The Journal of The Oceanological Society of Korea
Vol. 10, No. 2, p. 57-66, December 1975

CHARACTERISTICS OF STABILITY AND INTENSITY OF VERTICAL TRANSFER IN THE WESTERN CHANNEL OF THE KOREA STRAIT

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ABSTRACT

Structure of thermocline, characteristics of stability and intensity of vertical transfer have been studied with hourly oceanographic data in each layers on Line 207 from 1968 to 1969.

It is found that a typical thermocline is formed at depths of 10 to 50 meters in summer and early autumn and its core is located near depths of 25 meters.

The maximum diffusion coefficient of vertical turbulent is found to be 140cm^2/sec at the surface layer(i.e., 0—10 meters), while the minimum is 5cm^2/sec at depths of 25 meters, consistent with characteristics of stability and structure of thermocline layers.

Our computed diffusion coefficient and stability indicate that the mixing hardly takes place below depths of 80 meters during summer and early autumn, but for the rest of the season mixing could move up to the depth of 50 maters. It appears that the Western Channel of the Korea Strait consist of three different water masses during summer and autumn, and for the rest of the season, two kinds are present.

INTRODUCTION

Problems of stability and vertical turbulent diffusion in water masses have been considered critical to determining the structure of stratification and mixing of water masses.

The investigation on the stability of water mass was initiated by Hesselberg and Sverdrup (1915), followed by Defant (1929), Schubert (1935), Neumann(1948), Kolesnikov(1962) and many others by applying it various seas. Recently an intensive study has been made on the effect of stability on the intensity of vertical transfer by Malcus(1959), Proudman(1959), Kolesnikov(1962), et al.

The Western Channel of the Korea Strait is one of the widely surveyed area in the world since 1926. Unfortunately this survey has been done only on the qualitative basis.

The present study is to examine the characteristics of thermocline, stability and intensity of vertical transfer in the Western Channel of the Korea Strait in order to quantitatively explore the general orientation of stratification, formation of unstable layers, water masses and vertical mixing.

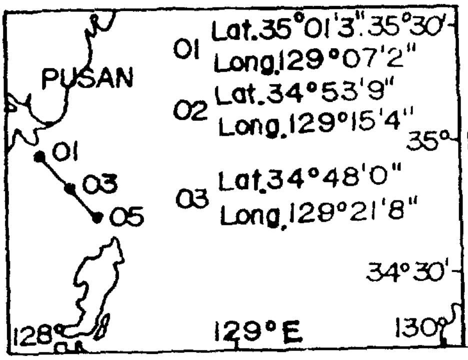


Fig. 1. Location of the studied area.

THEORETICAL FORMULATION

The stability, E of a water body is defined as the relative acceleration that a water particle experiences when displaced a unit length along a vertical distance from its rest position. Thus, it becomes

$$E=\lim_{\Delta z\to 0}\frac{dw/dt}{g}\cdot\frac{1}{\Delta z}=\frac{1}{\rho}\frac{\delta\rho}{dz} \qquad 1.$$

Since the density is functions of temperature, t, salinity, s and pressure, p., Equation 1 can be rewritten as

$$E=\frac{1}{\rho}\left[\frac{\partial\rho}{\partial s}\frac{ds}{dz}+\frac{\partial\rho}{\partial t}\left(\frac{dt}{dz}-\frac{d\theta}{dz}\right)\right] \qquad 2.$$

In the case of shallow water, the adiabatic temperature gradient, $(d\theta/dz)$ is negligibly small. Equation 2 is simplified as

$$E=\left(\frac{\partial\rho}{\partial s}\frac{ds}{dz}+\frac{\partial\rho}{\partial t}\frac{dt}{dz}\right)=\frac{d\rho}{dz} \qquad 3.$$

Replacing the density, ρ by σ_t,

$\sigma_t=(\rho-1)\ 1000$

Equation 3 becomes

$$E=10^{-3}d\sigma_t/dz \qquad 4.$$

It is found that Equation 4 represents fairly well the stability of water masses down to a depth of 1,400 meters.

