Effect of Ocean Thermal Diffusivity on Global Warming Induced by Increasing Atmospheric CO_2

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ABSTRACT

A global mean ocean model including atmospheric heating, heat capacity of the mixed layer ocean, and vertical thermal diffusivity in the lower ocean, proposed by Cess and Goldenberg (1981), is used in this paper to study the sensitivity of global warming to the vertical diffusivity. The results suggest that the behaviour of upper ocean temperature is mainly determined by the magnitude of upper layer diffusivity and an ocean with a larger diffusivity leads to a less increase of sea surface temperature and a longer time delay for the global warming induced by increasing CO_2 than that with smaller one. The global warming relative to four scenarios of CO_2 emission assumed by Intergovernmental Panel of Climate Change (IPCC) is also estimated by using the model with two kinds of thermal diffusivities. The result shows that for various combinations of the CO_2 emission scenarios and the diffusivities, the oceanic time delay to the global warming varies from 15 years to 70 years.

I. INTRODUCTION

One of the major means in the studies of climate change as the result of increasing atmospheric CO_2 concentration is numerical climate modelling. The early climate models are atmospheric general circulation models (AGCMs) and their recent versions are coupled atmospheric– oceanic general circulation models. Although the atmosphere is the central component of the climate system and a number of research works on climate changes have been completed by using AGCMs, the world ocean always plays a critical role and can not be ignored. As a matter of fact, the ocean effect must be taken into account in certain manner in any AGCMs. One of the earliest estimation of the global warming resulting from the doubling of CO_2 was given by Manabe and Wetherald (1975) by using the GFDL AGCM, in which the ocean was treated as a swamp without heat capacity at that time. The global mean temperature increase given by the model is about 2.9℃ and the response of the model's atmosphere to CO_2 doubling was accomplished within only 1 year. It is obvious that the ocean's delay for the global warming was not be taken into account in their studies. On the other hand, the magnitude of the global mean temperature increase estimated by Gates et al. (1981) with the OSU AGCM, in which the observational sea surface temperature (SST) is prescribed as the model's lower boundary condition, is only 0.3℃. This is not surprising because such a treatment for ocean in their model implies that an infinite ocean heat capacity is assumed and, therefore, the green house warming must be underestimated.

Such a kind of uncertainties in estimating the global warming stimulates the development of coupled atmosphere–ocean models to improve the representation of ocean's effect on climate change. At the present time, however, the majority of research works in this fields was conducted by using coupled atmosphere–mixed layer ocean models due to the limitation of computer resources and the less advanced development of full oceanic general circulation models (OGCMs). The global mean temperature increase estimated by using such a kind of models is ranging between 1.9–5.2℃ (IPCC, 1990) and the equilibrium response of model's atmosphere to the doubling of CO_2 can be reached within five years or so that represents only the delay effect of the mixed layer ocean but not the deep ocean.

The ocean's effect on the global warming is mainly exerted through the changes of SST, which governs the energy exchange between the atmosphere and the ocean. For instance, the slowly increasing SST is favourable to the penetration of atmospheric thermal anomalies into the lower ocean because of the maintenance of a dramatic thermal contrast between the atmosphere and the ocean. On the contrary, the ocean's delay effect on the global warming will not be significant while SST is increasing rather rapidly because of the air–sea energy exchange being depressed. Therefore, it is quite important to investigate the behaviour of SST variation in the warming event of the atmosphere–ocean system.

On the other hand, the changes of SST are closely related to the vertical thermal penetration processes in the lower ocean. Generally speaking, there are three ways for the thermal penetration, namely the cold water upwelling caused by vertical advection, the convective overturning in high latitude regions, and the vertical turbulence diffusivity. In sense of global average, the vertical diffusivity may play a substantial role due to the small fraction of areas for both the upwelling and the convective overturning. Unfortunately, the vertical diffusivity is one of the most uncertain subgrid–scale parameterized processes in ocean models due to the lack of observational data. It was estimated that the typical magnitude of order of the vertical diffusivity coefficient in the ocean is $10^{-4}m^2s^{-1}$ and it varies from 0.3 to $5.0\times10^{-4}m^2s^{-1}$ (F. Bryan, 1987).

