

Geochemical Study on the Abyssal Water in the Pacific

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Abstract

The steady state one-dimensional model was applied to the deep Pacific. This model is eligible for application to the deep water between 1.5 and 4.5 km deep in the North Pacific, and is not applicable in the location farther south than 20°S. The observed distribution of salinity, temperature, radium-226 and radiocarbon gives the estimate for the vertical eddy diffusivity, D_z , and the vertical advection velocity, w . The numerical values of D_z and w were evaluated to be 1 cm²/sec and 1×10^{-5} cm/sec respectively. The author has taken the particulate flux into consideration in interpreting the profiles of radium and radiocarbon. Using the foregoing values, the rate of supply for the dissolved silica by the dissolution of particulate in the deep water was computed to be 1.1×10^{-4} μ g atom Si/l/day.

Introduction

The steady state one-dimensional model was first applied to the distribution of ²²⁶Ra by KOCZY (1958), subsequently to oxygen by WYRTKI (1962), to T, S, ¹⁴C and O₂ by MUNK (1966), to ¹⁴C by CRAIG (1969), to Si by GRILL (1970), to O₂ by CRAIG (1971) and to nutrients by TSUNOGAI (1972). However, the steady state one-dimensional model can not be applied to all the oceans or all the depth levels. In applying the model to real profiles, several conditions must be satisfied.

The first is that the steady distribution is maintained. MUNK (1966) regarded the distribution as steady since he found that there was no significant alteration of temperature and salinity in the period of 21 years at the Mindanao Deep of the Pacific. It seems unreasonable to regard the distribution of elements in the deep water as being non-steady in the latest few thousand years, because there was no climatic change during these periods.

The second is the linear relationship between potential temperature and salinity. The observed temperature have to correct at each depth for the adiabatic increase. A linear relationship of T-S diagram can be attributed to the mixing of two independent water types. However, the linear relationship in T-S diagram gives us only an indication of a water column where one-dimensional model is eligible for application. Since potential temperature and salinity are regarded to be stable and conservative, the profile of temperature and salinity is expressed by the solution of the following equation,

$$D_z \frac{\partial^2 M}{\partial z^2} + w \frac{\partial M}{\partial z} = 0 \quad (1)$$

where D_z denotes the vertical eddy diffusivity and w the vertical advection velocity. As it is obvious from eq. (1), the feature of a profile is described only by the value of w/D_z .

The third criterion is that the values of D_z and/or w are kept constant over the entire column where the linearity in T-S relationship is maintained. Then, we must check the constancy of D_z and/or w .

In this study, the author intends to show the linear relationship between potential temperature and salinity in the deep Pacific along 170°W, and determine the depth levels to which one-dimensional

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model is applicable. The w/D_z value is determined over the region where the T-S relationship is linear, and the constancy of w/D_z in several subdivisions of a vertical column is discussed. To get the values of w and D_z individually, we need to introduce radioactive nuclids such as ^{226}Ra and ^{14}C into consideration. In this procedure, chemical and biological effects on ^{226}Ra and ^{14}C profiles are estimated from barium and stable carbon isotopes profiles respectively. Finally, an attempt is made to treat the profile of dissolved silica in the deep water by connecting the obtained D_z and w with the dissolution rate of silica.

Temperature and salinity

Figures 1 and 2 show the linear relationship between potential temperature and salinity in the deep water lower than $3 \sim 4^\circ\text{C}$ in the North Pacific and the South Pacific along the meridional line of 170°W on the basis of data from cruises KH-68-4 and KH-70-2 of the R. V. Hakuho-maru of the University of Tokyo (Ocean Research Institute, 1970, 1972).

As shown in Figure 1, there exist several features in T-S relationship in the North Pacific: a) the straight line shifts toward high salinity side from north to south, b) with the exception of locations at 24°N and 21°N , there is a bend in straight line, and c) all the lines converge to the point with the values of 1.12°C and 34.695% , and the conversion point may characterize the North Pacific bottom water. As