The vertical distribution of physical properties in the ocean, depend heavily on the turbulent diffusion coefficient along with the vertical current velocity.

In the presence of turbulent diffusion, the equation of continuity is given by

$$\frac{\partial s}{\partial t}+U\frac{\partial s}{\partial x}+V\frac{\partial s}{\partial y}+W\frac{\partial s}{\partial z}=\frac{\partial}{\partial x}\left(Kx\frac{\partial s}{\partial x}\right)+\frac{\partial}{\partial y}\left(Ky\frac{\partial s}{\partial y}\right)+\frac{\partial}{\partial z}\left(Kz\frac{\partial s}{\partial z}\right) \qquad 5.$$

where S denotes the properties in the ocean, and U, V, W and Kx, Ky, Kz represent the components of velocity and the coefficients of turbulent diffusion in the cartesian coordinate, respectively.

When the mean current U is taken parallel to the X-axis and the mixing is assumed to take place only along the vertical direction, the diffusion coefficient of the stationary current is given by

$$Kz=\frac{1}{\partial s/\partial z}\int_{z_0}^{z}U\frac{\partial s}{\partial x}dz \qquad 6.$$

MATERIALS AND METHODS

The present study is based on the data of hourly oceanographic observations in each layers on Line 207 from 1968 to 1969 made by the Fisheries Research and Development Agency in Korea. The currents data of Lee(1970) have also been employed for the computation of vertical turbulent diffusion coefficients, noting that his data were obtained in the same period as the hourly oceanographic obserations of F.R.D.A. were made.

In the present investigation, the vertical distribution of stability has been obtained by means of Equation 4. The physical cause of instability has also been examined by means of the temperature and salinity difference diagram of each unstable layers.

The criterion of defining thermocline layer has been adopted in accordance with that of Mazeika(1962), namely the temperature gradient should be greater than 0.37°C/10 meters. The effect of the thermocline on the stability

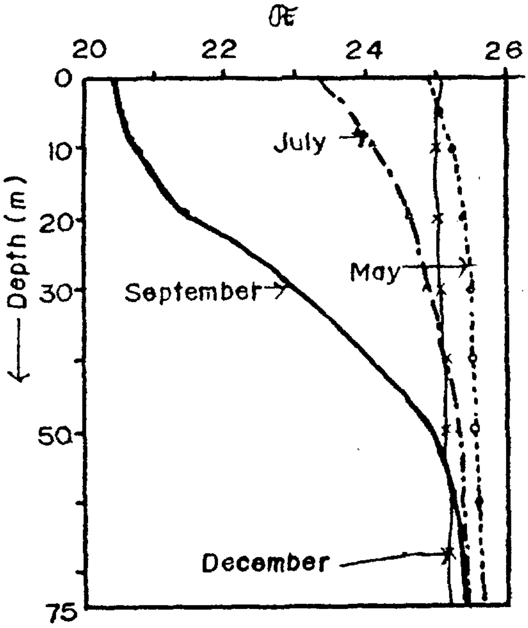


Fig. 2. Monthly variations of average σ_t.

has been studied by the vertical distributions of stability and temperature.

For the study of intensity of vertical transfer, 03 Station was selected because it is located at the central part of Western Channel, thus its data being less influenced by friction. With the aid of Equation 6, the intensity of vertical transfer has been computed by using the oceano-

