Heat Balance of the Surface Layer of the Sea at Ocean Weather Station T

Yoshikazu Kurasawa†, Kimio Hanawa† and Yoshiaki Toba†

Abstract: Heat balance of the upper 200 m of the sea south of Japan is studied, by the use of marine meteorological and oceanographic data at Ocean Weather Station T (29°N, 135°E), intensively obtained from June 1950 to November 1953. Local time change of the heat content in the surface layer and the net heat flux through the air-sea interface are calculated directly from these data, and the heat convergence in the sea is estimated from their residuals. Regarding the relative importance of one- and three-dimensional processes, it is found that, on a time scale of a few days to one month, the variation of heat content depends on heat convergence in the sea, while on a seasonal time scale, the heat content is determined primarily by the heat flux through the sea surface in December through February, by heat convergence within the sea from March to May, and by both processes from June to November. It is inferred that the heat convergence in the sea is caused by advection of water masses which are bounded by sharp fronts. Spectral analysis of sea surface temperature indicates that they typically take 2 to 3 days to pass the station, and their typical size is estimated as around 20 km by assuming the typical advection velocity of water masses to be 10 cm s⁻¹.

1. Introduction

The importance of the role of the ocean in climatic changes has been drawing special attention in recent years. The ocean contributes to climatic change through air-sea interaction, especially through heat exchange between the atmosphere and the sea through the sea surface. The heat exchange depends largely on the sea surface temperature (SST), which is approximately the same as the temperature of the ocean mixed layer. It is thus important to elucidate the processes which bring about the change of temperature structure in the surface layer of the sea.

In general, the temperature structure in the surface layer of the sea is determined, on the one hand, by one-dimensional processes such as local air-sea interactions and vertical mixing in the sea, and on the other hand, by three-dimensional processes including advection by currents and horizontal diffusion by oceanic turbulence. Tully (1964) classified the North Pacific Ocean into four regions depending on the type of dominant processes. He showed that in most parts of the North Pacific Ocean the temperature structure in the surface layer is determined by local surface processes, with the exception of one region near the northwestern boundary. The seas near Japan belong to this exceptional region, where three-dimensional processes are important as well as one-dimensional ones.

In their theoretical general discussion, Hanawa and Toba (1981) formulated the equations governing temperature and thickness of the oceanic mixed layer using a slab model which allows horizontal inhomogeneities of temperature, velocity and thickness of the mixed layer, and they presented a brief classification of physical processes of the upper ocean. They also pointed out the importance of three-dimensional processes in the variation of temperature of the mixed layer in the area adjacent to the Kuroshio south of Japan, by estimating the order of magnitude of terms in the heat balance equation by synthesizing published data and the results of various previous investigations. Maeda (1965, 1971), who examined the relationship between the wind speed and the thickness of the mixed layer from measurements with
BT, also reported that the heat balance of the upper ocean must be affected by three-dimensional processes, especially after the passage of typhoons.

In this paper, first, heat flux through the sea surface and time variance of the heat content in the surface layer of the sea are evaluated by use of marine meteorological and oceanographic data obtained at Ocean Weather Station T (hereafter call OWS-T) which is located south of Japan (29°N and 135°E). Heat convergence of the sea is also determined from the residuals of the above mentioned parameters. Secondly, by comparing these three quantities, we determine which of the one-dimensional or three-dimensional processes is dominant in causing variability in the temperature structure of the surface layer of the sea. Lastly, a hypothesis regarding the processes which cause changes in the heat content of the surface layer in the region adjacent to OWS-T is presented with some discussion.

Koizumi (1956) evaluated the heat exchange at OWS-T and the amount of heat brought by ocean currents. In the present study we re-examine his data to obtain a new estimate of the heat exchange between the atmosphere and the sea, by employing new empirical formulae for heat exchange and a different thickness of the water column to calculate the heat content. The variation of heat content on various time scales is also examined in this paper.

