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著者	Hanawa Kimio, Toba Yoshiaki
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Terms Governing Temperature and Thickness of the Oceanic Mixed Layer and Their Estimates for Sea Area South of Japan

KIMIO HANAWA and YOSHIAKI TOBA

Geophysical Institute, Faculty of Science, Tôhoku University
Sendai 980, Japan

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Abstract: Governing equations of temperature and thickness of the oceanic mixed layer have been formulated using a slab model which allows horizontal inhomogeneities of temperature, velocity and thickness of the mixed layer. Using the terms included in the above equations, a brief classification of physical processes of the upper ocean has been made. The orders of magnitude of terms in the equations are estimated for the sea area south of Japan (the area near the OWS-Tango), synthesizing published data and past investigations. When the time and spatial scales of 100 days and 1000 km, respectively, are selected (for the regional sea scale), the temperature change of the mixed layer is found to be mainly due to net heat exchange through the sea surface, entrainment at the bottom of the mixed layer and to horizontal mixing, whereas for a smaller scale such as the synoptic scale, horizontal advection becomes significant rather than horizontal mixing. In either case, the above results suggest that three-dimensional processes are important in the variation of the mixed layer in the area adjacent to the Kuroshio.

1. Introduction

The influence of oceanic conditions upon climate is given through the sea surface temperature (SST), as the lower boundary condition to the atmosphere. The SST can macroscopically be considered as the mixed-layer temperature above the seasonal thermocline. Consequently, studies of the dynamics which govern the mixed-layer variability are indispensable for the elucidation of the effect of the ocean on climatic change. It should also be noted that differential variabilities in the mixed layer conditions may cause variabilities of the oceanic side, as suggested by Toba *et al.* (1982).

Recently, a number of vertically one-dimensional models of the mixed layer have been proposed, including bulk type models (e.g., Kraus and Turner, 1967; Denman, 1973; Pollard *et al.*, 1973; Niiler, 1975) and continuous type models (e.g., Mellor and Durbin, 1975; Marchuk *et al.*, 1977; Kondo *et al.*, 1979; Kundu, 1980; Simpson and Dickey, 1981). These modelers have emphasized the agreement between their prediction and the observed deepening and temperature change of the mixed layer. The periods of the experiments, however, were rather short and the data were obtained in areas where advection was absent or small. The question arises whether such one-dimensional models can be applied to one particular sea area south of

Japan where the Kuroshio flowing nearby transports a large amount of heat, and where the loss of heat from the ocean is the largest in the North Pacific Ocean (e.g., Defant, 1961; Wyrtki, 1965).

The purpose of this paper is first to derive more realistic governing equations of temperature and thickness of the mixed layer. In this derivation, we will partly follow bulk models by Denman (1973) and others, but the assumption of horizontal homogeneities of temperature, velocity and thickness will be removed. Secondly, an estimate of orders of magnitude of terms in the temperature equation will be given for one particular sea area south of Japan.

2. Equations for temperature and thickness of the mixed layer

(1) *Derivation of equations*

As the local structure of the mixed layer, we assume a "slab model" of which the vertical structure is illustrated in Fig. 1, one which has been adopted in most of the bulk models. The mixed layer is assumed to have a thickness of $h(x, y, t)$ with bulk temperature which is vertically homogeneous: $T_s(x, y, t)$, where x and y are the horizontal coordinates and t is time. The horizontal velocity $\boldsymbol{V}_s(x, y, t)$ is also vertically homogeneous, except for the wind stirring layer of a thickness of $\delta'(\ll h)$. Below the mixed layer exists the transition layer (entrainment zone) with a thickness of $2\delta(\ll h)$ and the temperature and horizontal velocity abruptly decrease to the values of the lower layer, T_b and $\boldsymbol{V}_b$, respectively. The lower layer is assumed to be quiet, that is, the turbulent fluxes of heat and momentum are negligible.