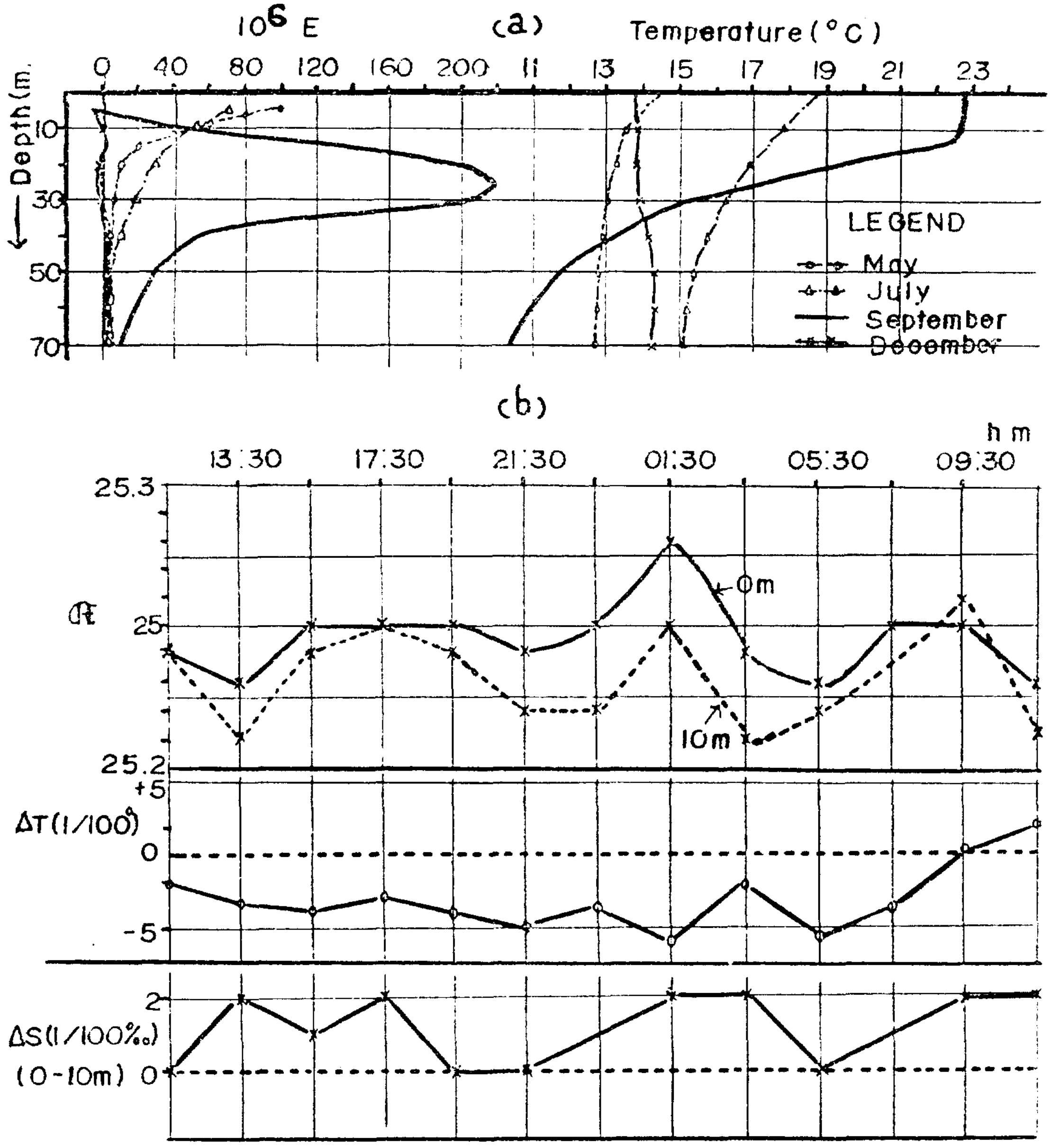


Fig. 3. (a) Left; Vertical stability, at 207-01 Station. shows maximum stability in September at the depth of 25 meters. Right; Vertical distributions of temperature at the same st.

The unstable layers formed in the layer between 0 to 10 meters in September and December.

(b) Upper part; Hourly values of σ_t at the surface (0 meter) and at 10 meters depth. Lower part; Temperature and Salinity differences T (0-10m) and S (0-10m), respectively, between 0 and 10 meters depth. December, 1968.