2. The data processing

2.1. Data

From June 1950 to November 1953, intensive marine meteorological observations were made by the Japan Meteorological Agency (JMA) at OWS-T. The station lay within a circular area with a radius of 50 miles centered on 29°N and 135°E. The data were taken almost continuously every three hours. The data used in this paper are based on "Results of Marine Meteorological and Oceanographical Observations" which was published by the JMA (1952, 1953, 1954, 1955). The times of no observation, mainly due to bad weather, accounted for 0.9% of all observations, and were sufficiently small to be negligible for the purposes of our investigation. These gaps in the data have been filled by linear interpolation when the time interval of no observation was less than nine hours, i.e., the time corresponding to two successive periods of no observation, but for a few cases when the time interval was longer the data has been left blank.

The oceanographic data is composed of results of serial observations down to 1000 m taken about 500 times during the above period. The standard observation depths were 0, 10, 20, 30, 50, 75, 100, 150, 200, 300, 400, 500, 600, 800 and 1000 m. The data has been supplied by the Japan Oceanographic Data Center (JODC) in the form of magnetic tapes. The interpolation of data is made after the calculation of heat content.

2.2. Heat balance equation

The local time change in the heat content $H$ of a water column of thickness $D$ is given by

$$\frac{\partial H}{\partial t} = \rho C_p \int_{-D}^{0} \frac{\partial T}{\partial t} \, dz = Q + F,$$  \hspace{1cm} (1)

where $\rho$ is the density of water, $C_p$ the specific heat under constant pressure, $T$ the temperature and $z$ is the vertical coordinate taken as positive upward from the sea surface. $Q$ is the net heat flux through the sea surface and is given by

$$Q = Q_s + Q_n + Q_h + Q_c,$$ \hspace{1cm} (2)

where $Q_s$ is the heat exchange by downward solar radiation, $Q_n$ the effective back radiation, $Q_h$ the latent heat, and $Q_c$ is the sensible heat, and a positive sign indicates as that the sea gains heat. $F$ in (1) is the heat convergence in the sea caused by horizontal advection and/or horizontal mixing, and is given by

$$F = -\rho C_p \int_{-D}^{0} \left( \mathbf{v} \cdot \nabla T + \mathbf{v} \cdot \mathbf{w} / T \right) \, dz + Q_{-D}$$ \hspace{1cm} (3)

where $\mathbf{v}$ is the horizontal gradient operator and $Q_{-D} = \rho C_p \mathbf{w} / T |_{z=-D}$ is the heat flux through the bottom of the column, $T$ and the water velocity $\mathbf{v}$ are the time averaged values for the time scale considered, and the prime indicates the fluctuation.

$Q$ is the heat gain by one-dimensional processes involving only local air-sea interaction, while $F$ is the gain by processes within the
sea. Consequently, by comparing $Q$ with $F$, we can find which process, air-sea heat exchange or heat convergence in the sea, is more important in the heat balance of the surface layer.

Since the oceanographic data was obtained discretely by serial observations in the vertical direction, we cannot estimate the term $Q_{-D}$ exactly. However, if we select $D$ as a little deeper than the depth of the seasonal thermocline, we may assume that $\frac{\partial}{\partial z}(w^T)_{z=-D}=0$, since there is no essential vertical temperature gradient at $z=-D$. Also, $D$ must be selected to be as thin as possible, since the thicker $D$ is the larger the contribution of $F$ becomes relative to $Q$. We have thus selected 200 meters as an appropriate thickness of the water column at OWS-T, and disregard $Q_{-D}$ hereafter.

2.3. Method of calculation of heat flux and heat content

For our detailed discussion of the heat balance, daily values of $Q_S$ and $Q_A$ together with instantaneous values of $Q_E$ and $Q_C$ every 3 hours had to be evaluated. $Q_S$ is estimated by use of a formula proposed by Kondo (1967) and $Q_E$ and $Q_C$ are evaluated by use of the aerodynamic bulk coefficients under diabatic conditions after Kondo (1975). Kondo (1976) reported that estimates of $Q_E$ and $Q_C$ using these formulae agreed in general with values estimated by atmospheric budget computations by Nitta (1976) for the period of Air Mass Transformation Experiment (AMTEX). $Q_S$ cannot be estimated from Kondo’s formula which uses information about types and heights of clouds, since only data for cloud amount is available. $Q_S$ is estimated by a newly devised empirical formula which uses a correction for the solar radiation due to the effects of cloud. The detailed forms of the formulae used in the present study are shown in the Appendix.

