The Journal of the Oceanological Society of Korea
Vol. 18, No. 1, p. 43-48, June 1983

ON THE GENERATION OF TEMPERATURE INVERSIONS IN THE UPPER LAYER OF THE OCEAN

Yong Q. Kang

Department of Oceanography, National Fisheries University of Busan, Busan 608, Korea

ABSTRACT

Oceanic temperature inversions, with unstable stratifications, are frequently found in the surface layer of a few tens meters in the Japan Sea and the Yellow Sea in winter. Mechanisms responsible for the generation of temperature inversions include the followings: (1) The net heat loss at the sea surface requires an upward transport of heat from the interior of the ocean by convection, and this convection leads to the temperature inversions. (2) The downward propagation of the annual variation of the sea surface temperature, with an exponential decrease of amplitude and a linear change of phase with depth, generates the surface inversion layer in winter. (3) The cold water advection by Ekman drift, of which magnitude decreases exponentially with depth, generates temperature inversions in the surface layer. I present simple analytic models of the temperature inversions for the three possible mechanisms mentioned above.

1. INTRODUCTION

The ocean is usually stably or neutrally stratified. That is, the density of seawater usually does not decrease with depth. Vertical distribution of temperature in the ocean is characterized with the surface mixed layer of a few tens meters above the thermocline layer. When the temperature increases with depth, we call such phenomena as oceanic temperature inversions. Note that the oceanic temperature inversion has an opposite meaning to the atmospheric inversion in which the temperature increases with height.

Sometimes, however, temperature inversion layers are found in the ocean (Federov, 1978; Kim and Cho, 1982; Nagata, 1967a, 1967b, 1968; Nakao, 1977). The oceanic temperature inversion does not necessarily accompany a density inversion because the distribution of salinity can compensate unstable straifications associated with temperature inversons. Kim and Cho (1982), however, showed that distributions of both temperature and salinity are responsible for an unstable stratification of density inversion in the surface layer in the Japan Sea in winter. They showed that more than a half of the Japan Sea has density inversions in winter.

Fig. 1 shows the difference between the sea surface temperature (SST) and the temperature at 50m at the Ocean Data Buoy Station No. 6 (37°45′N, 134°23′E) in the Japan Sea for one year (November 1979 to October 1980). This figure is made by using the data published by the Japan Meteorological Agency (1981). This figure shows that the sea surface temperature is usually cooler than the temperature at 50m by 0.2~0.5°C from November to March. From this figure, and also from the results by Kim and Cho (1982), we learn that temperature inversions are rather persistent and wide-spread phenomena in the Japan Sea in winter.

In this paper I present simple analytic models for generation of temperature inversions. Mechanisms of temperature inversions discussed in this paper are; (1) vertical convection to com-

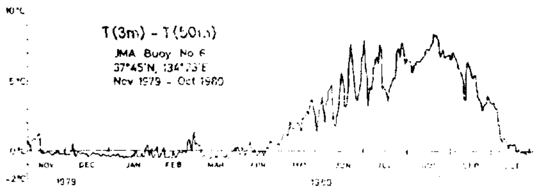

Fig. 1. Difference between the temperature at the sea surface (3m) and that at 50m at JMA Buoy Station No. 6 (37°45′N, 134°23′E) in the Japan Sea from November 1979 to October 1980.

pensate the net heat loss at the surface, (2) vertical phase change of the downward propagating annual temperature variation, and (3) cold water advection by Ekman drift in the surface layer. The theoretical models for temperature inversions should be able to account for the following observed features: (1) Inversion layers are usually found in the surface layer of a few tens meters. (2) The inversion layers are found usually in winter. (3) Density inversions are sometimes apparently persistent even though the vertical stratification is unstable and consequently causes convection. (4) Temperature inversions are more pronounced in oceanic frontal areas where horizontal temperature gradients are large.

2. NET HEAT LOSS AT THE SEA SURFACE

In winter there is a net heat loss at the sea surface. For example, the net heat loss in the Japan Sea in winter is about 500 cal cm^{-2} day^{-1} or 5×10^{-3} cal cm^{-2} sec^{-1} (Wyrtki, 1966; Maizuru Marine Observatory, 1972). If we assume that the net heat loss at the sea surface is compensated by vertical convection of heat from the interior of the ocean, then a steady state heat balance will be

$$\rho_0 c_p \overline{w'T'} = -\rho_0 c_p \left(K \frac{\partial \overline{T}}{\partial z}\right) = \dot{Q}, \qquad (2.1)$$

where ρ_0 is the density of sea water, c_p is the specific heat, $\overline{w'T'}$ is the average correlation between the fluctuations of vertical velocity and those of temperature, K is the eddy conductivity, $\overline{T}(z)$ is the mean temperature, z is the vertical coordinate (positive upward), and $\dot{Q}$ is the net heat flux per unit area per unit time. In (2.1) I neglected the molecular diffusion of heat because it is much smaller than the heat transported by convection and the thickness of inversion layer associated with the molecular diffusion of heat is less than 1 cm order (Stern, 1975, p. 186).