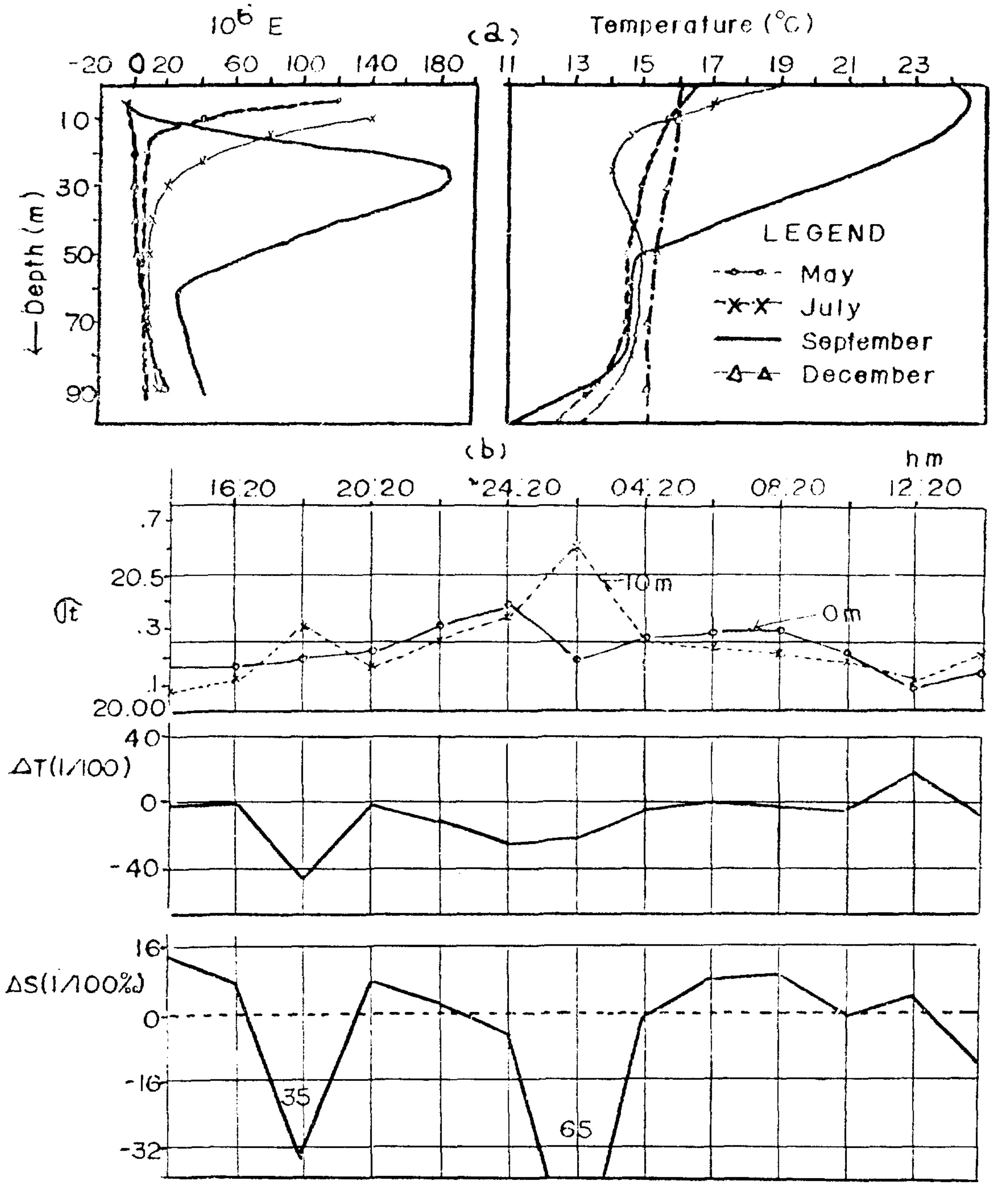
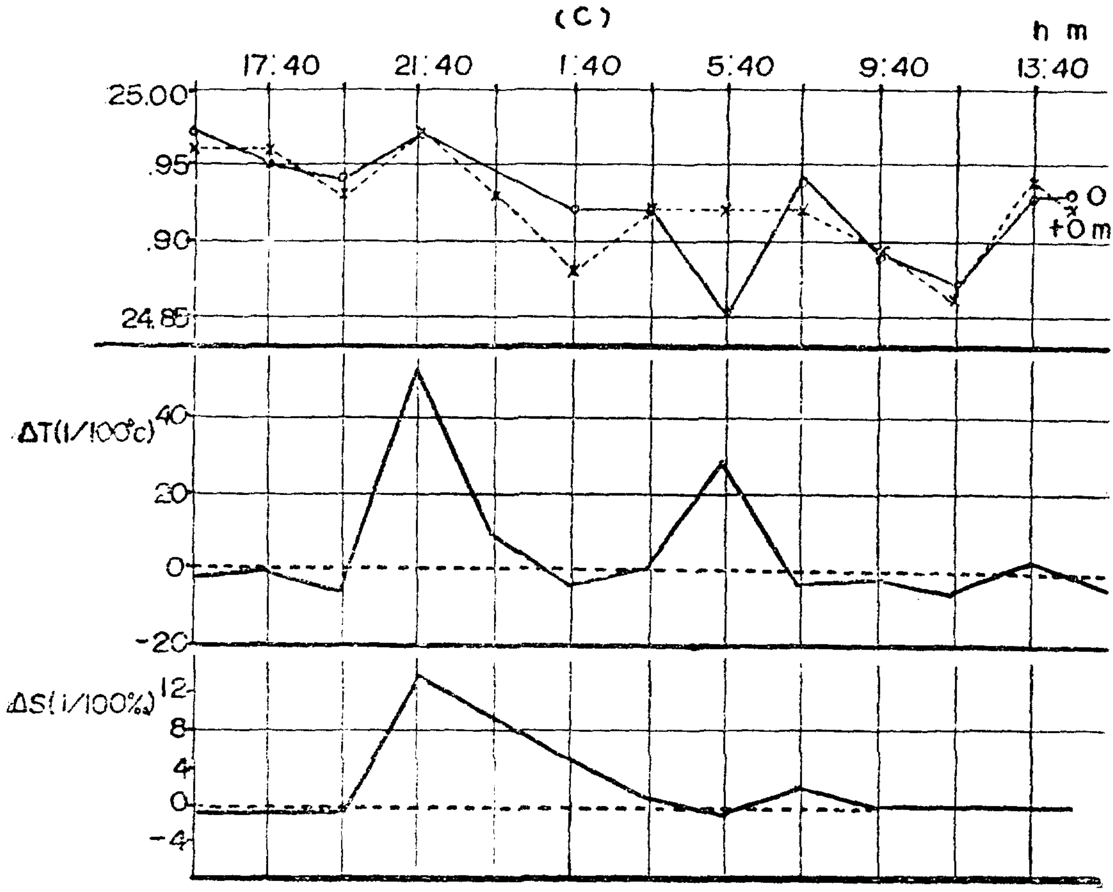