The conservation equation of temperature T and the equation of continuity for an incompressible fluid with velocity $\boldsymbol{V}$ are written as follows,

$$\frac{\partial T}{\partial t} + \boldsymbol{V} \cdot \nabla T + w \frac{\partial T}{\partial z} = (\rho_0 C_p)^{-1} q\,, \tag{1}$$

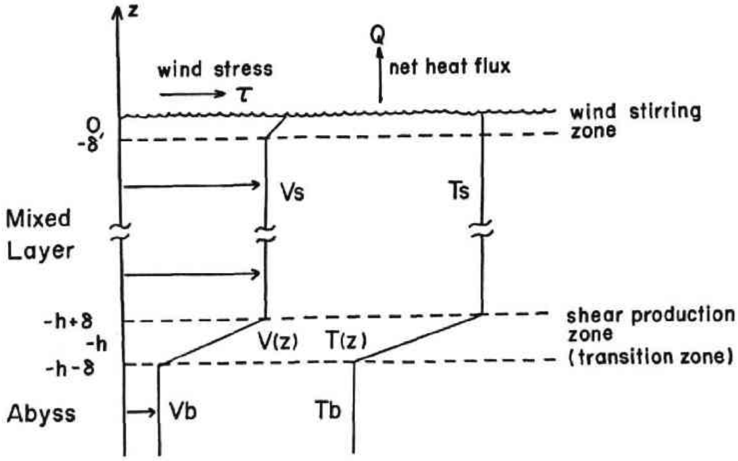

Fig. 1. Schematic picture of the vertical structure of the present mixed-layer model.

and

$$\nabla \cdot \boldsymbol{V} + \frac{\partial w}{\partial z} = 0 , \tag{2}$$

where $\nabla = \boldsymbol{i} \frac{\partial}{\partial x} + \boldsymbol{j} \frac{\partial}{\partial y}$ is the horizontal gradient operator, z the vertical coordinate taken positive upwards, ρ_0 the reference density of sea water and C_p the specific heat at a constant pressure; q indicates the heat source due to the absorption of the short wave solar radiation and is expressed by the form (Denman, 1973),

$$q = -\gamma Q_S \exp(\gamma z) , \tag{3}$$

where Q_S represents the incident net solar radiation (the penetrative component, positive upward) at the ocean surface, and γ the extinction coefficient.

When T, $\boldsymbol{V}$, w and q are expressed by the sums of the values averaged with respect to the time and the flucutations: $T=\bar{T}+T'$, etc., and are substituted into (1), the equation for mean temperature is obtained with the aid of (2), as

$$\frac{\partial T}{\partial t} = \frac{q}{\rho_0 C_p} - \nabla \cdot \boldsymbol{V} T - \frac{\partial}{\partial z}(wT) - \nabla \cdot \overline{\boldsymbol{V}'T'} - \frac{\partial}{\partial z} \overline{w'T'} , \tag{4}$$

where the over bar representing the mean values is abbreviated to simplify the notations, except for the eddy-correlation quantities.

When we integrate (4) through the transition layer from $z=-h-\delta$ to $z=-h+\delta$ in order to obtain the heat flux at the lower boundary of the mixed layer, following the method first applied to the atmospheric well-mixed layer by Geisler and Kraus (1969) and to the oceanic boundary layer by Denman (1973), we obtain

$$\begin{aligned} &\frac{\partial}{\partial t} \int_{-h-\delta}^{-h+\delta} T dz - T_s \frac{\partial}{\partial t}(h-\delta) + T_b \frac{\partial}{\partial t}(h+\delta) \\ &= -\nabla \cdot \int_{-h-\delta}^{-h+\delta} \boldsymbol{V} T dz + \boldsymbol{V}_s T_s \cdot \nabla (h-\delta) - \boldsymbol{V}_b T_b \cdot \nabla (h+\delta) - T_s w \big|_{-h+\delta} \\ &\quad + T_b w \big|_{-h-\delta} - \nabla \cdot \int_{-h-\delta}^{-h+\delta} \overline{\boldsymbol{V}'T'} dz + \overline{\boldsymbol{V}'T'} \big|_{-h+\delta} \cdot \nabla (h-\delta) - \overline{w'T'} \big|_{-h+\delta} , \end{aligned} \tag{5}$$

where the term of the short-wave solar radiation is omitted. This treatment is considered to be appropriate when the mixed layer is deeper than twenty meters, since the existinction coefficient is smaller than $(20\,\mathrm{m})^{-1}$ for most of the spectral bands of short wave solar radiation (Woods, 1980).