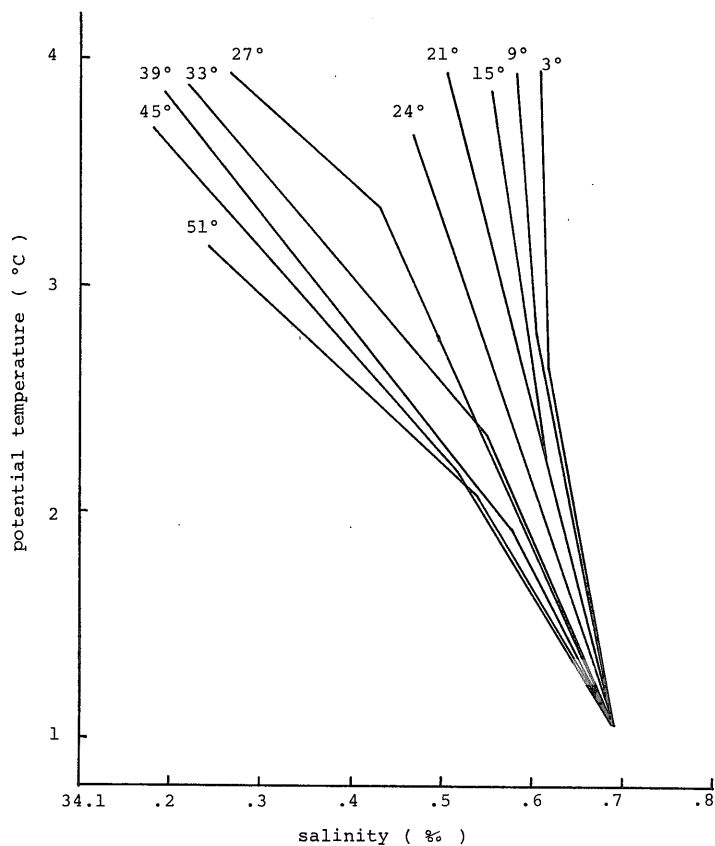


Fig. 1 T-S relationship in the North Pacific along 170°W .

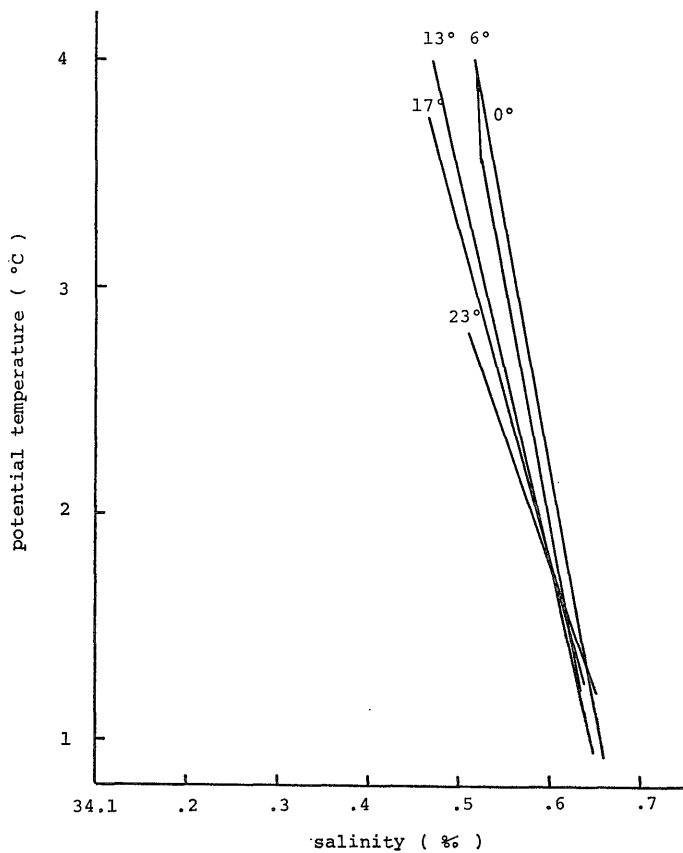


Fig. 2 T-S relationship in the South Pacific along 170°W.

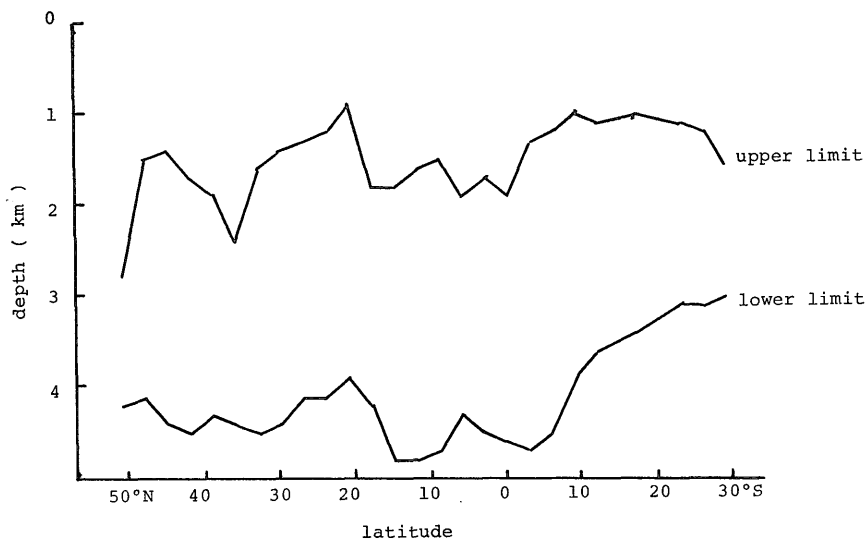


Fig. 3 Boundary of water column, in between these boundaries the linear relationship of T-S is maintained. Profile is along 170°W.

shown in Figure 2, the straight lines in the South Pacific gradually move toward the low salinity side from equator southward.