In the present paper, we evaluate daily values of $Q_S$ and $Q_B$ by use of daily mean values of marine meteorological data, and those of $Q_E$ and $Q_C$ by a linear integration of the values estimated every 3 hours, and we also estimate monthly mean values by averaging the daily values.

Heat content of the surface layer of the sea, $H$, is evaluated from a linear integration of discrete values of temperature observed at the standard depths using the formula

$$H = \rho C_p \sum_{n=1}^{\infty} (T_n + T_{n+1})(D_{n+1} - D_n)/2,$$

where $T_n$ is the water temperature at the $n$-th standard depth and $D_n$ is the value of the $n$-th standard depth. The blanks in the time series of data of heat content are filled by linear interpolation, regardless of the time interval of the blank. Heat convergence in the sea, $F$, is determined as the residual $(\partial H/\partial t - Q)$.

3. Heat exchange between the atmosphere and the sea

The time series of the monthly mean values of $Q$, $Q_S$, $Q_B$, $Q_E$ and $Q_C$, and their average annual variation are shown in Figures 1(a) and (b), respectively. It is seen in Figure 1(a) that for $Q$ seasonal variation is dominant, though there is a small yearly fluctuation. The principal features of the seasonal variation of $Q$ is that minimum value is $-275$ W m$^{-2}$ ($-569$ ly day$^{-1}$) in January, and the maximum value is $+99$ W m$^{-2}$ in June, in other words, maxi-

![Fig. 1. Components of monthly mean heat flux at the sea surface at OWS-T from 1950 to 1953, and their average seasonal variations. For the meaning of symbols refer to Equation (2).](image-url)
mum heat loss occurs in January and maximum heat gain of the sea occurs in June.

Another interesting feature is that, in the cooling season, the heat loss through latent heat is very large. The annual range of the net heat exchange $Q$ is $-75 \text{ Wm}^{-2}$, and the annual average supply of heat by the sea to the atmosphere is $2.3 \times 10^8 \text{ Wm}^{-2}\text{s}$ ($56 \text{ kcal cm}^{-2}$) per year. This value is equivalent to a temperature fall of the water column 200 m in thickness at a rate of $2.8^\circ \text{C year}^{-1}$. However, such a large rate of temperature fall has never been found anywhere in the world. This implies that heat convergence within the sea has to be taking place in this region.

4. **Heat balance of the sea**

4.1. **Primary contribution to the annual variation of heat balance**

First we examine the characteristics of the variation in heat content of the sea. The time series of the monthly mean values and the annual variation of the heat content of the water column of 200 m thickness, $H$, is shown in Figures 2(a) and (b), respectively. It is seen in Figure 2(a) that as in the case of $Q$ the seasonal variation of $H$ is dominant, though there is a slight yearly fluctuation. It is clear from Figure 2(b) that the notable features of the seasonal variation of $H$ are that the minimum value of $1.6 \times 10^{10} \text{ Jm}^{-2}$ is seen in February and March, and the maximum value of $1.8 \times 10^{10} \text{ Jm}^{-2}$ in September and October. The annual range is $2.5 \times 10^9 \text{ Jm}^{-2}$ and it is equivalent to a temperature variation for $3^\circ \text{C}$ of the water column of 200 m thickness.

Next we evaluate the heat convergence in the sea by use of the method shown in section 2, and examine the characteristics of its variation. The time series of monthly mean values and the annual variation of the heat convergence in the sea, $F$, is shown in Figures 3(a) and (b), respectively. It is seen that in the case of $F$, the monthly fluctuation is dominant compared with the seasonal variation. However, heat convergence is noticeable from March to May and in August, and heat divergence in November.