If the sea surface is regarded as a rigid lid, then, by Prnadtl's mixing length hypothesis, one may assume that the eddy conductivity is proportional to depth, i.e., $K=-az$, where a is a positive constant (cf. Madsen, 1977). The sea surface, however, cannot be regarded as a rigid lid due to the presence of wave-induced turbulences at the sea surface. Hence, a more realistic parameterization of the eddy conductivity with an allowance of wave-induced turbulences should be

$$K = K_0 - az, \qquad (2.2)$$

where K_0 is the eddy conductivity associated with the surface wave-induced turbulences. A similar parameterization for the case of momentum stress can be found in Kraus (1972, p. 137).

Using (2.2) we can write the heat balance equation (2.1) as

$$\frac{\partial \overline{T}(z)}{\partial z} = -\frac{1}{K_0 - az} \frac{\dot{Q}}{\rho_0 c_p}. \qquad (2.3)$$

The solution of this equation is

$$\overline{T}(z) = \frac{\dot{Q}}{a \rho_0 c_p} \ln\left(1 - \frac{a}{K_0} z\right) + \text{const.} \qquad (2.4)$$

For numeral values of $\dot{Q}=5\times10^{-3}$ cal cm^{-2} sec^{-1}, $\rho_0=1$ gm cm^{-3}, $c_p=1$ cal gm^{-1} °C^{-1}, $a=5\times10^{-2}$cm sec^{-1}, and $K_0=50$ cm^2 sec^{-1}, eqn (2.4) becomes

$$\overline{T}(z) = 0.1 \ln(1-0.1z) + T_0 \ (°\text{C}) \qquad (2.5)$$

where T_0 is the sea suface temperature and z is in meters. Eqns. (2.4) and (2.5) are plotted

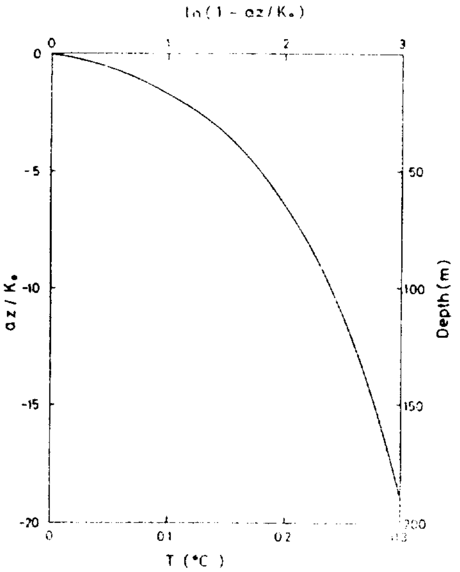

Fig. 2. Vertical distribution of temperature associated with net heat loss at the sea surface. The plot represents $\bar{T}=\ln(1-az/K_o)$ and $\bar{T}(z)=0.1\ \ln(1-0.1z)$.

in Fig. 2. This figure shows that for the case of a net heat loss at the sea surface the mean temperature increases with depth and, therefore, temperature inversions are generated in the surface layer. From (2.5) we learn that the temperature at 50m is about 0.2°C higher than that at the sea surface.

3. DOWNWARD PROPAGATION OF ANNUAL TEMPERATURE VARIATIONS

The sea surface temperature (SST) varies predominantly with an annual cycle. If the advection of heat by ocean current is neglected, the fluctuations of temperature, associated with the annual SST variations, in the interior of the ocean are described by

$$\frac{\partial T(z,t)}{\partial t}=\frac{\partial}{\partial z}\left(K\frac{\partial T}{\partial z}\right), \qquad (3.1)$$

subject to the surface boundary condition

$$T(0,t)=T_0+T_1\cos(\omega t+\delta) \text{ at } z=0, \quad (3.2)$$

where K is the eddy conductivity, T_0 is the annual mean temperature at the sea surface, T_1 is the amplitude of the annual temperature variation at the sea surface, and $\omega(=2\pi$ rad/year) and δ are the angular frequency and the phase of the annual temperature variations at the sea surface, respectively. If K is constant and $|\partial^2\bar{T}/\partial z^2|\ll T_1/K$, where $\bar{T}(z)$ is the time-averaged mean temperature at z, the solution of (3.1) and (3.2) is

$$T(z,t)=\bar{T}(z)+T_1\exp\left(\sqrt{\frac{\omega}{2K}}\,z\right)\cos\left(\omega t+\sqrt{\frac{\omega}{2K}}\,z+\delta\right). \qquad (3.3)$$