Moreover, the vertical diffusivity is far from homogeneous in vertical direction e.g., according to K.Bryan et al. (1979) the vertical profile of the diffusivity may be fitted by using the following equation.

$$k = k(z) = 10^{-4}(0.8 + 1.05/\pi \cdot \tan^{-1}(4.5 \cdot (z-2500)/1000)) \ , \tag{1}$$

where z is the ocean depth in meter. Thus the values of k are about $0.3\times10^{-4}m^2s^{-1}$at the surface and $1.3\times10^{-4}m^2s^{-1}$ at the bottom respectively, and there is a turning point at $z=2500$ meters. However, Pacanowski et al.(1981) proposed another different choice for k in their tropical OGCM, which is proportional inversely to Richardson number (Ri) and has a value at the sea surface being one order larger than those in the deep ocean. In brief, the vertical diffusivity coefficient, as a critical parameter in ocean models, has not been precisely determined.

The purpose of this paper is to study the effect of oceanic thermal diffusivity process on the global warming. The global mean ocean model proposed by Cess and Goldenberg (1981) including the radiation heating of the surface–troposphere system due to increasing atmospheric CO_2, the heat capacity of the mixed layer of the ocean, and the vertical thermal diffusivity in the lower ocean is used in this study with an emphasis on the sensitivity of the warming event to the choice of the vertical diffusivity. The heat penetrating into the lower ocean and the long term oceanic response to the atmospheric thermal anomaly are described in Section 2. In Section 3, the sensitivity of the time delay of the ocean for the global warming to vertical diffusivities is dis–cussed based on a series of numerical experiments with the model. Finally, the variation of the ocean's delay for global warming related to the four scenarios of future CO_2emission proposed by Intergovernmental Panel on Climate Change (IPCC, 1990) is estimated in Section 4.

II. THERMAL PENETRATING TOWARD DEEP OCEAN GOVERNED BY DIFFUSIVITY

Let $T = T(z,t)$ be the global horizontal averaged ocean temperature, where z is height (negative beneath the sea level) and t is time, the process of vertical thermal diffusion in the ocean can be described by using the following equation and boundary conditions:

$$\frac{\partial T}{\partial t} = \frac{\partial}{\partial z}\left(k\frac{\partial T}{\partial t}\right) , \tag{2}$$

$$k\frac{\partial T}{\partial t}\bigg|_{z=0} = \frac{1}{\rho_0 c_p} F_0(t) , \tag{3}$$

$$k\frac{\partial T}{\partial t}\bigg|_{z=H} = 0 , \tag{4}$$

$$T(z,0) = \tilde{T}_0(z) , \tag{5}$$

where k is the thermal diffusivity coefficient, ρ_0 is the density of sea water, c_p is the specific heat capacity at constant pressure, $F_0(t)$ is the heat flux from the atmosphere to the ocean, H is the global mean depth of the world ocean, $\tilde{T}_0(z)$ is a certain given initial vertical distribution of the ocean temperature. In this calculation, we take

$$\rho_0 = 1029 \text{ kg m}^{-3} \ ; \ c_p = 3901\text{J} \cdot \text{kg}^{-1} \cdot \text{k}^{-1} \ ; \ H = 4000 \text{ m} .$$

Here $\tilde{T}_0(z)$ is the observed global mean ocean temperature based on Levitus (1982), in which the global mean temperature in the uppermost layer of 50 meters is about 18℃, sharply decreases with depth within 50–1500 meters, and has a very small lapse below 1500 meters. $F_0(t)$ is calculated by using the bulk aerodynamic formula:

$$F_0(t) = \mu(\tilde{T}_a - T_s) , \tag{3}$$

where μ is relaxation coefficient, $\tilde{T}_a$ is the observed global and annual mean sea level air temperature, taken as 18.8℃ , T_s is the model's SST i.e., $T_s = T(0,t)$. Thus the initial air–sea temperature difference is 0.8℃ at the sea surface, which determines an initial downward heat flux from the atmosphere to the ocean. To solve the problem numerically, the model's ocean is divided into forty layers with 100 meter interval and the differential equation is replaced by the corresponding finite–difference approximation. The time step of the model's integration is six hours.