We now examine which of the one-dimensional or three-dimensional processes is dominant in determining the local temporal changes of heat content of the sea. The monthly values in the annual variation of $\partial H/\partial t$, $Q$ and $F$ are shown in Table 1. It is clear from the values that from December to February, $\partial H/\partial t$ is determined by $Q$, or one-dimensional processes, while from

![Fig. 2. Monthly mean heat content $H$ of the upper 200 m of the water column at OWS-T from 1950 to 1953, and its average seasonal variation.](image)

![Fig. 3. Monthly mean heat convergence $F$ for the upper 200 m of the water column at OWS-T from 1950 to 1953, and its seasonal variation. Shaded areas indicate large values appearing from March to May.](image)
March to May it is determined by $F$, or three-dimensional processes, and from June to November, $Q$ and $F$ are equally important in determining $\frac{\partial H}{\partial t}$. This implies that the dependence of $\frac{\partial H}{\partial t}$ upon $Q$ and $F$ varies seasonally.

4.2. Primary contribution to the heat balance on various time scales

In this section, we reexamine the characteristics of the variation of the dependence of $\frac{\partial H}{\partial t}$ upon $Q$ and $F$ by selecting various time scales, taking the case of 1951 as an example. The daily values of $\frac{\partial H}{\partial t}$, $Q$ and $F$ are shown in Figure 4, where the scale of $Q$ in the figure is magnified 20-times, relative to the others. It is seen that the daily value of $\frac{\partial H}{\partial t}$ shows pronounced and rapid variation and the range of variation reaches $5-10 \text{ kW m}^{-2}$ in the time scale of a few days. It is clear that on this time scale, the large variation of $\frac{\partial H}{\partial t}$ cannot be caused by variations in daily $Q$, but rather is caused by heat convergence or divergence in the sea, i.e., the heat balance of the surface layer of the sea on a time scale of a few days depends mainly upon three-dimensional processes rather than one-dimensional ones.

Next, we select one month as a longer time scale of variation. The plots in Figure 5 have been made by taking a thirty-one day moving average of the daily values in Figure 4. It is seen that on this time scale, the general trend of the annual variation of $\frac{\partial H}{\partial t}$ resembles that of $Q$, i.e., the annual trend of variation of $\frac{\partial H}{\partial t}$ depends upon one-dimensional processes. However, when one looks at the fluctuations in detail, the dependence of $\frac{\partial H}{\partial t}$ upon $F$ is still large at this time scale.

One more interesting feature which can be found from Figure 5 is that, from June to December both heat convergence and divergence of the sea take place, while from January to May only convergence takes place. The latter agrees with the finding in the previous section that heat convergence is large from March to May.

Table 1. Seasonal variation in the dependence of temporal variations of heat content $\frac{\partial H}{\partial t}$ upon the air-sea heat flux $Q$ and the heat convergence in the sea $F$.

<table>
<thead>
<tr>
<th>Month</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>
<th>9</th>
<th>10</th>
<th>11</th>
<th>12</th>
</tr>
</thead>
<tbody>
<tr>
<td>$\frac{\partial H}{\partial t}$</td>
<td>-229</td>
<td>-152</td>
<td>105</td>
<td>347</td>
<td>228</td>
<td>19</td>
<td>143</td>
<td>426</td>
<td>-4</td>
<td>-166</td>
<td>-412</td>
<td>-213</td>
</tr>
<tr>
<td>$Q$</td>
<td>-271</td>
<td>-189</td>
<td>-103</td>
<td>-37</td>
<td>8</td>
<td>99</td>
<td>93</td>
<td>73</td>
<td>12</td>
<td>-126</td>
<td>-184</td>
<td>-262</td>
</tr>
<tr>
<td>$F$</td>
<td>42</td>
<td>37</td>
<td>208</td>
<td>384</td>
<td>220</td>
<td>-80</td>
<td>50</td>
<td>353</td>
<td>-16</td>
<td>-40</td>
<td>-228</td>
<td>49</td>
</tr>
</tbody>
</table>

$(\text{W m}^{-2})$

Fig. 4. Time series of daily values of local temporal variation of heat content $\frac{\partial H}{\partial t}$, the surface heat flux $Q$ and the heat convergence in the sea $F$.  

Fig. 5. Time series of daily values of local temporal variation of heat content $\frac{\partial H}{\partial t}$, the surface heat flux $Q$ and the heat convergence in the sea $F$.  

Fig. 6. Time series of daily values of local temporal variation of heat content $\frac{\partial H}{\partial t}$, the surface heat flux $Q$ and the heat convergence in the sea $F$.  