Fig. 4. (a) Left; Vertical stability, at 03 Station. Right; Vertical distributions of temperature at the same station.
(b) Upper part; Hourly values of σ_t at the surface(0 meter) and at 10 meters depth. Lower part; Temperature and Salinity differences, T(0-10m) and S(0-10m), respectively, between 0 and 10 meters depth. September, 1969.



(c) Upper part; Hourly values of σ_t at the surface and at 10 meters depth. Lower part; Temperature and Salinity differences, T(0-10 m) and S(0-10m), respectively, between 0 and 10 meters depth. December, 1968.

graphic data of F.R.D.A along with currents data of Lee(1970).

It should be, however, noted that like many others(e.g., Fjeldstad(1933), Defant(1936), Stockman(1946), Proudman(1957), Kolesnikov (1962),) the vertical velocity component of currents has been neglected in our computation.

RESULTS AND DISCUSSIONS

Thermocline Structure: Typical thermocline is formed at depths of 10 to 50 meters in September and 0 to 50 meters in July at all stations. Among those thermocline, the strongest thermocline is found at the 01 Station in September (See Figure 3a). Weak thermocline is found in May at all stations. Interestingly, the thermocline is not found in December as indicated in Figures 3a, 4a and 5a. These phenomena are consitent with the monthly variations of average σ_t as shown in Figure 2. Below the depth of 50 meters, the average value of σ_t merges together within a very narrow range of σ_t.