Figure 3 shows the boundary of vertical column in the Pacific, and the linear T-S relationship is maintained between these boundaries. From the figure it is evident that in the North Pacific there is a water mass in the depth region between 1.5 and 4.5 km in which the one-dimensional model is applicable, but in the South Pacific we encounter typically different water mass that penetrates wedgewise toward the north from polar region, so that the overlying water mass is pushed up and the lower limit of the column becomes shallower. For this reason, in the location farther south than 20°S, one-dimensional model is not eligible for application to the deep water.

The water column to which the one-dimensional model is applicable is divided further into blocks with the thickness of 2 km. The value of w/D_z is derived from the best-fit of eq. (2), the solution of eq. (1), to the real profile of salinity and/or potential temperature,

$$\frac{M - M_1}{M_2 - M_1} = \frac{\exp[-w/D_z \cdot (z - z_1)] - 1}{\exp[-w/D_z \cdot (z_2 - z_1)] - 1} \quad (2)$$

where M_1 and M_2 represent the boundary values of salinity or potential temperature at z_1 and z_2 km depth respectively. Here z is taken to be positive downward and $z_1 = 0$ is set at the real depth of z_1

Table 1-a Calculated w/D_z from salinity profile.

Station	Latitude	Depth (m)	w/D_z (km ⁻¹)	Station	Latitude	Depth (m)	w/D_z (km ⁻¹)
KH-70-2-2	48°00'N	1800-3800	0.9	KH-68-4-5	26°55'N	1400-3400	1.3
KH-70-2-3	45°00'N	1400-3400	1.0			1800-3800	1.0
		1800-3800	1.0	KH-68-4-8	23°59'N	1400-3400	1.2
		2200-4200	1.1			1800-3800	1.1
KH-70-2-4	42°03'N	1800-3800	1.2	KH-68-4-11	21°01'N	1000-3000	0.7
KH-70-2-5	39°01'N	2200-4200	1.4			1400-3400	0.9
KH-70-2-6	35°56'N	2600-4600	0.9			1800-3800	1.2
KH-70-2-7	33°03'N	1800-3800	1.2	KH-68-4-13	17°59'N	1800-3800	0.9
		2200-4200	0.8	KH-68-4-14	14°58'N	1800-3800	1.0
KH-70-2-8	30°03'N	1400-3400	1.5			2200-4200	0.8

All stations are along 170°W.

Table 1-b Calculated w/D_z from salinity profile.

Station	Latitude	Depth (m)	w/D_z (km ⁻¹)	Station	Latitude	Depth (m)	w/D_z (km ⁻¹)
KH-68-4-15	12°00'N	1800-3800	0.8	KH-68-4-21	06°09'S	1400-3400	1.0
KH-68-4-16	09°02'N	1800-4600	?			1800-3800	0.7
KH-68-4-17	06°05'N	2200-4200	?			2200-4200	0.5
KH-68-4-18	03°02'N	1800-3800	0.6	KH-68-4-22	09°05'S	1400-3400	1.0
		2200-4200	0.7			1800-3800	1.0
KH-68-4-19	00°03'N	2200-4600	?	KH-68-4-23	12°00'S	1400-3400	1.3
KH-68-4-20	03°02'S	1400-3400	0.8	KH-68-4-24	17°01'S	1000-3000	0.8
		1800-3800	0.8			1400-3400	0.8
		2200-4200	0.6				
		2600-4600	0.7				

km. As shown in Table (1), the obtained values of w/D_z range from 0.5 to 1.5 km^{-1} . Even in a water column where T-S relationship is linear, the value of w/D_z for each subdivided column is not always constant. For example, the value of w/D_z in 30°N station decreases from 1.5 to 1.0 km^{-1} downward, but in 45°N station it is almost constant over an entire column. Therefore, it is by no means correct that the profile of conservative tracer over the entire column where T-S linear relationship is established throughout can be described by a constant value of w/D_z . The value of w/D_z is taken to be 1 km^{-1} as an average for the deep Pacific.

The sign for w determined over the entire Pacific in the present study is always positive. In other words, the upward advection prevails in the abyssal water of the Pacific. This upwelling of water is not only a characteristic feature in the deep North Pacific (KNAUSS, 1962), but also prominent in the deep South Pacific. Then, it may also be reasonable to state that the vertical distribution of a conservative component is governed by a balance between the vertical diffusion and the effect of the rising water.