When the entrainment occurs under the condition convoluted by large-scale motion, the location of the bottom of the mixed layer is specified by

$$\frac{d}{dt}(-h) = w \big|_{-h} + w_e , \tag{6}$$

where w_e is the entrainment velocity (Phillips, 1972). At the top and the bottom of the transition layer, (6) may be formally rewritten as follows:

$$\frac{\partial}{\partial t}(h-\delta) = -\boldsymbol{V}\,|_{-h+\delta}\cdot\nabla(h-\delta) - w\,|_{-h+\delta} - w_e\,, \tag{6a}$$

$$\frac{\partial}{\partial t}(h+\delta) = -\boldsymbol{V}\,|_{-h-\delta}\cdot\nabla(h+\delta) - w\,|_{-h-\delta} - w_e\,. \tag{6b}$$

Substituting $T_s\times$(6a)$-T_b\times$(6b) into (5), and taking the limit of $\delta\to 0$, we obtain the turbulent heat flux specified by the entrainment velocity as

$$(T_s - T_b)w_e = \overline{V'T'}_+\cdot\nabla h + \overline{w'T'}_+\,, \tag{7}$$

where the subscript $+$ specifies the value at the top of the transition layer.

Lastly, the integration of (3) through the mixed layer, i.e., from $z=-h+\delta$ to $z=0$, taking the limit of $\delta\to 0$, and the substitution into it of (7), give the final form

$$\frac{\partial}{\partial t}(T_s) = h^{-1}[-(\rho_0 C_p)^{-1}Q + (T_s - T_b)\,w_e - h\boldsymbol{V}_s\cdot\nabla T_s - \nabla\cdot h\overline{\boldsymbol{V}_s' T_s'}]\,, \tag{8}$$

(T–1) (T–2) (T–3) (T–4)

where Q is the net heat flux through the ocean surface, and is the sum of the short-wave radiation (Q_S), the long-wave back radiation (Q_B), the heat fluxes by the latent heat (Q_E) and the sensible heat (Q_C). The Q_S, etc., are expressed by,

$$-Q_S(\rho_0 C_p)^{-1} \sim \int_{-h+\delta}^{0} (\rho_0 C_p)^{-1} q\,dz\,,$$

and

$$-(Q_B + Q_E + Q_C)(\rho_0 C_p)^{-1} = -\overline{w'T_s'}\,]_0\,.$$

The thickness of the mixed layer is given by the combination of (6a) and (2) and after taking the limit of $\delta\to 0$, by

$$\frac{\partial h}{\partial t} = -w_e - h\nabla\cdot\boldsymbol{V}_s - \boldsymbol{V}_s\cdot\nabla h\,. \tag{9}$$

(H–1) (H–2) (H–3)

It is worthwhile to note here that the equation of heat balance or heat content obtained by the combination of (8) and (9) does not always have a physical meaning. When (8) and (9) are combined, the following equation for heat content is obtained,

$$\frac{\partial}{\partial t}(hT_s) = -Q(\rho_0 C_p)^{-1} - w_e T_b - h\boldsymbol{V}_s\cdot\nabla T_s - hT_s\nabla\cdot\boldsymbol{V}_s - T_s\boldsymbol{V}_s\cdot\nabla h - \nabla\cdot h\overline{\boldsymbol{V}_s' T_s'}\,.$$

This equation is under the influence of measuring reference temperature as Montgomery (1974) pointed out in connection with the work by Niiler and Richardson (1973) on the heat transport of the Florida Current. When a new temperature scale $T_r = T + a$ (a is an arbitrary constant) is chosen and is substituted into the above equation, the term of the left hand side and the 2nd, 4th and 5th terms of the right hand side change form. Namely, the rate of contribution of the terms of the right hand side to the local change of heat content depends on the temperature scale selected. This fact is caused by the volume change of the water column of the mixed layer under consideration. Consequently, when we want to examine temporal changes of the heat content of the upper ocean, a column of a fixed thickness must be taken in order to estimate appropriately the magnitudes of terms.

Equations (8) and (9) express the variation of the mixed layer at a fixed point. In cases when stationary eddies or meanders of the mean flow pattern exist in the sea area to be considered, several correlation terms among the spatially fluctuating quantities should be included in the equations for quantities averaged with respect to the space, in addition to the terms already described. In the case when there is no such stationary flow pattern, (8) and (9) may be regarded as applicable to quantities averaged spatially as well as temporally.

(2) *Classification of upper ocean processes*

If the process of the mixed-layer variation is one dimensional, the mixed-layer temperature is determined only by the net heat exchange through the sea surface and the entrainment of water of the lower layer, and the thickness of the layer increases only through the entrainment process. However, when the horizontal inhomogeneities of temperature, velocity and thickness exist, three-dimensional processes are incorporated.