This solution shows that the annual SST variations propagate downward. The amplitude of fluctuations decreases exponentially with depth and the phase varies linearly with depth. Because of the downward propagation of the annual temperature variation, the temperature in the interior of the ocean can, for a certain period of time, be higher than that at the sea surface.

As an illustrative example, let the annual temperature propagating downward has an e-folding depth of 100m. The corresponding eddy conductivity is 10 cm² sec⁻¹. If the amplitude of the annual variation at the sea surface is 5°C, then the amplitude at 30m would be 3.7°C. The sea water temperature at 30m has a phase shift of 17° with respect to that at the sea surface. This means that the maximum temperature at 30m occurs 17 days after an occurance of the maximum SST. Fig. 3 shows the temperature variations at the sea surface and at 30m. In this figure it is assumed that the mean temperatures at the sea surface and at 30m are respectively 15°C and 14°C, and the maximum SST occurs on the last day of August. This illustrative example (Fig. 3) shows that the temperature at 30m is higher than that at the sea surface during winter (December to March), and this period of temperature inversion agrees reasonably with the

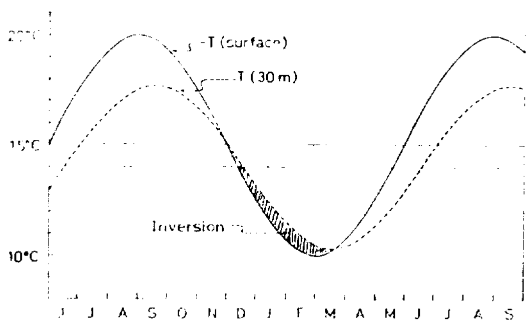

Fig. 3. The annual temperature variations at the sea surface and at 30m. Hatched area represents temperature inversions.

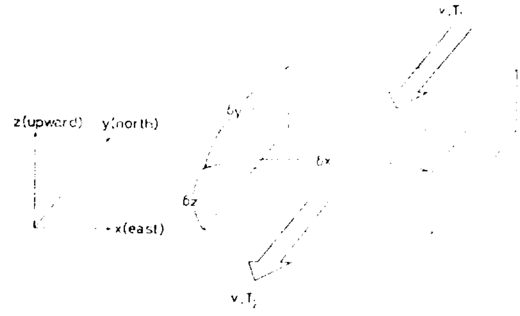

Fig. 4. A box with dimensions of $\delta x, \delta y$ and δz. The temperature at the northern and southern walls are respectively T_1 and T_2, and the southward flows across both walls are the same.

observed inversion period (see Fig. 1).

4. COLD WATER ADVECTION

When the surface Ekman current flows from a colder area to a warmer region, there is a net negative supply of heat to the oceanic area of interest fixed in space. In the Japan Sea, for example, the isotherms at the sea surface are almost parallel to latitudes and the northwesterly monsoon generates southward Ekman current in the surface layer in winter. If the spatial distribution of temperature in the surface layer is maintained steady despite cold water advection, the heat loss by cold water advection should be compensated by the vertical eddy conduction of heat.

Let's consider a steady state heat balance in a small box with dimensions of δx, δy and δz fixed in space as shown in Fig. 4. We assume, for a mathematical simplicity, that the steady state mean temperature $\bar{T}$ is independent of longitudes and decreases linearly with latitude, i.e.,

$$\frac{\partial \bar{T}}{\partial x}=0, \quad \frac{\partial \bar{T}}{\partial y}=M=\text{const}, \ (M<0). \quad (4.1)$$

We consider southward flowing Ekman current of which magnitude is horizontally uniform but decreases exponentially with depth, as shown in Fig. 5. That is,

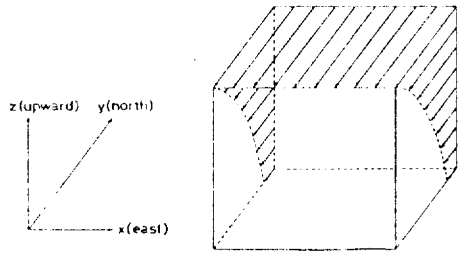

Fig. 5. Southward Ekman current. The magnitude of current is horizontally uniform, but decreases exponentially with depth.