Radium-226 and radiocarbon

The stable conservative tracers such as salinity and/or temperature give us only ratio of w and D_z on the basis of the steady state one-dimensional model. To get the values of w and D_z individually, we need further information pertinent to the problem. For this purpose, it is suitable to introduce radioactive nuclides such as ^{226}Ra and ^{14}C into consideration. ^{226}Ra and ^{14}C are appropriate tracers to study the mixing and circulation problems in the deep water since both nuclides have adequate half-lives, although there are only few observed data available.

The concentration of a radioactive nuclide M is expressed by the following equation under the condition of the steady state one-dimension,

$$D_z \frac{\partial^2 M}{\partial z^2} + w \frac{\partial M}{\partial z} + P - \lambda M = 0 \quad (3)$$

where P denotes the production rate and λ the decay constant of a radioactive nuclide.

The production rate of ^{14}C is negligible in the deep water since ^{14}C is cosmic-ray product. As to ^{226}Ra , the production of ^{226}Ra from ^{230}Th is negligible owing to much lower concentration of ^{230}Th in comparison with that of ^{226}Ra (MOORE and SACKETT, 1964, SOMAYAJULU and GOLDBERG, 1966, MIYAKE, SUGIMURA and YASUJIMA, 1970, and MIYAKE, SUGIMURA and MATSUMOTO, 1972). Therefore, eq. (3) is simplified to

$$D_z \frac{\partial^2 M}{\partial z^2} + w \frac{\partial M}{\partial z} - \lambda M = 0 \quad (4)$$

Eq. (4) has the following solution under the conditions $M = M_1$ at $z = z_1 (=0)$ and $M = M_2$ at $z = z_2$,

$$M = \frac{M_2 (e^{\alpha z} - e^{\beta z}) + M_1 (e^{\alpha z_2 + \beta z} - e^{\beta z_2 + \alpha z})}{e^{\alpha z_2} - e^{\beta z_2}} \quad (5)$$

where

$$\alpha = 1/2 \cdot (-w/D_z + \sqrt{(w/D_z)^2 + 4w/D_z \cdot \lambda/w})$$

$$\beta = 1/2 \cdot (-w/D_z - \sqrt{(w/D_z)^2 + 4w/D_z \cdot \lambda/w})$$

By using the value of w/D_z obtained by the profile of salinity, the numerical value of a new parameter,

λ/w can be determined from the best-fit of eq. (5) to the real profile of a radioactive nuclide. When the value of λ/w is determined, w can be calculated since λ is known. As w/D_z is already known, D_z can be calculated in turn.

MUNK (1966) has used eq. (5) to interpret the ^{14}C and ^{226}Ra profiles. However, CRAIG (1969) has pointed out in an elaborate manner that the vertical distribution of ^{14}C of inorganic carbon is affected by the particulate flux. Also, WOLGEMUTH and BROECKER (1970) has found that there is a close relationship between ^{226}Ra and barium distribution in sea water. They suggested that the enrichment of barium in the deep water relative to the surface water must reflect the uptake of barium by particulate formed in surface water and subsequent release of barium to the deep water as particulates decompose on the way to bottom. This conclusion may also be valid interpreting the radioactive nuclide with the effect of particulate dissolution. We need one more term relevant to the particulate flux in eq. (4). The fundamental equation can be written in the following form,

$$D_z \frac{\partial^2 M}{\partial z^2} + w \frac{\partial M}{\partial z} - \lambda M + J = 0 \quad (6)$$

where J denotes the dissolution rate that is assumed to be independent on both concentration and depth. Eq. (6) has the solution under the conditions of $M = M_1$ at $z = z_1$ and $M = M_2$ at $z = z_2$ as follows:

$$(M - J/\lambda) = \frac{(M_2 - J/\lambda)(e^{\alpha z} - e^{\beta z}) + (M_1 - J/\lambda)(e^{\alpha z_2 + \beta z_2 - e^{\beta z_2 + \alpha z}})}{e^{\alpha z_2} - e^{\beta z_2}} \quad (7)$$

The new parameter, J/λ of eq. (7) is transformed as follows:

$$J/\lambda = J/w \cdot w/\lambda$$

The value of J/w can not be directly known, but estimated if we take a element with the same chemical behavior or stable isotope into consideration. Therefore, only unknown parameter to be determined in eq. (7) as well as in eq. (5) is λ/w .