Fig. 7. Time series of daily values of local temporal variation of heat content $\frac{\partial H}{\partial t}$, the surface heat flux $Q$ and the heat convergence in the sea $F$.  

Fig. 8. Time series of daily values of local temporal variation of heat content $\frac{\partial H}{\partial t}$, the surface heat flux $Q$ and the heat convergence in the sea $F$.  

Fig. 9. Time series of daily values of local temporal variation of heat content $\frac{\partial H}{\partial t}$, the surface heat flux $Q$ and the heat convergence in the sea $F$.  

Fig. 10. Time series of daily values of local temporal variation of heat content $\frac{\partial H}{\partial t}$, the surface heat flux $Q$ and the heat convergence in the sea $F$.  

Fig. 11. Time series of daily values of local temporal variation of heat content $\frac{\partial H}{\partial t}$, the surface heat flux $Q$ and the heat convergence in the sea $F$.  

Fig. 12. Time series of daily values of local temporal variation of heat content $\frac{\partial H}{\partial t}$, the surface heat flux $Q$ and the heat convergence in the sea $F$.  

Fig. 13. Time series of daily values of local temporal variation of heat content $\frac{\partial H}{\partial t}$, the surface heat flux $Q$ and the heat convergence in the sea $F$.  

Fig. 14. Time series of daily values of local temporal variation of heat content $\frac{\partial H}{\partial t}$, the surface heat flux $Q$ and the heat convergence in the sea $F$.  

Fig. 15. Time series of daily values of local temporal variation of heat content $\frac{\partial H}{\partial t}$, the surface heat flux $Q$ and the heat convergence in the sea $F$.  

Fig. 16. Time series of daily values of local temporal variation of heat content $\frac{\partial H}{\partial t}$, the surface heat flux $Q$ and the heat convergence in the sea $F$.  

Fig. 17. Time series of daily values of local temporal variation of heat content $\frac{\partial H}{\partial t}$, the surface heat flux $Q$ and the heat convergence in the sea $F$.  

Fig. 18. Time series of daily values of local temporal variation of heat content $\frac{\partial H}{\partial t}$, the surface heat flux $Q$ and the heat convergence in the sea $F$.  

Fig. 19. Time series of daily values of local temporal variation of heat content $\frac{\partial H}{\partial t}$, the surface heat flux $Q$ and the heat convergence in the sea $F$.  

Fig. 20. Time series of daily values of local temporal variation of heat content $\frac{\partial H}{\partial t}$, the surface heat flux $Q$ and the heat convergence in the sea $F$.
5. Discussion — The inferred mechanism of horizontal mixing

It is found that variation of the heat content depends primarily on heat convergence in the sea on the time scale of a few days to one month. Also, seasonal variation of heat content is determined primarily by the heat flux from December to February, and primarily by heat convergence from March to May, while both effects are important from June to November.

As to the mechanism of variation of the heat content of the sea, the following reasoning is presented. Figure 6 shows the time series of the heat content per unit volume for each depth range of the water column at OWS-T in 1951. It is seen that the heat content varies in a very coherent way in the vertical direction. Variation of the order of 1°C occurs on a time scale of two or three days. The temperature variation of the water column is so large that \( \frac{\partial H}{\partial t} \) cannot be balanced by \( Q \), and is mostly controlled by \( F \) as was already seen in Figure 4. We may infer that small water masses of different temperature bounded by fronts were advected through the points of sampling. Though drifting of the observation vessel across the
water masses can cause similar variation, the effect of water mass advection will be dominant especially for the time scales longer than a few days since the observation vessel at OWS-T adjusted its location a few times a day. Also, if the variation were caused by up and down motion of the seasonal thermocline due to the passing of internal waves, variation would be concentrated near the thermocline, but this is not the case. The small water masses may be similar features to the one reported by Toba et al. (1983) from satellite infra-red images of the Japan Sea. It should be noted again that for a moving average of one month, \( \frac{\partial H}{\partial t} \) is still controlled by \( F \) as was seen in Figure 5.