We now give the classification of the upper ocean processes based on the equation of mixed-layer thickness (9). The first term (H–1) of the right hand side of (9), is the entrainment term, and efforts to parameterize it have been made in various bulk models. The energy inflow, which induces the turbulent motion and entrainment at the bottom of the mixed layer, is performed through two ways. One is heat exchange at the sea surface, that is, the inflow of potential energy, which creates free or thermal convection. The other is wind stress, that is, the inflow of kinetic energy. The latter is also divided into two types according to the nature of occurrence of turbulence. One is the surface-originated turbulence, and includes direct stirring by wind waves and a relatively organized motion of the Langmuir circulation. These motions may be termed as "forced convection" as contrasted with "free convection". In the other, kinetic energy first enters the mean motion to form an Ekman current or an inertial oscillation; these currents have shear near the pycnocline, and the turbulence occurs owing to the shear instability. The entrainment term, therefore, represents the above four processes.

The second term (H–2) describes the effect of divergence (or convergence) of the horizontal velocity, caused by Ekman pumping and internal wave or tide. The Ekman pumping is generated by the curl of wind stress.

The third term (H–3) expresses the effect of movement of the inclined thermocline, owing to the ageostrophic motion including the passing of eddies by the mean flow.

Fig. 2 shows schematically this classification. Though the processes are presented

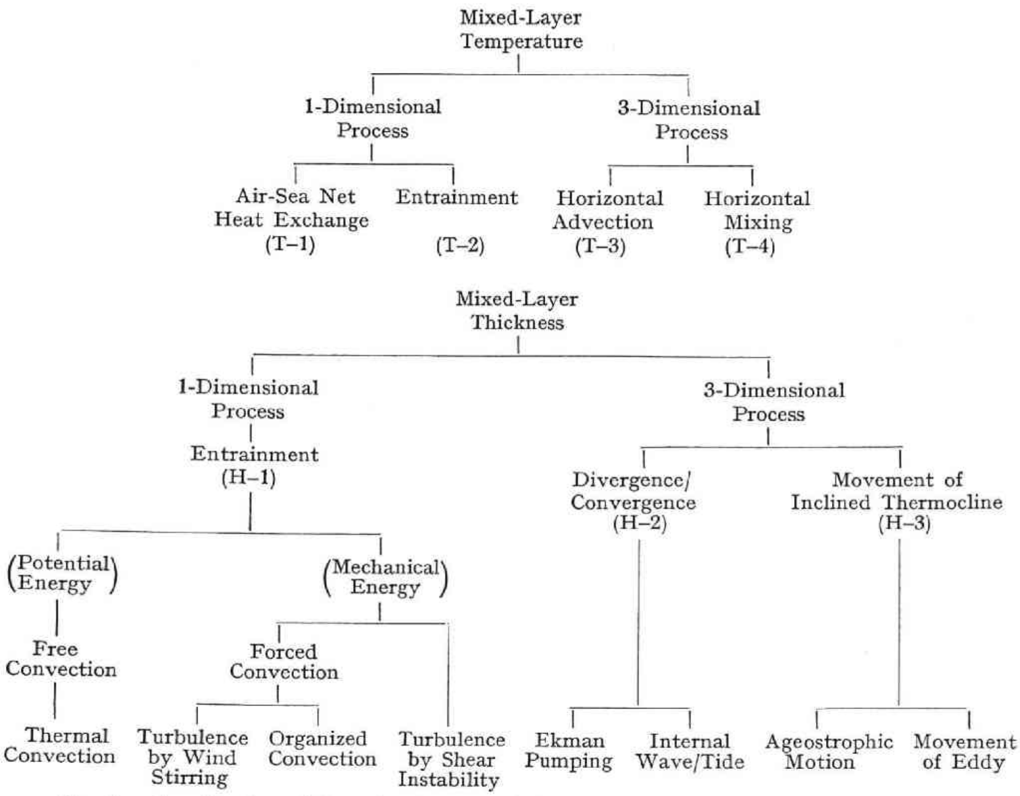

Fig. 2. Classification of the mixed-layer variation. The number in the parentheses corresponds to the term in eqs. (8) and (9).

side by side here, the essential processes for the actual mixed-layer variation will be different for seasons and sea areas. For example, for the mixed layer deepening in the temporal and spatial scales of atmospheric disturbance, the entrainment due to the surface-originated turbulence is the dominant process, as already shown by many simulations in vertically one-dimensional bulk models (e.g., Denman, 1973). The entrainment due to the free convection is important for a time scale of one day, and is also essential during the cooling season (e.g., Woods, 1980). When the large-scale wind field varies, the Ekman pumping must be considered as corresponding to the scale of variability of the wind field (de Szoeke, 1980). As for the internal tides, when averaged over a period longer than one day, their effect becomes negligible.