The numerical value of J_{Ra}/w is determined from the profile of barium as follows:

$$J_{\text{Ra}}/w = J_{\text{Ba}}/w \cdot (C_{\text{Ra}}/C_{\text{Ba}}) \text{ surface water}$$

where C_{Ra} and C_{Ba} denote the concentration of radium and barium in surface water respectively. This implies that the ratio of radium to barium concentration in the particulate can be taken to be similar to that of surface water (SZABO, 1967). The value of J_{Ba}/w is given by the solution of the following equation,

$$D_z \frac{\partial^2 M}{\partial z^2} + w \frac{\partial M}{\partial z} + J_{\text{Ba}} = 0 \quad (8)$$

The solution is

$$\frac{M - M_1}{M_2 - M_1} = \frac{e^{-w/D_z \cdot z} - 1}{e^{-w/D_z \cdot z_2} - 1} \left(1 + J_{\text{Ba}}/w \cdot \frac{z_2}{M_2 - M_1} \right) - J_{\text{Ba}}/w \cdot \frac{z}{M_2 - M_1} \quad (9)$$

Previously reported ^{226}Ra and barium profiles are badly scattered and there are only few available data. ^{226}Ra profile to be used here is taken from BROECKER, LI and CROMWELL (1967) and barium profile from WOLGEMUTH and BROECKER (1970) both for the North Pacific.

The curve of barium profile in Figure 4 was subjected to the fitting by eq. (9). The range of water column is taken to be from 1.5 to 4.5 km as previously determined, where one-dimensional model is applicable in the North Pacific. The best-fit was found when J_{Ba}/w was 3.6×10^{-6} g/l/km with the assumption of $w/D_z = 1.0 \text{ km}^{-1}$, which is reasonable from the previous discussion in salinity profile.

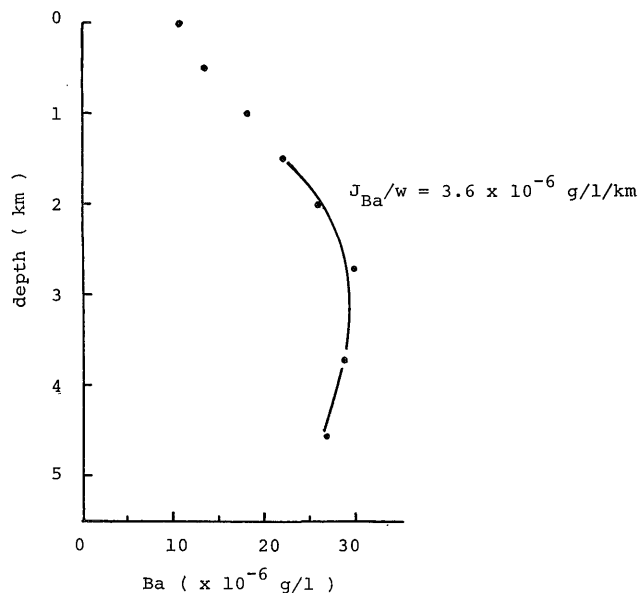


Fig. 4 Profile of barium in the Northeast Pacific given by WOLGEMUTH and BROECKER (1970).

By using $C_{Ra} = 4.0 \times 10^{-14}$ g/l and $C_{Ba} = 10.5 \times 10^{-6}$ g/l, the value of J_{Ra}/w comes out to be 1.4×10^{-14} g/l/km. On the basis of these known values, the radium profile shown in Figure 5 can be represented by eq. (7) with the best-fit value of $\lambda/w = 0.14$ km⁻¹. From this value w is calculated to be 1.0×10^{-5} cm/sec, and subsequently D_z to be 1.0 cm²/sec. Without the particulate flux, the best-fit value of diffusivity turns out to be nearly twice the above value when eq. (5) is used.

The rate of ²²⁶Ra supply by the dissolution of particulate, J_{Ra} was estimated to be 4.3×10^{-17} g/l/yr by using the obtained value of $w = 1.0 \times 10^{-5}$ cm/sec, and the decay rate, λM , to be 6.0×10^{-17} g/l/yr by using the value of $M = 14 \times 10^{-14}$ g/l for the average concentration of ²²⁶Ra in the deep layer. As seen from the above result, J_{Ra} is about the same as λM , which means that the dissolution plays an important role as well as decay does in ²²⁶Ra profile.

CRAIG (1969) has pointed out that J_{C^*}/w is determined from the total CO₂ and ¹³C profiles as follows:

$$J_{C^*}/w = \alpha_C \cdot J_C/w$$

where C* denotes the normalized radiocarbon concentration defined by CRAIG (1969) and α_C the conversion factor determined on the basis of a profile of ¹³C.