Since the transport of the Kuroshio and the Subtropical Counter Current, which flow so as to surround the region enclosing OWS-T, is largest from March through May (Fujiiwara, 1981, White et al., 1978, Sekine and Toba, 1981), it is inferred that the heat transported by these currents converges into this region by horizontal advection and/or by horizontal mixing, and consequently results in an increase in heat content in this region. The upper four panels of Figure 7 show vertical cross sections of the temperature field for four seasons along the 29°N line of latitude which passes through OWS-T. The bottom panel in Figure 7 shows the mean zonal gradient of the heat content of a water column of 200 m thickness in the four seasons. Negative values correspond to the eastward lowering of temperature. The data used in drawing these figures has been obtained from the Marine Environmental Atlas compiled by the Japan Oceanographic Data Center. It is found that the zonal gradient of temperature is largest in spring especially in the region east of the Kuroshio.

We have insufficient data to draw a vertical section along the longitudinal direction through OWS-T corresponding to Figure 7. However, it may be speculated that there is a temperature distribution which causes heat transport by lateral mixing from the Kuroshio located north of OWS-T, and from the Subtropical Counter Current located to its south.

The features of these water masses may further be brought to light by examining their typical time and space scales. Passages of these water masses may cause variation of SST as well as of the heat content in the surface layer. Fortunately, the time resolution is high enough to detect variation of SST on a time scale of a few days, since the interval of observation is three hours. Thus the dominant periods of the variation of SST in each month from June in 1950 to October in 1953 were determined from power spectra based on these data. For practical reasons the power spectrum for a particular month was calculated by using 512 consecutive data, corresponding to the data for two months, beginning on the 16th day of the preceding month; the power spectrum is thus influenced a little by variation in adjacent months.

Fig. 7. Temperature sections along the 29°N-line of latitude for four seasons (upper four figures), and the horizontal gradient of heat content in the surface layer (bottom figure). The negative values correspond to eastward flow of heat.
and $Q_E$ were investigated (instead of $Q$, since $Q$ was evaluated every day and not every three hours). It is found that both have peaks at periods longer than one day, with ranges almost the same as for SST. However, the peaks of $Q_E$ do not agree with those of the SST, nor with those of the wind stress, and this implies that variation of SST is affected more by three-dimensional processes.

In conclusion, it is inferred that these three-dimensional processes are caused by the advection of water masses, which typically take on average 2 to 3 days to pass the station. The typical space scale can be estimated by multiplying this time scale by the mean drift velocity of water masses. However, the drift velocity was not obtained during these observations. If a value of 10 cm s$^{-1}$ is assumed, which seems to be reasonable as the mean drift velocity in the area adjacent to OWS-T, the typical space scale is estimated to be about 20 km.

This is of comparable size to the water mass described by Toba et al. (1983) from the Japan Sea. Further, continuous recording of subsurface temperature by ferryboats running between Nagoya and Tomakomai, also shows relatively large temperature variations with space scales of about 10 to 20 km (Hanawa, 1983).

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References


Appendix

<Empirical formulae for heat exchange>

The empirical formulae for the estimation of each component of the heat exchange used in the present paper (Eq. 2) are as follows:

\[ Q_s = \begin{cases} (1 - r)(0.67 \times Ld^*), & (C \leq 0.3) \\ (1 - r)(0.84 - 0.58 \times C)Ld^*, & (C > 0.3) \end{cases} \]

(A-1)

\[ Q_s = \varepsilon (L_1 - \sigma \theta_a^4), \]

\[ L_1 = \sigma \theta_a^{-4} \left[ 1 - (0.49 - 0.66 \times e_a^{-1/2}) C^* \right], \]

\[ C^* = 1 - 0.53(1 - 0.0095 \times e_a) \times (C + 0.5 \times C^2) \]

(A-2)

\[ Q_E = \rho L C_E (q_s - q_a) w, \]

(A-3)

\[ Q_0 = \rho C_p C_H (T_s - T_a) w, \]