In order to understand the oceanic thermal diffusion process we have performed three long–term integrations with different choices of the diffusivity coefficient i.e., (a) $k = k(z)$ given by Eq.(1), (b) $k = 0.3 \times 10^{-4}\text{m}^2\text{s}^{-1}$, and (c) $k = 1.3 \times 10^{-4}\text{m}^2\text{s}^{-1}$. For all these cases, the model has been integrated for one thousand years started from the same initial condition. The results show that the thermal penetration into the deep ocean is an extremely slow process and the full equilibrium has not been reached by the end of the integration for all three cases. The increases of the ocean temperature at depths of 50 m, 650 m, 1250 m, and 3950 m during the last one hundred years are listed in Table 1.

Table 1. The Ocean Temperature Increase from the 900th Year to the 1000th Year

	k ($10^{-4}\text{m}^2\text{s}^{-1}$)	–50 m	–650 m	–1250 m	–3950 m
Case (a)	0.3–1.3	0.009℃	0.181℃	0.280℃	0.209℃
Case (b)	0.3	0.02℃	0.127℃	0.322℃	0.105℃
Case (c)	1.3	0.109℃	0.216℃	0.343℃	0.556℃

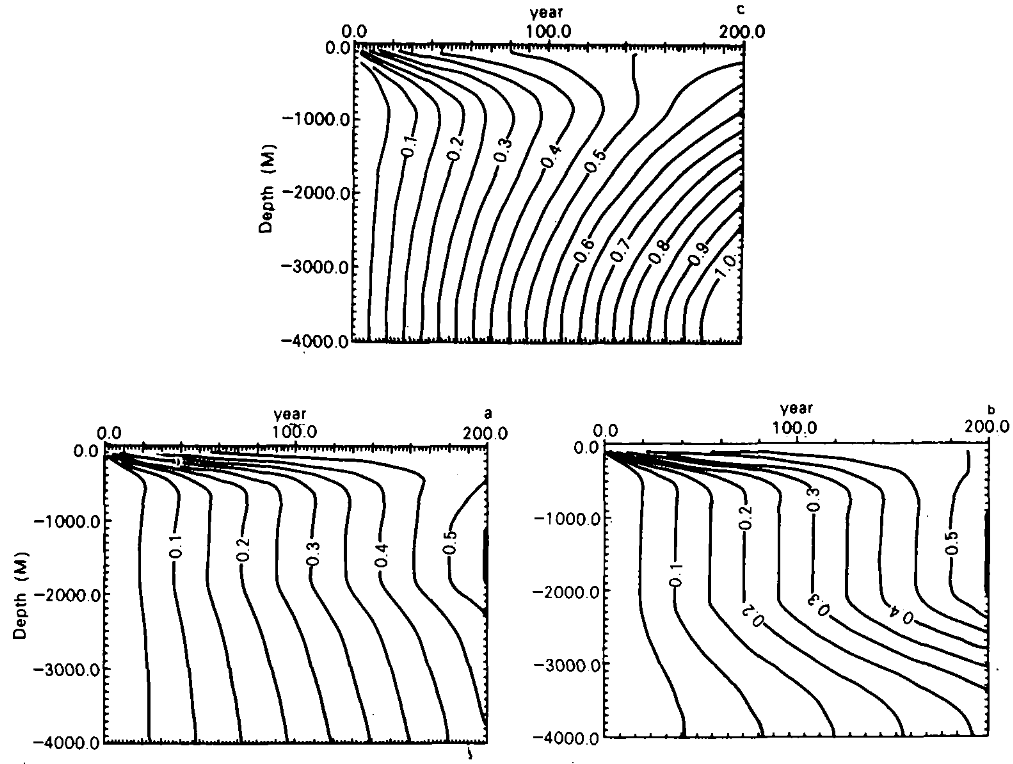

Fig.1. The ocean temperature increase at each depth during the 200 years. The interval of isothermohyps is 0.05°C. (a) $k = k(z)$, (b) $0.3 \times 10^{-4} m^2 s^{-1}$, (c) $k = 1.3 \times 10^{-4} m^2 s^{-1}$.

It can be seen from Table 1 that the surface layer equilibrium may have been reached in Case (a) and (b), reflecting only 0.01—0.02℃ temperature increases at 50 m depth during the last 100 years. But this may not be true for Case (c) because the temperature increase during the last 100 years is 0.109℃, being one order greater than those in Case (a) and (b). In Case (c), the temperature increase at the depth of 3950 m during the last 100 years is 0.556℃ which is about two or five times of those in Case (a) and Case (b) respectively. Comparing the three choices of diffusivity we found that the surface layer temperature is mainly determined by the magnitude of surface layer diffusivity coefficient. The temperature evolution in the deep ocean, however, is influenced by not only the local magnitude of diffusivity but also its vertical structure, and is even controlled mainly by the latter.

On the basis of these 1000 year integrations, three further integrations have been performed for another 200 years with a global mean atmospheric thermal anomaly of 0.5℃ to see how the thermal anomaly is penetrating into the lower ocean due to the effect of the vertical diffusivity.