The data observed at Geosecs station were used in this study since there are all values required for the calculation. CRAIG and WEISS (1970) have reported that the numerical values of J_C/w and w/D_z are calculated to be 0.0474 m mole/kg/km and 1.16 km⁻¹ on the basis of total CO₂ and salinity profiles respectively. The numerical value of α_C was determined to be 0.938 on the basis of ¹³C profile given by KROOPNICK, DEUSER and CRAIG (1970). Therefore, the value of J_{C^*}/w comes out to be 0.0445 m mole/kg/km. Finally, the value of λ/w was determined from the best-fit of eq. (7) to the profile of a radiocarbon given by ÖSTLUND and NISKIN (1970), using the foregoing values. The best-fit value of λ/w was 0.039 km⁻¹, and from this value w is evaluated to be 1.0×10^{-5} cm/sec and subsequently D_z to be 0.9 cm²/sec, as shown in Figure 6. These values agree well with those that is calculated from the radium profile as

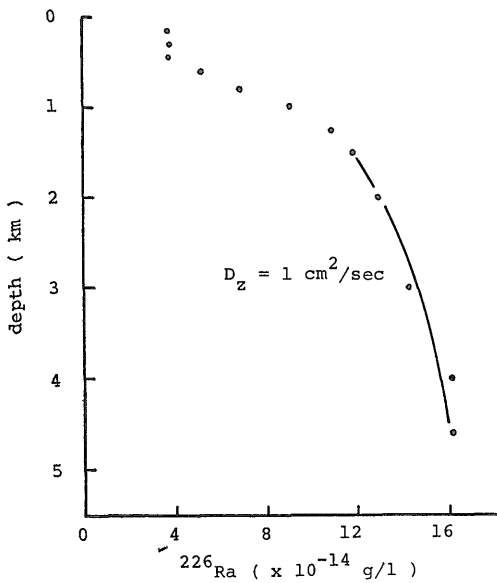


Fig. 5 Profile of ^{226}Ra in the Northwest Pacific given by BROECKER, LI and CROMWELL (1967).

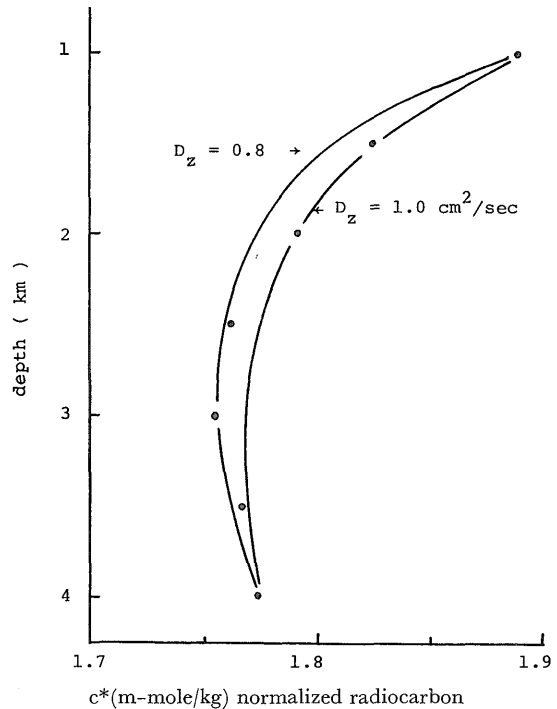


Fig. 6 Profile of ^{14}C in the North Pacific given by ÖSTLUND and NISKIN (1970).

previously mentioned.

By using the values of $J_c/w = 0.0474$ m mole/kg/km and $w = 1.0 \times 10^{-5}$ cm/sec, the value of J_c comes out to be 4.1×10^{-7} m mole/l/day. Therefore, the amount of integral supply for the inorganic carbon by the particulate carbon in the deep layer is 20 mg C/m²/day, when the thickness of the deep layer is taken to be 4 km. Provided that the primary production rate is 150 mg C/m²/day for an average of the North Pacific (ICHIMURA, 1965), this value is about 13% relative to the assimilated amount of inorganic carbon in the surface layer by biota.

MUNK (1966) derived the vertical diffusivity and advection velocity from ^{14}C profile without taking particulate flux into consideration. He estimated $D_z = 1.3$ cm²/sec and $w = 1.4 \times 10^{-5}$ cm/sec in the interior Pacific. CRAIG (1969) estimated $D_z = 0.6 - 6$ cm²/sec and $w = 0.6 - 6 \times 10^{-5}$ cm/sec in the deep Pacific. The obtained values of D_z and w in his calculation give a wide range because of the scatter of individual measurements given by BIENS, RAKESTRAW and SUESS (1965). Recently, TSUNOGAI (1972) derived the numerical value of J_{Ra}/w from calcium profile considering the particulate flux. He estimated to be 1.2×10^{-14} g/l/km for the value of J_{Ra}/w on the basis of the assumption that the increase of alkalinity is due to the dissolution of calcium carbonate and J_{Ra} is proportional to J_{Ca} , and the estimated value agrees well with that obtained in this study. KNAUSS (1962) estimated that the vertical upward advection velocity would be 2.5×10^{-5} cm/sec, if the Pacific deep water moving from south to north rises uniformly over the entire North Pacific north 10°N. Since KNAUSS' estimate is made by a simple and rough way, his value is larger than $w = 1 \times 10^{-5}$ cm/sec that is calculated in the present study.