$$\bar{V}(z)=-V_0 e^{\gamma z}, \ (V_0>0) \quad (4.2)$$

where

$$V_0=\frac{|\tau|}{\sqrt{A\rho_o f}}, \ \gamma=\sqrt{\frac{\rho_0 f}{2A}}, \quad (4.3)$$

V_o is the Ekman current at the sea surface, τ is the wind stress, A is Austausch coefficient, and f is the Coriolis parameter. The temperature at the northern and southern walls of Fig. 4 are respectively $\bar{T}_1$ and $\bar{T}_2$ ($\bar{T}_2>\bar{T}_1$). The inflow of heat through northern wall is $\rho_o c_p \bar{T}_1 \bar{V}\delta x\delta z$, and the outflow of heat through the southern wall is $\rho_o c_o \bar{T}_2 \bar{V}\delta x\delta z$. Since the sea water temperate at the southern wall is higher than that at the northern wall and the magnitude of horizontally uniform southward Ekman flow is the same at both walls, the heat leaving the box through the southern wall is larger than that entering through the northern wall. The net heat loss inside of the box by mean flow is

$$\rho_o c_p(\bar{T}_2-\bar{T}_1)\bar{V}\delta x\delta z. \tag{4.4}$$

In order to maintain a steady thermal state in the box, the loss of heat must be compensated by vertical convection, which may be parameterized as

$$\rho_o c_p\overline{w'T'}\delta x\delta y=-\rho_o c_p K\frac{\partial\bar{T}}{\partial z}\delta x\delta y. \tag{4.5}$$

Since the sum of heat loss by advection and heat gain by convection should vanish for a steady state, from (4.4) and (4.5) we obtain a heat balance equation

$$\frac{\partial\bar{T}}{\partial y}\bar{V}(z)+K\frac{\partial^2\bar{T}}{\partial z^2}=0, \tag{4.6}$$

or by using (4.1), (4.2) and (4.3),

$$-MV_o e^{\gamma z}+K\frac{\partial^2\bar{T}}{\partial z^2}=0. \tag{4.7}$$

The solution of (4.7) with a bounday condition $\bar{T}(-\infty)=T_d$, where T_d is the mean temperature at an infinite depth, is

$$\bar{T}(z)=T_d+\frac{MV_o}{K\gamma^2}e^{\gamma z}, \tag{4.8}$$

or

$$\bar{T}(z)-T_d=2\frac{\partial\bar{T}}{\partial y}\frac{\tau}{K}\sqrt{\frac{A}{\rho_o f}}\frac{1}{\rho_o f}\times\exp\left(\sqrt{\frac{\rho_o f}{2A}}z\right). \tag{4.9}$$

Notice that the magnitude of temperature inversion, $\bar{T}(0)-T_d$, is proportional to the meridional gradient of mean temperature and

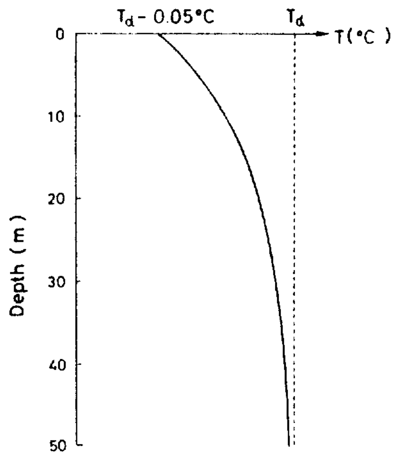

Fig. 6. Vertical distribution of temperature associated with cold water advection.

wind stress. With typical values of $A=10^2$ gm cm^{-1} sec^{-1}, $K=10^2$ cm^2 sec^{-1}, $\tau=1$ dyne cm^{-2}, $f(35°\text{N})=8.4\times10^{-5}$ sec^{-1}, and $\partial\bar{T}/\partial y=2\times10^{-7}$ °C cm^{-1}(10°C change over 5 degrees of latitude), (4.9) becomes

$$\bar{T}(z)=T_d-0.05\exp(z/15m). \tag{4.10}$$

Fig. 6 shows the vertical distribution of $\bar{T}(z)$ of (4.10) in the surface layer of the ocean.