Figure 1 shows the temperature increases at each depth mentioned above during the 200 years in relation to the final states of the former three integrations. We can see that the surface layer temperature increases much more rapidly in Case (b) ($k = 0.3$ $m^2 s^{-1}$) than in Case (c) ($k = 1.3$ $m^2 s^{-1}$), e.g., a 0.45℃ surface layer anomaly at 50 m depth, which may be regarded as a quasi—equilibrium state, occurs in the 57 th year for Case (b) and in the 80th year for Case (c) respectively. On the other hand, the depths reached by the thermal anomaly are about 500 m in Case (b) and 1000 m in Case (c) respectively. These results suggest that a smaller diffusivity is favourable to the increase of surface layer temperature and a larger diffusivity is favourable to the penetration of thermal anomaly into the lower ocean. It is found that the evolution of the temperature within the upper 500 meter layer in Case (a) is quite similar to that in Case (b) though $k(z)$ in Case (a) has dramatic vertical variation and is far from a single constant. Again, this shows that the behavior of the upper ocean temperature is mainly determined by the magnitude of the upper layer diffusivity in this model.

It should be pointed out that the temperature increase in the lower ocean (below 1000 meters) for all the three cases shown in Fig.1 does not represent the penetration of the initial thermal anomaly and just reflects the residual drift resulting in the unequilibrium in the first 1000 year integration as mentioned above.

III. SENSITIVITY OF OCEANIC DELAY TO VERTICAL DIFFUSIVITY

We have shown in the last Section that the magnitude of the vertical diffusivity in the upper ocean exerts a strong influence upon the response speed of surface layer temperature to atmospheric thermal anomalies. Now we turn to investigate the time scale of the oceanic delay for the global warming induced by increasing CO_2 with a global ocean model proposed by Cess and Goldenberg (1981). The model consists of a 70 m mixed layer of the ocean and the lower ocean which can be represented by using the following equation and the initial–boundary conditions:

$$\frac{\partial\theta_0}{\partial t}=\frac{\partial}{\partial t}\left(k\frac{\partial\theta_0}{\partial z}\right), \tag{6}$$