(A-4)

where \( Ld^* \), \( r \) and \( C \) in (A-1) are the daily total value of the solar radiation at the top of the atmosphere, the reflectivity of the sea surface, and the daily mean amount of cloud, respectively; \( L_1 \), \( \sigma \theta_a^{-4} \), \( \varepsilon \) and \( e_a \) in (A-2) are the downward atmospheric radiation, the upward radiation of the sea corresponding to the absolute sea surface temperature \( \theta_a \), the emissivity of the sea surface, and the daily mean vapor pressure, respectively; \( \rho, L \) and \( C_p \) are the air density, the latent heat of evaporation, and the specific heat of air at constant pressure, respectively; \( T_s \) and \( q_s \) are the temperature, and the specific humidity at the sea surface; \( T_a \), \( q_a \) and \( w \) are the air temperature, the specific humidity and the wind speed at a level 10 m above the sea surface; and \( C_E \) and \( C_H \) are the transfer coefficients under diabatic conditions.

Among the above equations, those for \( Q_s, Q_E \) and \( Q_0 \) are Kondo’s (1967 and 1975) empirical formulae. However, (A-1) for \( Q_s \) is a new equation, as mentioned in section 2. In order
To formulate this equation, we selected eight observational stations, as shown in Figure 9(a), which are located on the coast of Japan and islands which surround OWS-T, and have used daily data for the total solar radiation and cloud amount at these stations in 1978. We have assumed the simplest form for the relationship between $Q_s/I_d^*$ and $C$ as follows

$$Q_s/I_d^* = F(C), \quad (A-5)$$

and $F(C)$ has been determined as a function of $C/I_d^*$ is given by

$$I_d^* = \left(\frac{Q_0}{\pi}\right) \left(\frac{\bar{d}}{d}\right)^2 \left( H \sin \varphi \sin \delta + \cos \varphi \cos \delta \sin H \right), \quad (A-6)$$

where $Q_0$ is the solar constant, $\bar{d}$ and $d$ are the mean and the instantaneous distances between the sun and the earth, respectively, $\varphi$ is the latitude, $\delta$ the solar declination, $H$ is the half-day length and is given by

$$H = \cos^{-1}(-\tan \varphi \tan \delta). \quad (A-7)$$

Figure 9(b) shows the relationship between $Q_s/I_d^*$ and $C$. It is clear that this ratio is nearly constant for $C \leq 0.3$ with change of $C$, and it increases inversely and almost linearly with $C$ for $C > 0.3$. Thus, where $C \leq 0.3$ the total solar radiation does not depend on the amount of cloud. This is because, as explained by Kondo (1967), the downward reflection of solar radiation by cloud compensates the loss due to cloud cover, when the cloud cover is slight. We have approximated the relationship for $C > 0.3$ with a straight line determined by the method of least squares. The error in values calculated from the new empirical formula is at most 10%. This error is sufficiently small for the purposes of discussion of heat exchange in the present paper.
海洋気象観測点 T における海洋表層の熱収支

倉沢由和*, 花輪公雄*, 鳥羽正明*

要旨：日本南方海域の上層 200 m の熱収支を、集中的に 1950 年 6 月から 1953 年 11 月まで海洋気象観測点 T（北緯 29 度、東経 135 度）で得られた海上気象資料・海洋資料を用いて調べた。表層の熱収支の局所的時間変化と、大気海洋境界を通しての正味の熱フラックスをこれらの資料から評価し、海洋内の熱収支を上記の 2 つの量の残差として求めた。面での熱交換と海洋内の熱収支の相対的な重要性に関しては、数日から 1 月の時間スケールでは、合熱流の時間変化は表に現在の熱収支に依存していること。季節変化の時間スケールでは、12 月から 2 月までは海面での熱交換が主に決まる。3 月から 5 月までは海洋内部の熱収支。7 月から 11 月までは双方の過程で決まることがわかった。短周期の海洋の熱収支は、フロントを持つ水塊の移流によって生じていると推察される。海洋表面水温のスペクトル解析から、水塊が観測点を通過する典型的な時間スケールは 2～3 日であり、その空間スケールは 10 cm s⁻¹ の移流速度を仮定すると 20 km 程度と見積られる。

* 東北大学理学部地球物理学教室
〒980 仙台市荒巻字育栄
† 現所属：石油公団 〒100 東京都千代田区内幸町 2-2-2