5. DISCUSSION AND CONCLUSIONS

In this paper I presented simple analytic models that can account for the observed temperature inversion phenomena in the upper layer of the ocean in winter. I showed that the net heat loss at the sea surface, the downward propagation of annual temperature variation, and the cold water advection can generate temperature inversions. Although each mechanism was modelled separately for a mathematical simplicity, it should be reminded that all of the above mentioned mechanisms work simultaneously in actual situations.

The models in this paper show that the temperature inversions in the upper layer of the ocean are generated by the external thermal and/or momentum forcings applied at the sea surface and the corresponding adjustments in the interior of the ocean. The adjustments in the interior of the ocean depend on the magnitude of eddy conductivity. Eddy conductivity of 10 to 10^2 cm^2 sec^{-1} yields an inversion of an order of 10^{-1} °C. The corresponding surface inversion layers are shown to be generated in the upper tens of meters.

A favourable condition for a generation of temperature inversion layer is found in winter. The net heat loss at the sea surface and the cold water advection by monsoon winds in the Japan Sea occur in winter. Besides, the downard propagation of seasonal temperature variations can generate temperature inversions only in winter (see Fig. 3). Hence we expect that the temperature inversions in the surface layer of the ocean occur mainly in winter, and this expectation is confirmed by observations by Kim and Cho(1982).

ACKNOWLEDGEMENTS

I wish to thank Drs. H. J. Kim and K. D. Cho, National Fisheries University of Busan, Dr. I. J. Choi, Chungnam National University, and Dr. H. J. Lee, Korea Ocean Research and Development Institute, for their helpful comments and suggestions. I wish to extend my thank to Mr. S. S. Yuk for his preparation of Fig. 1 and to Mr. H. S. Choo for his drawing of figures in this paper.

REFERENCES

Federov, K. N., 1978. The Thermohaline Finestructure of the Ocean. Pergamon Press, Oxford, 170 pp.

Japan Meteorological Agency, 1981. Data from Ocean Data Buoy Stations, No. 4 (1978—1980). Japan Mateorological Agency, 158 pp.

Kim, H. J. and K. D. Cho, 1982. Inversion phenomena of density in the Japan Sea. J. Oceanol. Soc. Korea, **17**:51-58.

Kraus, E. B., 1972. Atmosphere-ocean interaction. Clarendon Press. Oxford, 275 pp.

Madsen, O. S., 1977. A realistic model of the wind-induced Ekman boundary layer. J. Phys. Oceanogr., 7:248-255.

Maizuru Marine Observatory, 1972. Marine Meteorological Study of the Japan Sea. Tech. Rep. Japan Meteorol. Agency, No. 80, 116 pp. (in Japanese).

Nagata, Y., 1967a. Shallow temperature inversions at Ocean Station V. J. Oceanogr. Soc. Japan, **23**: 194-200.

Nagata, Y., 1967b. On the structure of shallow temperature inversions. J. Oceanogr. Soc. Japan, **23**: 221-230.

Nagata, Y., 1968. On the structure of shallow temperature inversions. J. Oceanogr, Soc. Japan, **23**:221-230.

Nagata, Y., 1968. Shallow temperature inversions in the sea to the east of Honshu. J. Oceanogr. Soc. Japan, **24**:103-114.

Nakao, T., 1977. Oceanic variability in relation to fisheries in the East China Sea and the Yellow Sea. J. Fac. Mar. Sci. Tech., Tokai Univ., Japan, Spec. No., Nov-1977, 190-367.

Stern, M. E. 1975. Ocean circulation physics, Academic Press, N.Y., 246 pp.

Wyrtki, K., 1966. Seasonal variation of heat exchange and surface temperature in the North Pacific Ocean. Hawaii Institute of Geophysics Report HIG-66-3, Univ. of Hawaii, 8 pp. with 72 Figs.

해양 표층 수온 역전의 원인

강 용 규

부산수산대학 해양학과

要 約

동계 동해와 황해 표층 수십 메터에서 불안정 성층을 수반한 해양 온도 역전 현상이 종종 발생한다. 수온 역전층 발생의 원인으로서는 (1) 해면으로부터의 열손실을 보상하기 위한 해양 내부의 대류에 기인한 수온 역전, (2) 계절적 수온 변화가 하방으로 전달되면서 진폭과 위상이 바뀜에 따른 동계의 수온 역전, 그리고 (3) 냉수역에서 온수역으로 취송류가 이동함에 따른 수온 역전 등을 들 수 있는 바, 이 논문에서는 위의 세가지 수온 역전 발생 원인에 대하여 해석적인 모델을 제시한다.