon^3/2}}{a_4 l} .
\]  

(21)

Lastly, the mixed-layer depth, \( h \), is defined as the depth at which the TKE drops below \( 10^{-6} \text{(m/s)}^2 \). Model parameters appearing in the above equations assume the following values: \( S_m = 0.39, a_1 = 0.19, a_2 = 7.8, a_3 = 1.56, a_4 = 16.6 \).

The equation of state used is the approximate empirical formula proposed by Friedrich and Levitus [8]. In terms of the temperature \( T \) and salinity \( S \) this formula is given by

\[
\sigma(T, S) = C_1 + C_2 T + C_3 S + C_4 T^2 + C_5 ST + C_6 T^3 + C_7 ST^2
\]  

(22)

where the values of the constants \( C_1 - C_7 \) are given in their paper and \( \sigma \) denotes the density anomaly defined by \( \sigma = [\rho - 1000] \text{(kg/m}^3\text{)} \). The above polynomial expression ignores the effect of compressibility and is accurate for depths down to \( 2000 \text{(m)} \) which is well within the domain of this study.

The thermal expansion and haline contraction coefficients are determined from \( \sigma \) using

\[
\alpha = -\frac{1}{\rho_0} \frac{\partial \sigma}{\partial T}, \quad \beta = \frac{1}{\rho_0} \frac{\partial \sigma}{\partial S} .
\]  

(23)

Boundary conditions imposed at the surface \( z = 0 \) take the form of flux conditions. The continuity of stress gives

\[
\overline{u'w'^0} = -\frac{\tau_x}{\rho_0} , \quad \overline{v'w'^0} = -\frac{\tau_y}{\rho_0} ,
\]  

(24)

where \( \vec{\tau} = (\tau_x, \tau_y) \) represents the wind stress. The surface heat flux is given by

\[
\overline{T'w'^0} = -\frac{Q}{\rho_0 c_p} ,
\]  

(25)

where \( Q \) represents the net non-solar heat. It is made up of sensible, \( Q_s \), latent, \( Q_L \), and net long-wave, \( Q_{LW} \), heat fluxes; \( Q_s \) and \( Q_L \) are computed using the standard bulk formulae (see [9])

\[
Q_s = \rho_a C_T C_P a (T_s - T_a) U_w ,
\]  

(26)
where \( \rho_a \) is the density of air, \( c_{Pa} \) is the specific heat of air at constant pressure, \( U_w \) is the mean wind speed, \( T_s \) and \( T_a \) are the surface water and air temperatures respectively, \( L \) is the latent heat of vapourization, \( \rho_s \) and \( \rho_v \) are the water vapour densities at the saturation sea-surface temperature (SST) and air temperature respectively. The bulk transfer coefficients \( C_T \) and \( C_E \) are computed according to Smith [10]. The total heat flux is \( Q = Q_s + Q_L + Q_{LW} \).

The turbulent kinetic energy at the surface is specified as

\[
\bar{\varepsilon}_0 = \alpha_1 u_*^2 ,
\]

where \( \alpha_1 = 3.25 \). Here, \( u_* \) is the friction velocity which is related to the wind speed and stress through

\[
u_0^2 = \frac{\sqrt{\tau_x^2 + \tau_y^2}}{\rho_0} = \frac{\rho_a}{\rho_0} C_D U_w^2 ,
\]

with \( C_D \) denoting the drag coefficient which is also computed according to Smith [10]. For salinity, the flux condition is

\[
s'w' = -\overline{S}_0 (E - P^*)
\]

where \( \overline{S}_0 \) is the mean surface salinity and \( E = -Q_L/(\rho_0 L) \) and \( P^* \) are the evaporation and precipitation rates respectively. The incoming short-wave radiation, net long-wave radiation and precipitation are obtained from the National Centers for Environmental Prediction (NCEP) data.