3. Order estimate of terms for the sea area south of Japan

Miller (1980) developed a convective adjustment model of the mixed layer deepening which was suitable for the cooling season, and simulated the SST variation in the North Pacific to obtain the values in January with the climatological initial temperature profiles from BT data obtained in September. The results showed a good agreement with the observed SST distribution except for several areas such as the Kuroshio

Extension and the Alaska Bay. Gill and Niiler (1973) also asserted that the SST was mainly controlled by the local air-sea heat exchange in the mid ocean. The above two examples state that the SST, as the representative of the mixed-layer temperature, is almost entirely determined by one-dimensional processes. However, as the sea area south of Japan is located adjacent to the Kuroshio which transports a large amount of heat (e.g., Pullen, 1977), it remains to be examined whether or not the temperature in the seasonal zone is determined mainly by the one-dimensional processes.

In this section, the orders of magnitude of terms representing the one-dimensional processes in eq. (8) will be estimated for the sea area south of Japan as shown in Fig. 3. The data used are based on complied climatological data, e.g., monthly averaged values. The area to be estimated has the spatial scale of the order of 1000 km, including the Ocean Weather Station Tango (hereafter written OWS-T) located at 29°N and 135°E. The time scale of 100 days is chosen corresponding to this spatial scale. First, we will present the data, and next the orders of magnitude for the warming and the cooling seasons will be examined.

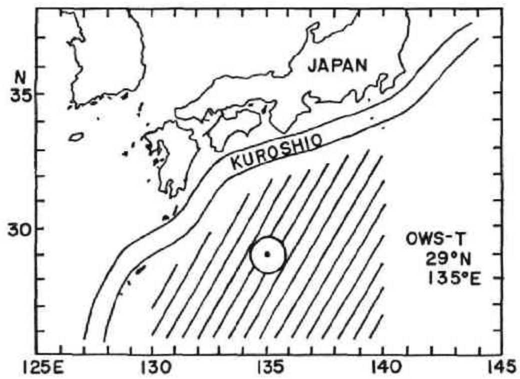

Fig. 3. Sea area south of Japan. The circle represents the Ocean Weather Station Tango (OWS-T) and the hatched area the test area.

(1) *Presentation of data*

a. Heat exchange

The distribution of yearly-averaged net heat exchange has been obtained, for example, by Defant (1961), Budyko (1956) and Wyrtki (1965) in the North Pacific Ocean. The Kuroshio area south of Japan is the region where an extremely large amount of heat is transported to the atmosphere in the North Pacific. The sea loses a quantity of heat of about 150–200 ly day^{-1} on average, which is equivalent to a temperature decrease of 3° or 4°C in the uppermost 200 m water column per year. The seasonal change estimated by various authors within or near this area is listed in Table 1. The regional average value in the warming season (June through August) is about -250 ly day^{-1} (heat gain of the sea), and in the cooling season (November through January) it is $+440$ ly day^{-1}.

Table 1. Net heat exchange (ly day^{-1}) in the sea area south of Japan. Positive values show the heat flux from the sea to the air.

Region	Month												Yearly mean
	Jan.	Feb.	Mar.	Apr.	May	June	July	Aug.	Sep.	Oct.	Nov.	Dec.	
OWS-Tango * (29°N, 135°E)							−206	−293	−181	229			
OWS-Tango **	569	391	213	77	−17	−205	−192	−150	−25	261	379	540	153
32°–34°N *** 135°–140°E	1120	360	170	10	−140	−190	−260	−140	10	420	440	640	210
24°–26°N *** 130°–135°E	580	300	−20	−30	−280	−300	−110	−260	−140	230	0	270	20
24°–26°N *** 140°–145°E	550	290	−110	−200	−420	−370	−370	−290	−200	90	0	250	−30
mean	700	340	60	−40	−210	−270	−230	−230	−110	250	200	430	90