$$k\frac{\partial\theta_0}{\partial z}\bigg|_{z=-70M}=\frac{1}{\rho_0 c_p}F_0(t), \tag{7}$$

$$k\frac{\partial\theta_0}{\partial z}\bigg|_{z=-H}=0, \tag{8}$$

$$\theta_0(z,0)=0, \tag{9}$$

where $\theta_0=T_0(z,t)-T_0(z,1860)$ is the departure of the ocean temperature from that in 1860 when the concentration of atmospheric CO_2 was at a low level, and $F_0(t)$ is the downward heat flux through the interface between the mixed layer and the lower ocean. Considering the heat budget of the mixed layer ocean, we may get an expression for $F_0(t)$ as follows,

$$F_0(t)=\Delta F_0(t)-B\theta_{os}-R_m\frac{d\theta_{os}}{dt}, \tag{10}$$

where $\Delta F_0(t)$ is the heating of the surface–troposphere system arising from the increasing atmospheric CO_2 concentration related to its value in 1860 (287 ppmv), $B=1.26\ \mathrm{W\cdot m^{-2}C^{-1}}$ and $-B\theta_{os}$ represents the modification of the surface–atmosphere radiation budget resulting from increased mixed layer temperature (or SST), θ_{os}, and the last term in Eq.(10) represents the heat storage of the mixed layer with a heat capacity, R_m, of $3\times10^8\mathrm{W\cdot s^{-1}m^{-2}}$.

According to Cess and Goldenberg (1981), $\Delta F_0(t)$ may be approximated by the expression

$$\Delta F_0(t)(\mathrm{Wm}^{-2}) = 0.0277(\exp\ (t / t_c) - 1)\ , \tag{11}$$

where $t_c = 33.3$ years. Thus, the radiation heating, $\Delta F_0(t)$, in 2025 will be 4 W^2m^{-2} which corresponds to a CO_2 concentration of 574 ppmv or the doubling of its value in 1860.

First of all, let us discuss the time delay of the mixed layer for the global warming based on Eq.(10). Let $F_0(t) = 0$ and $R_m = 0$, we have $\theta_{os} = \Delta F_0(t) / B = \theta_{ls}$ which represents the CO_2–induced global surface warming regardless of ocean effect (neither the lower ocean, nor the mixed layer ocean). As a matter of fact, θ_{ls} can be regarded as a reference temperature increase which reflects the immediate response of the land surface temperature to $\Delta F_0(t)$ without any time delay. Let $F_0(t) = 0$, which implies that only the mixed layer is taken into account, then θ_{os} can be solved from Eq.(10) directly. Thus, the "realistic" global warming should be

$$\theta_s = f \cdot \theta_{os} + (1 - f)\theta_{ls}\ , \tag{12}$$

where $f = 0.7$ is the fraction of the world ocean, and the time delay of the mixed layer could be estimated by comparing θ_s and θ_{ls}. For instance, the time for θ_{ls} and θ_s. and qs to reach 2.6℃ is 2020 and 2025 respectively, therefore, the time delay of the mixed layer for 2.6℃ global warming is about 5 years. It should be pointed out that the time delay of the mixed layer is almost independent from the phase of global warming, however, the time delay of both the mixed layer and the lower ocean is strongly dependent upon the behavior of $\Delta F_0(t)$, that will be discussed in the next section.

In order to investigate how the lower ocean exerts the influence on the time delay and its dependence upon the choice of vertical diffusivity, we have solved Eqs.(6)–(10) using numerical integration procedure similar to that in Section 2. In doing so, the lower ocean is also divided into 40 layers with interval of 100 m. Different from the calculation in Section 2, here only three constant diffusivities i.e., $k = 0.1$, 1.0 and 5.0 ($10^{-4}m^2s^{-1}$) are selected for the sake of simplicity. Table 2 illustrates results of the three integrations with different k.