Dissolved silica

The numerical values of D_z and w in the deep layer of the North Pacific have already been obtained, and hence, the rate of dissolution of silica can be estimated on the basis of the obtained values of D_z and w .

All the elements involved in the biological cycle of the ocean used to show variation in concentration with depth. The primary photosynthetic process occurs in the top 100 meters of the ocean. Silicon composes the skeletal part of some plankton, and silicon as one of the nutrient components is readily removed from the surface water as well as phosphorous and nitrogen. During the descent of biologically derived particles they decompose and most of elements in particles regress to sea water.

OKUBO (1954, 1956) and GRILL (1970) proposed mathematical models for the dissolved silica profile. OKUBO's model consists of the balance between loss by vertical diffusion and gain by the dissolution of particulate silica. GRILL (1970) proposed the model that contains the term of diffusion, advection and the rate of silica dissolution dependent on depth. However, they applied the model to too shallow depth region, where the model is not eligible for application, because the shallow depth region must be treated in a different manner against the deep layer. In the shallow region, the physical parameter, D_z and w , would not be equal to those of the deep layer, and the rate of dissolution may also depend on depth. On the other hand, it is possible to interpret the profile by rather simple treatment in the deep layer, since the parameters are regarded to be relatively constant over a certain range of depth.

The profile of dissolved silica in the deep layer can be described by the same equation as eq. (8) for barium profile. The term J_{Si} implies that assimilated silica in particulate matter in surface water sinks and supplies the dissolved silica with a constant rate through the dissolution of particulate into the deep water.

The dissolved silica profiles in this study are restricted to the same depth ranges as those of salinity profiles found to be suitable for the application of one-dimensional model, and the calculation was done on the several stations for high-, middle- and low-latitudes of the North Pacific along 170°W and on the depth region between 1.8 and 3.8 km. As previously mentioned in barium profile, the value of J_{Si}/w is derived from the best-fit of eq. (9) to the real profile of dissolved silica.

The obtained values of J_{Si}/w are listed in Table 2 together with w/D_z obtained from the salinity profile, and the values range 7 to 18 $\mu\text{g atom Si/l/km}$, and an average value of J_{Si}/w is taken to be 13 $\mu\text{g atom Si/l/km}$. An example of the fitting of the calculated profile to the observed profile is shown in Figure 7. By using the values of $J_{Si}/w = 13 \mu\text{g atom Si/l/km}$ and $w = 1 \times 10^{-5} \text{ cm/sec}$ obtained on the basis of the profiles of radium and radiocarbon, then $J_{Si} = 1.1 \times 10^{-4} \mu\text{g atom Si/l/day}$. Therefore, the rate of supply for the dissolved silica by the dissolution of particulate in the deep layer is calculated to be 0.44 $\text{mg atom Si/m}^2/\text{day}$ when the thickness of the deep layer is taken to be 4 km.

REDFIELD, KETCHUM and RICHARDS (1963) showed that the ratio of the assimilated carbon to silicon should be about 5:1 by atom in the case of phytoplankton. According to ICHIMURA (1965), it is reasonable to assume the range of 100 to 300 $\text{mg C/m}^2/\text{day}$ for the primary production rate in the open ocean of the North Pacific. Provided that the primary production rate is 150 $\text{mg C/m}^2/\text{day}$ and the ratio of carbon to silicon is taken to be 5:1, the amount of integral supply for dissolved silica by the biogenous silica in the deep layer is about 20% relative to assimilated amount of dissolved silica in the surface