The monthly variation of density and temperature are shown in Figures 2, 3a, 4a and 5a. From these figures one can easily find that there is a very close relationship between them. When the surface density decreases, there is a tendency that the thermocline becomes stronger, as evidenced, for example in September where

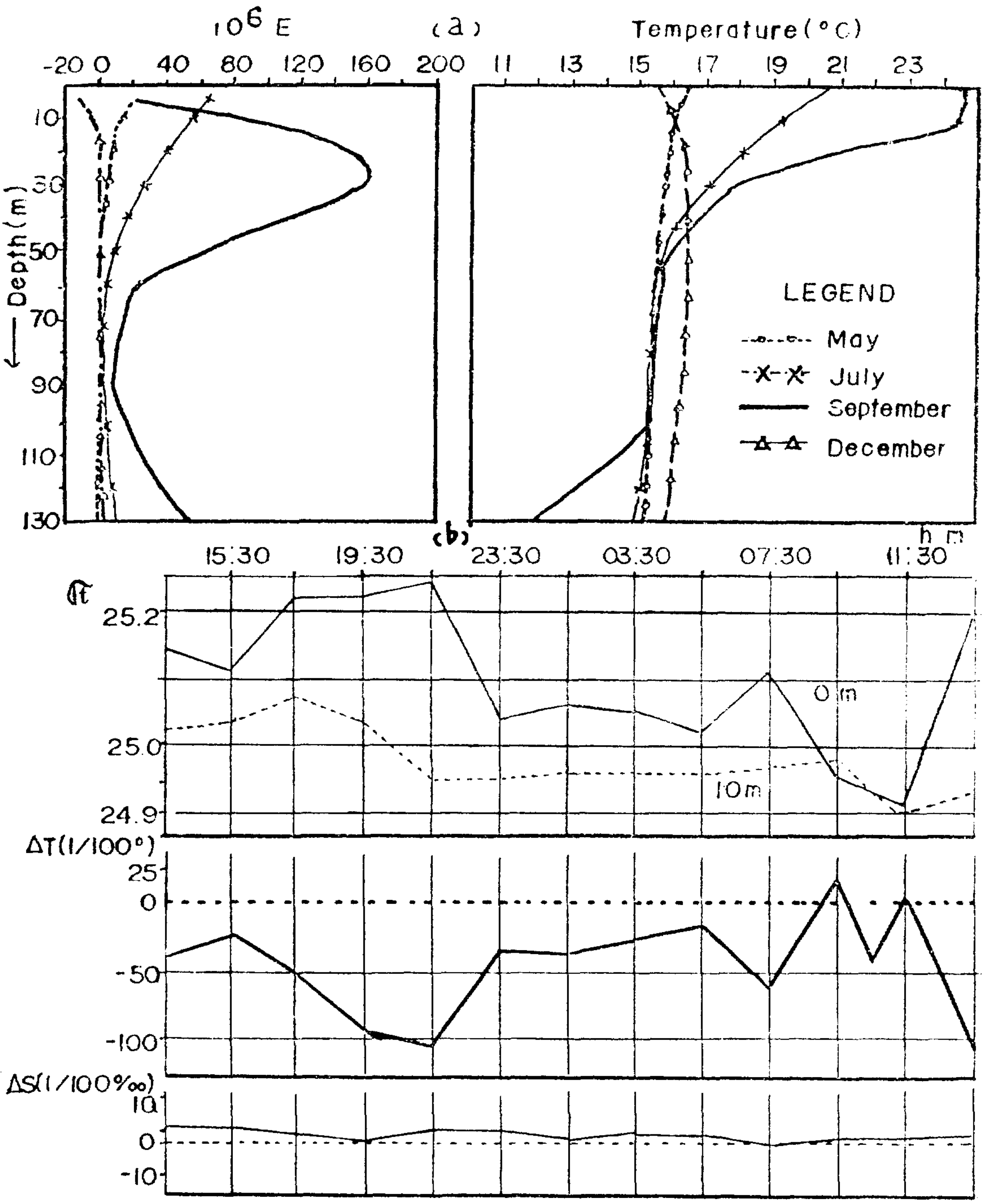


Fig. 5. (a) Left; Vertical stability, at 05 Station. Right; Vertical distributions of temperature at the same station.
(b) Upper part; Hourly values of σ_t at the surface(0 meter) and at 10 meters depth. Lower part; Temperature and Salinity differences, T(0-10m) and S(0-10m), respectively, between 0 and 10 meters depth. December, 1968.

when a very strong thermocline is formed, the density reveals minimum.