Table 2. The Global Temperature Increase From 1860 to 2025

k ($10^{-4}m^2s^{-1}$)	θ_{ls} (2025)	θ_{os} (2025)	θ_s (2025)	time delay
0.1	3.05℃	1.76℃	2.5℃	6.3 years
1.0	3.05℃	1.27℃	1.9℃	15.5 years
5.0	3.05℃	0.845℃	1.52℃	22.8 years

It is clear that the magnitude of θ_{os}, which characterizes the SST increase due to the increasing CO_2, is quite sensitive to the vertical diffusivies. For example, with $k = 0.1 \times 10^{-4}$ m^2s^{-1} the SST increase is 1.76℃ in 2025, corresponding to a global warming of 2.5℃, while that with $K = 5.0 \times 10^{-4} m^2s^{-1}$ it is only 0.845℃, corresponding to a global warming of 1.5℃. In the former, the oceanic delay for 2.5℃ global warming defined above is 6.55 years, which is only 1.5 years longer than that of the mixed layer of ocean and implies that the effect of the lower ocean is neglegible, while in the latter, the oceanic delay for 1.5℃ global warming is 22.8 years, which is more than four times of that of the mixed layer. For a typical diffusivity of $1.0 \times 10^{-4} m^2s^{-1}$, the SST increase in 2025 is 1.27℃ corresponding to a global warming of 1.9℃, for which the oceanic time delay is 15.5 years.

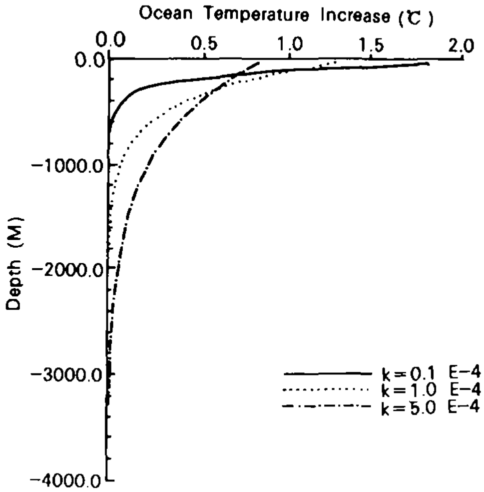

Fig.2. The ocean temperature increase from 1860 to 2025, as a function of depth, for three choices of *k*.

Fig.2 shows the ocean temperature increase from 1860 to 2025, as a function of depth, for three choices of k. It can be seen that the depth of significant temperature increase for $k=0.1\times10^{-4}\ m^2s^{-1}$ is only about 200 m, which is much shallower than the depth of about 1000 m for $k=5.0\times10^{-4}m^2s^{-1}$. Denoting that the depth of temperature increase in the lower ocean represents a measure of energy storage in it, we might say that an ocean with larger vertical diffusivity would capture more energy resulting from the increasing atmospheric CO_2 concentration and result in a longer time delay for the global warming than that with smaller one.

IV. EFFECT OF CO_2 EMISSION CHANGE ON GLOBAL WARMING

The determination of $\Delta F_0(t)$, one of the critical factors influencing the SST increase, is dependent upon the prediction of the future atmospheric CO_2 concentration. The Intergovernmental Panel of Climate Change (IPCC) assumed four possible scenarios for future atmospheric CO_2 emission (see Fig.3a). Scenario A assumes that few or no steps are taken to limit CO_2 emissions. The energy supply is coal intensive and on the demand side only modest efficiency increase are achieved. In 2100, the CO_2 concentration in the atmosphere will reach 800 ppmv, which is doubling the present one. In Scenario B, the energy supply mixing shifts towards lower carbon fuels, notably natural gas. Large efficiency increases are achieved. In Scenario C, a shift towards renewables and nuclear energy takes place in the second half of next century. For Scenario D, a shift to renewables and nuclear in the first half of the next century reduced the emissions of CO_2 , which is 50% of 1985 levels by the middle of the next century. Based on these assumptions, the radiative forcing increase, $\Delta F_0(t)$, relative to mid–18th century (1765) was predicted by IPCC and is shown in Fig.3b.