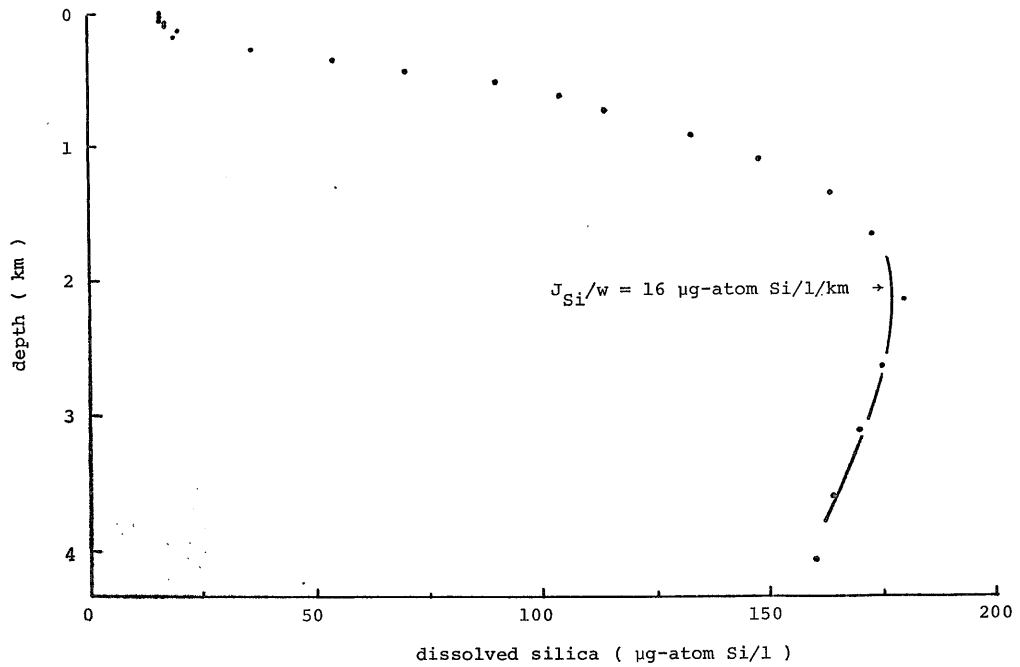


Fig. 7 Profile of dissolved silica at KH-70-2-4 of the North Pacific.

Table 2 Calculated values of J_{Si}/w and w/D_z on the basis of the dissolved silica and salinity profile respectively.

Station	Latitude	w/D_z (km^{-1})	J_{Si}/w ($\mu\text{g-atom Si/l/km}$)
KH-70-2-2	48°00'N	0.9	11
KH-70-2-3	45°00'N	1.0	18
KH-70-2-4	42°00'N	1.2	16
KH-70-2-7	33°03'N	0.8	7
KH-70-2-8	30°03'N	1.2	13
KH-68-4-5	26°55'N	1.0	15
KH-68-4-13	17°59'N	0.9	13
KH-68-4-14	14°58'N	1.0	11
KH-68-4-15	12°00'N	0.8	14

All stations are along 170°W and the calculation is done on the depth region between 1.8 and 3.8 km.

layer.

Conclusions

The steady state one-dimensional model was applied to the deep Pacific. The linear relationship between potential temperature and salinity gives us an indication of a water column where one-dimensional model is eligible for application. From point of view of the T-S relationship, this model is eligible for application to the deep water between 1.5 and 4.5 km deep in the North Pacific, but is not applicable in the location farther than 20°S because of a water mass that penetrates wedgewise toward the north

from polar region. The value of w/D_z is derived from the best-fit to the profile of salinity, and obtained values of w/D_z range from 0.5 to 1.5 km^{-1} and the value is taken to be 1.0 km^{-1} as an average for the deep Pacific. The profiles of radium-226 and radiocarbon give the estimate for the vertical eddy diffusivity, D_z , and the vertical advection velocity, w . The numerical values of D_z and w were evaluated to be 1 cm^2/sec and 1×10^{-5} cm/sec respectively. The profiles of these nuclides are affected by the particulate flux. The value of J/w is estimated if we take barium profile for radium and stable carbon isotopes for radiocarbon into consideration. By using the foregoing value of w , then $J_{Ra} = 4.3 \times 10^{-17}$ $\text{g}/\text{l}/\text{yr}$, $J_C = 4.1 \times 10^{-7}$ $\text{m mole}/\text{l}/\text{day}$ and $J_{Si} = 1.1 \times 10^{-4}$ $\mu\text{g atom Si}/\text{l}/\text{day}$. The amount of integral supply for dissolved carbon and silica by the particulate flux in the deep layer is about 13 and 20% relative to the assimilated amount in the surface layer respectively.

Acknowledgements

I am indebted to Dr. S. MATSUO for his discussions and suggestions of these problems.

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太平洋の深層水についての地球化学的研究

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要 旨

太平洋の深層水に、定常状態—次元モデルを適用した。このモデルは、北太平洋においては、1.5 から 4.5 km の深さの深層水について適用可能であり、南太平洋の 20°S 以南では、モデルの適用は不可能である。塩分、温度、²²⁶Ra および ¹⁴C の分布から、深層水の鉛直渦動拡散定数および鉛直移流速度を求めると、おのおの 1 cm²/sec および 1 × 10⁻⁵ cm/sec の値が得られた。この値を用いて、深層水中での粒状シリカの溶解速度を計算すると、1.1 × 10⁻⁴ μg atom Si/l/day であった。