* mean values of 1957–1960, after Maeda (1965)
** mean values of 1950–1953, after Kurasawa *et al.* (MS)
*** mean values of 1962–1963, after Hishida and Nishiyama (1969)

b. Mixed-layer temperature

The seasonal change of temperature of OWS-T was obtained by various authors (e.g., Koizumi, 1956; Kawai, 1972; Kurasawa *et al.*, MS). The temporal rates of change of the temperature in the uppermost layer are $+0.07$ and -0.07°C day^{-1} for the warming and the cooling seasons, respectively. The regional average values based on the Oceanographic Atlas of the Northwestern Pacific (JMA, 1975) and the Marine Environmental Atlas (JODC, 1975) are almost same as those of OWS-T.

The horizontal temperature gradient exists from the south toward the north in the southern region, and from the Kuroshio to the interior region in the western and northern region.* In the eastern region, the isotherms lie almost zonally. Therefore, the sea area south of Japan is the region where heat convergence by horizontal mixing in the sea is likely to occur.

c. Mixed-layer thickness

Bathen (1972) obtained the seasonal change of mixed-layer depth for each area of 2° (latitude)$\times$10°(longitude) in the North Pacific Ocean, based on about 135000 BT casts. Fig. 4 shows the depth for the sea area south of Japan from his results, drawn

* In the figures contours were obtained as a simple geographical average; the gradient near the south coast of Japan seems sometimes to be from the ocean toward the coast through the Kuroshio. This feature, however, is apparently caused by the lack of consideration of the flucutation of the Kuroshio axis. If the averaging of the temperature data were made with respect to the natural coordinates of the Kuroshio axis, as pointed out by Rossby (1981), the temperature maxima should appear in the Kuroshio area. This is evident also from continuous recordings of the surface temperature by vessels cruising across the Kuroshio. Rossby, compiling the data for the Gulf Stream, calls this high temperature belt "the warm advective core".

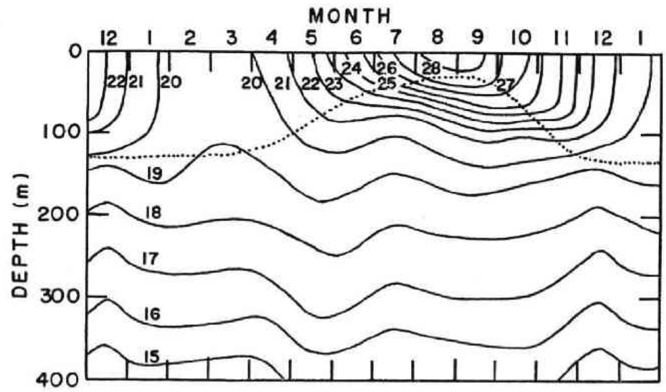

Fig. 4. Seasonal temperature variation at OWS-T redrawn from Kawai (1972). The dotted line shows the depth of mixed layer cited from Bathen (1972).

on the temperature fields of OWS-T by Kawai (1972). In the warming season, the mixed-layer thickness is about 30 m, and from September, the mixed layer begins to deepen at a rate of 15 m month^{-1}, and has a maximum depth of about 130 m in February or March. Average value in the cooling season can be taken as 100 m.

d. Entrainment velocity

Tabata *et al.* (1965) investigated the relationship between the depth of daily thermocline and the wind speed averaged for 12 hours based on the data of OWS-Papa (50°N, 145°W). Their results show the deepening of 15 to 20 m per half a day for the wind speed of 10 m s^{-1}. Maeda (1965) studied a similar relation for OWS-T and for Victory (34°N, 164°E) in the summer season, and also reproted the same deepening rate for atmospheric disturbance of a duration of two or three days with the maximum wind speed above 10 m s^{-1}. If these deepenings are regarded as the result of the entrainment process only, the entrainment velocity should have an order of magnitude of 5×10^{-5} or 10^{-4}m s^{-1}.

In the cooling season, on the other hand, if the deepening of the mean seasonal thermocline described in the preceding subsection can be regarded as the result of thermal convection, the entrainment velocity becomes of the order of 6×10^{-6}m s^{-1} corresponding to the rate of 15 m month^{-1}; it is one or two orders of magnitude smaller than that for atmospheric disturbance.