We calculated the global temperature increase from 1765 to the next century under the four scenarios using the model described in Section.3 with $k=k(z)$ given by Eq.(1) and $k=1.0\times10^{-4}m^2s^{-1}$ respectively. Table 3 lists the calculated global temperature increase in 2031 under the different scenarios and diffusivities.

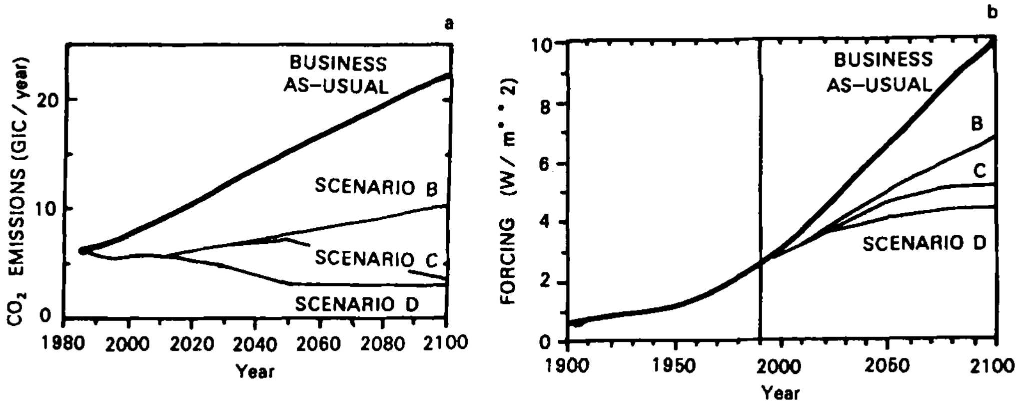

Fig.3. (a) Emissions of carbon dioxide to the year 2100, in the four scenarios developed by IPCC Working Group III (from IPCC, 1990). (b) Increase in radiative forcing since the mid−18th century, and predicted to result from the four IPCC emissions scenarios (from IPCC, 1990).

Table 3. Global Temperature Increase in 2031 with Two Choices of Diffusivity in the Four Scenarios

Scenario	θ_{ls}	θ_s $(k = 1.0 \times 10^{-4} m^2 s^{-1})$	time delay	θ_s $(k = k(z))$	time delay
A	2.49℃	1.75℃	25 years	2.0℃	15 years
B	1.91℃	1.41℃	31 years	1.61℃	18 years
C	1.89℃	1.39℃	34 years	1.59℃	18 years
D	1.75℃	1.32℃	70 years	1.51℃	27 years

Clearly from Table 3, θ_s is about 0.3℃ (for $k = k(z)$) or 0.5℃ (for $k = 1 \cdot e^{-4}$ (m^2s^{-1})) lower than θ_{ls} for all the scenarios and the time delay can vary from 15 years to 70 years for various combinations of the scenarios and the diffusivities. This means that it is difficulty to precisely predict the global warming without knowledge about the ocean diffusivity and the CO_2 emission in the future. Comparing time dependent behaviour of $\Delta F_0(t)$ in scenario A with that in scenario D, shown in Fig.3b, we can find a large difference between them. In the former, $\Delta F_0(t)$ keeps up a rapid increase in the next century, which implies that the energy arising from increasing radiative forcing will pour into the ocean speedily, so that it will take a relatively short time for θ_s. to approach θ_{ls}. Thus the time delay is about 15 years (for $k = k(z)$) for 2.49℃ increase of θ_s. Fig.4a shows that the horizontal distance between the solid line (θ_{ls}) and dashed line (θ_s), which represents the oceanic time delay, has little change with time. While in the latter, after 2025 $\Delta F_0(t)$ increases very slowly due to CO_2 emission reduction and tends towards a equilibrium. This means the energy input from the atmosphere to the ocean almost keeps a constant and θ_{os} rise slows down. Therefore it will spend longer time for θ_s to reach θ_{ls} corresponding a longer time delay. Fig.4d displays that both θ_{ls} and θ_s tend towards their equilibriums respectively. Obviously, it is not difficult to find that the time delay after 2050 is much longer than that before 2050 . In fact, if $\Delta F_0(t)$ helds constant, it will take infinite time for θ_s to reach θ_{ls}, therefore, it is not necessary to discuss time delay.