A constant temperature layer is found between depths of 50 to 80 meters for 03 Station and 50 to 90 meters for 05 Station. Below these depths, the temperature is fluctuates considerably. This means that another water mass is intruded. At this level, the temperature rapidly decreases from July to September and then it recovers to a steady state.

In winter, the density is almost constant, slightly dense water is covered in the upper layers of 0 to 10 meters. At this layer, a temperature inverse occurs.

The hourly variations of unstable layers are shown in Figures 3b, 4b, 4c and 5b, respectively. The upper part of each of these figures represents the hourly variation of σ_t at the surface and and at 10 meters, which turns out to be useful for stability determination. The lower part of these figures shows the temperature and salinity differences between 0 and 10 meters. These figures can be used to determine the important physical factors which influence the formation of unstable layers. According to our analysis, the unstable layers found in 03 and 05 Stations appears to be due to variations of temperature and salinity, while 01 station is primarily due to temperature change alone.

Stability: The computed values of stability are presented in Figures 3a, 4a and 5a, respectively. The maximum stability is found to reside in the layers between 25 to 30 meters with $E=160\times10^{-6}\sim220\times10^{-6}$ (i,e., $16,000\times10^{-8}\sim22,000\times10^{-8}$ in conventional unit). Negative stability is located at the surface layer of 0 to 10 meters in December at all stations and 03 Station in September.

The characteristics of monthly variations of stability are as follows; from May to July, the stability takes its maximum at the surface(0-10m) (while its minimum occurs in September and December). During this period, the stability decreases gradually downward to the depth of 50 meters and then reaches to a quasi-neutral state. In September, however, the stability increases rapidly from the surface(where the value is minimum) downward until it reaches the maximum value at the depth of 50 meters. Eventually, it reaches to the quasi-neutral state too. During the period from July to September, stability increases from the lower depth of the quasi-neutral layer to the bottom of the channel. This trend is more typical in September than July.

According to the results coming from the characteristics of vertical temperature distribution and stability (Fig. 3a, 4a and 5a), the thickness of the constant temperature layer (i,e., the quasi-neutral stability layer) varies from station to station. At the Station 03, the constant temperature layer appears in depths between 50 and 80 meters but at the Station 05, it lies from 50 to 90 meters.

Finally, we wish to remark that all the figures on our stability analysis are consistent with the thermocline structure: namely, (1) the negative stability occurs in the temperature reversing layer; (2) the maximum stability appears in the core of thermocline; (3) the quasi-neutral stability is related with the constant temperature layer; (4) the slight increase of stability found in the lower transition depth appears to be associated with the cold water intrusion.

Intensity of Vertical Transfer: The computed vertical turbulent diffusion coefficient is presented in Figure 6. In this computation, the currents data of September at 03 Station has been employed. Because it is solely available.

Figure 6. shows that the stability tends to be inversely proportional to diffusion coefficient just as found in Atlantic Ocean by Kolesnikov

(1962). The maximum diffusion coefficient is found in the surface mixing layer, which is

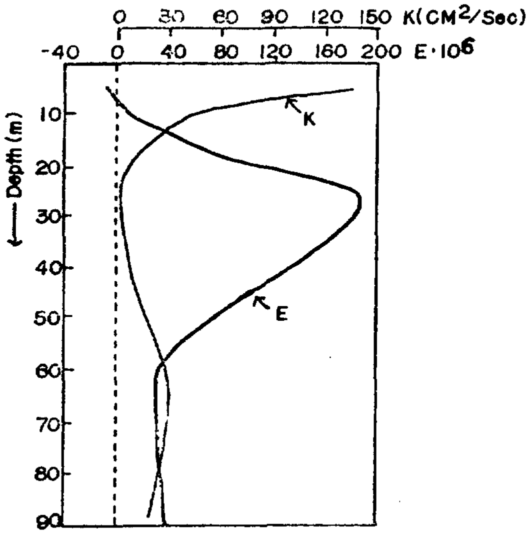


Fig. 6. The vertical distribution of diffusion coefficient of vertical turbulent (K) and of stability(E).