3 Results and comparisons

The set of conservation equations (1)-(4) and (6) were numerically integrated in time using a semi-implicit procedure. To allow for accurate differencing, computations were carried out on a staggered grid whereby the turbulent fluxes, and TKE were computed midway between the grid points for the mean quantities \( \overline{U}, \overline{V}, \overline{T} \) and \( \overline{S} \). The region \(-D < z < 0\) was discretized into \( M \) unequal intervals having grid points located at

\[
z_i = -\frac{D}{\lambda} \ln[1 - \xi_i(1 - e^\lambda)] \quad \text{where} \quad \xi_i = \frac{i}{M} \quad \text{and} \quad i = 0, 1, \ldots, M.
\]

In this formula \( D \) denotes the computational boundary, while \( \lambda \) is a resolution parameter. In the limit that \( \lambda \to 0 \) it can be easily shown that the above produces equally spaced grid points having a spacing of \( \Delta z = D/M \). The depth \( D \) is chosen to lie below the base of the mixed layer, and well away from the influence of the surface forcing. At this boundary we impose the condition that all turbulent fluxes vanish along with the TKE, and that all mean quantities assume constant and undisturbed values.
The equations were then discretized using a control volume formulation with the prognostic variables treated implicitly to increase numerical stability. At each time step the turbulent fluxes were determined from the system of equations (8)-(14). Quantities such as $h$, $\tau$, $K_m$, $K_h$ and $K_s$ were computed explicitly based on information from the previous time step since these quantities are expected to vary more slowly with time. This also accelerates the convergence of the numerical scheme. At each instant in time the numerical solution procedure involved solving a linearized system of equations for each of the mean quantities having a coefficient matrix that is either tri-diagonal or block tri-diagonal. Efficient routines were then utilized to solve these linear systems of equations. Computational parameters used were $\Delta t = 900(s)$, $\lambda = -1$, $M = 200$ and a bottom located at $z = -D = -1000(m)$. With $\lambda = -1$ the grid spacing is about $3(m)$ near the surface while about $10(m)$ near the bottom. The average grid spacing is $\Delta z = 5(m)$.

The model was tested against observational data taken from ocean weather station Bravo located in the Labrador Sea at $56.5^\circ N$ and $51.85^\circ W$. With this data set several simulations were carried out over the period 1964-1971. The results to be presented correspond to one simulation conducted for the period 1969-1971. In this simulation day 0 denotes January 1st, 1969. The Bravo data set is a good one to test mixed-layer models since it offers more rigorous testing due to its high-latitude location and intense winter conditions. Also, salinity is known to play an important role in stabilizing the water column since the equation of state (22) suggests that the density of cold water is sensitive to salinity. Initial profiles used for temperature and salinity are shown in Figures 2a,b respectively. The density anomaly is shown in Figure 2c and confirms that although the initial $T$ profile is unstable the overall density is stably stratified owing to the contribution made by salinity.

The data set consists of 3-hourly surface meteorological measurements, as well as subsurface temperature and salinity profiles. The observed temperature and salinity profiles were used to initialize the temperature and salinity at the beginning of the simulation. From the meteorological data surface fluxes of heat, momentum, and evaporation were computed using the bulk formulae presented in the previous section. Fluxes required by the model between the 3-hourly measurements were obtained by interpolation. As mentioned earlier solar radiation, net long-wave radiation and precipitation were not directly measured. Instead these quantities were estimated indirectly from NCEP data.

The simulations indicate that the mixed-layer model performs reasonably well in reproducing the observed seasonal changes in SST as portrayed in Figure 3a; however, the model is not as successful in predicting the surface salinity as pointed out in Figure 3b. Modelling salinity is difficult since it is sensitive to horizontal advection and also because there is considerable uncertainty in the precipitation measurements. A salinity balance calcula-
tion has shown that advection is indeed significant during the year 1970. Also, it must be remembered that a poor prediction of salinity can have consequences on the simulated SST which in turn can impact the climate. Lastly, shown in Figure 4 is the simulated mixed-layer depth. This plot illustrates that during the period 1969-1971 deep convection was inhibited due to the famous “Great Salinity Anomaly” that passed through the region. During this period the oceanic mixed layer reached depths of about 200(m) instead of the more usual 1500(m).

4 Conclusions

In summary, this paper discussed mixed-layer modelling at ocean weather station Bravo using a second-order turbulence closure scheme. The results obtained suggest that while the model is capable of tracking the observed SST, there is room for improvement in predicting the surface salinity. This work also demonstrates the importance of salinity in high-latitude locations. Applying relaxation may be a necessary option in order to improve the agreement between the observed and simulated surface salinity. In lower latitudes this is not essential since salinity usually plays a less significant role.

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References


![Figure 1: Map of the Labrador Sea showing the location of ocean weathership station Bravo (taken from Lazier [11]).](image-url)
Figure 2a: Initial temperature profile.

Figure 2b: Initial salinity profile.
Figure 2c: Initial density anomaly profile.

Figure 3a: Comparison of observed and simulated SST.
Figure 3b: Comparison of observed and simulated surface salinity.

Figure 4: The simulated mixed-layer depth.