It should be pointed out that the emphasis of our paper is to study the sensitivity of the global warming event to the uncertainty of vertical diffusivity of the ocean model and different scenarios for furture atmospheric CO_2 emission rather than to predict global warming

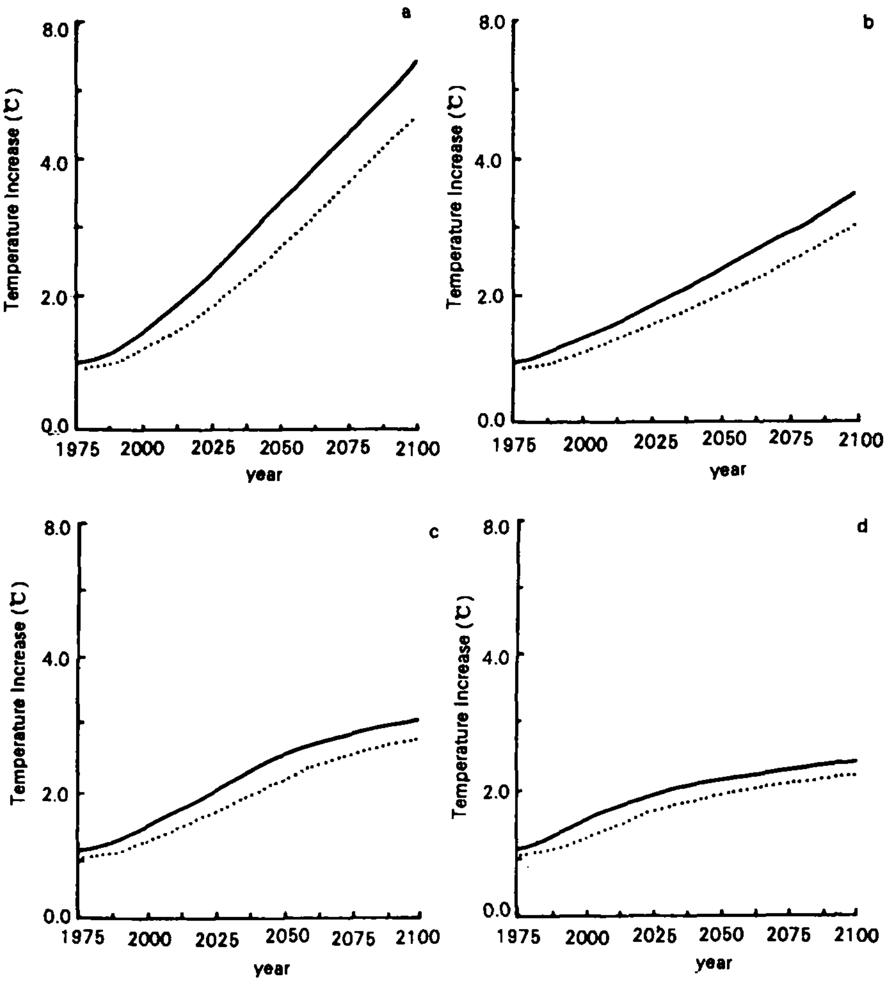

Fig.4. The changes of θ_{ls} (sold line) and θ_s (dashed line) (for $k = k(z)$). (a) Scenario A; (b) Scenario B (c) Scenario C; (d) Scenario D.

in furture. Our results suggest that the vertical diffusivity needs further studies in developing oceanic general circulation model.

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