approximately $140cm^2/sec$. In the thermocline layer, the computed coefficient varies from 5 to $10cm^2/sec$. At the very core of thermocline, it becomes 5 cm^2/sec, which is the minimum in this channel. In the layer of constant temperature, the coefficient ranges from 15 to $30cm^2/sec$, with the average of $22cm^2/sec$, amounting to one sixths of its maximum value.

It is interesting to note that according to Kolesnikov(1962), no mixing takes place when the diffusion coefficient becomes either less than $30cm^2/sec$ or less than one fifths of the maximum value of the surface mixing layer.

On the basis of the characteristics of our computed diffusion coefficient and stability, the waters of the station 03 consist of three layers; namely, the surface layer (i.e., the mixing and thermocline layer)ranging from 0 to 50 meters, the middle layer roughly from 50 to 80 maters and the cold bottom layer appearing only in summer and early autumn. The formation of the mixing layer in the surface(0-10m) appears to be caused by insolation, rainfall and fresh water inflows from the continent.

Figure 6 can be used to estimate the diffusion coefficient for other months. Making use of the fact that the stability is inversely proportional to diffusion coefficient, it is expected that the vertical mixing could move up to 50 meters in Winter and Spring, but for summer and early autumn, the mixing is restricted only below 80 meters. This phenomenon appears to be associated with the appearance of cold bottom water mass and a rapid decrease of surface density.

Water Masses: The monthly variations for density, vertical distribution of temperature, stability and diffusion coefficient(Fig. 2, 3a, 4a, 5a and 6.) indicate that the main water masses of this channel can be distinguished in three different groups.

1) The Surface Water; It is characterized by high temperature and low salinity. It appears in summer and disappears in winter. Its thickness is about 50 meters.

2) The Middle Water(Constant Temperature Layer); It is characterized by high salinity and nearly constant temperature. This water is located between 50—80 meters for the station 03 and 50—90 meters for the Station 05. Its physical properties remain nearly the same throught the year, except for the summer and early autumn when the cold water is intruded into the layears below the depth of 50 meters.

3) The Cold Water; It is characterized by low temperature and low salinity. This cold water shows up strongly whenever the Tsushima Current becomes strong in August and September(Miyazaki, 1952., Yi, 1966). At this period the temperature of this channel becomes minimum at the bottom.

SUMMARY

As discussed in the previous section, the ch-

aracteristics of stability and intensity of vertical transfer are all in consitent with the fluctuations of density, thermocline structure and stratification of water masses.

It is found that a typical thermocline is formed at the depth of 0 to 50 meters in summer and 10 to 50 meters in early autumn. Its core is located near depths of 25 meters. Unstable layers are found at the surface(0-10 m) from September to December, and the cause appears to be due to the variations of temperature and salinty for the Station 03 and 05. However, for the Station 01, its instability is due to the temperature change alone.

The maximum diffusion coefficient is found to be 140cm^2/sec at the surface(0-10 m), while its minimum takes 5cm^2/sec at the depth of 25 meters, consistent with the characteristics of stability and thermocline struture. The Western Channel appears to be made of three kinds of water masses during summer and autumn, but for the rest of the season, only two kinds of water masses are found.

It may be summed up that the mixing hardly takes place between the depth of roughly 80 meters during summer and early autumn, but for the rest, the mixing could move up to the depth of 50 meters.

ACKNOWLEDGEMENTS

The auther wishes to express his gratitude to Park, Young Hyang for his generous assistance for the preparation of this paper and to Yi, Suk-U, Yun, Hong Sik, Han, Sangbok, for their helpful suggestions and constructive criticism. Also thank to colleague Kim, Wan Soo, Park, Yong Ahn, for their encouragement.

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