The temperature difference between the mixed layer and the lower layer may be selected to be 5°C and 2°C on an average for the warming and the cooling season, respectively (cf. Fig. 4)

(2) *Order estimate of terms for one dimensional processes*

Orders of magnitude of the terms (T–1), (T–2) and the local change term are estimated as shown in Table 2, on the basis of the data described in the preceding subsection. The entrainment term has been obtained in two ways. The upper values

Table 2. Order of magnitude of terms representing one-dimensional processes. Units are in °C s^{-1}. See the text on the double values for the entrainment term.

	warming season	cooling season
$\frac{\partial}{\partial t}T_s$	$+8\times10^{-7}$	-8×10^{-7}
$-\frac{1}{h}\frac{1}{\rho_0 C_p}Q$	$+10\times10^{-7}$	-5×10^{-7}
$+\frac{1}{h}(T_s-T_b)w_e$	0	-1×10^{-7}
	-25×10^{-7}	-3×10^{-7}

show the values by use of the entrainment velocity for the variation of the mean depth of mixed layer; for the warming season, the entrainment velocity is taken to be zero. The lower values show the values by use of the entrainment velocity under atmospheric distrubance, i.e., $w_e=5\times10^{-5}$m s^{-1}, under the assumption that the distrubances occupy one-third of the estimated period.

The results show that the all three terms have an order of magnitude of 10^{-7}°C s^{-1} for both the cooling and the warming seasons, and also that the net heat flux through the ocean surface is the dominant term. Since the data used here are monthly or seasonally averaged, we may be unable to consider further details from this table. However, the following considerations may be possible. In the warming season, the upper value indicates that an excess heat flux exists compared with the temperature increase. On the other hand, under the assumption of atmospheric disturbance, which is more realistic, the heat budget is extremely imbalanced and requires heat convergence by three-dimensional processes. In the cooling season, 60 per cent of the decrease of temperature is due to heat loss to the atmosphere, and most of the rest is due to the entrainment of lower water. This estimate for the cooling season (November through January) supports the result by Miller (1980).

The imbalance among the local change term and the one-dimensional terms in the warming season suggests that the three-dimensional processes participate in it. Maeda (1971), who observed the response of the mixed layer to the passage of a typhoon by use of BT's and XBT's reported that the heat balance of the upper ocean should include the effect of three-dimensional processes. Price (1981), who analyzed the buoy data and simulated the response of the sea to a slowly moving hurricane by the use of a three-layer and three-dimensional numerical model, also reported a similar fact.

The GEK data compiled by the Japan Oceanographic Data Center (Marine Environmental Atlas, 1979) show that the mean surface absolute velocities are less than 0.4 knot, and the directions are random, though they do not cover the whole area under consideration. The stabilities of the current velocity defined by the vector mean velocity divided by the scalar mean speed are also very low here on each grid point. These facts show nonexistence of dominant flow in this area. Consequently, we can regard the convergence or divergence of heat as the result of horizontal mixing, when the temporal and spatial scales are selected as 100 days and 1000 km, respectively.

4. Discussion

Studies on the variabilities of the oceanic mixed layer are important to clarify the role of the ocean in weather or in climatic change. Vertically one-dimensional models are considered to have been established, under the assumption of horizontal homogeneities of velocity, temperature and thickness of the layer. The real ocean, however, has a three-dimensional structure. Kurasawa *et al.* (MS), for example, have shown that the variation of heat content in the uppermost 200 m water column is largely influenced by the three-dimensional structure of the upper ocean for a short-time scale, on the basis of the data analyses of the OWS–T. They have also found that, when the net heat flux and the heat content are averaged over a long term, these values are almost consistent with each other, although the heat convergence within the ocean still exists and varies seasonally. Since the seas adjacent to Japan are complicated in boundary geometries and the distribution of currents, the essential and dominant processes in the mixed-layer variation will be different for different sea areas and for seasons. These points should be clarified by further field observations and data analyses.

It should be mentioned here that the study of the mixed layer variability may also afford a clue to the study of the mechanism of seasonal variability of oceanic conditions including currents. For example, Toba *et al.* (1982) concluded that the seasonal variability of the Tsushima-Tsugaru Warm Current System is primarily controlled by the differential variability of the sea level, presumably by processes including the local air-sea hert exchange between the East China Sea and the sea area east of Tsugaru. Further work along these lines should also be expected for the regional seas adjacent to Japan.

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