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Research Article

Carbon cycle and climate change during the Cretaceous inferred from a biogeochemical carbon cycle model

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Abstract The carbon cycle and climate change during the Cretaceous are reconstructed by using a carbon cycle model, and discussed. The model takes into account the effects of the enhanced magma eruption and organic carbon burial rates, both of which characterize the carbon cycle during the Cretaceous. The result for the CO_2 variation is roughly consistent with the pattern of paleoclimate change inferred from the geological record. The CO_2 level during the mid-Cretaceous is estimated to be 4–5 times the present atmospheric level, corresponding to a surface temperature of 20–21°C. The warm, equable Cretaceous resulted from the effects of tectonic forcing such as enhanced CO_2 degassing, although the enhanced organic carbon burial has a tendency to decrease the CO_2 level. The organic carbon burial rate during the Cretaceous is generally larger than those for the Cenozoic, and is characterized by three major peaks (~ 1.5–1.8 times the present-day value) corresponding to the major oceanic anoxic events. In the case for the extensive mantle plume degassing, although the CO_2 levels are only 10% higher than those for the standard case during 120–100 Ma, the causes for the enhanced CO_2 levels would be quite different. If the globally averaged surface temperature had increased due to paleogeographic forcing effects, the greenhouse effect of CO_2 (and thus the CO_2 level) should be lower than the values estimated for the standard case. If the CO_2 levels are similar to, but the surface temperature is higher than, those for the standard case, either the parameter β (an influence of the Himalayas–Tibetan Plateau on the global weathering today) may be unreasonably large or the dependence of the silicate weathering rate on the CO_2 partial pressure and the surface temperature should be much weaker than those previously proposed.

Key words: carbon cycle, climate change, Cretaceous climate, mantle degassing, organic carbon burial, silicate weathering.

INTRODUCTION

A number of lines of evidence suggest that the climate during the mid-Cretaceous was very warm: the globally averaged surface temperature was probably 6–14°C higher than that of today; the latitudinal temperature gradient (equator-to-pole temperature contrast) might be 17–26°C compared to the present-day value of 41°C; there is no evidence of permanent and seasonal ice; forests and vertebrates lived near both poles; the ocean bottom water temperature was as much as 18°C; and warm anoxic oceans may have promoted organic carbon burial (Barron 1983; Frakes *et al.* 1992). Such a warm, equable Cretaceous climate could be attributed to aspects of the paleogeography such as topography, continental positions, and sea-level at that time (Barron *et al.* 1980, 1981; Barron & Washington 1982, 1984; Barron 1983), and/or to the greenhouse effect of CO_2, maintained by intensive degassing of CO_2 due to metamorphism of carbonates at subduction zones and due to rapid production rates of mid-ocean ridge basalt and oceanic plateau basalt, in addition to the small land area caused by sea-level transgression (Berner *et al.* 1983; Kasting 1984; Lasaga *et al.* 1985; Caldeira & Rampino 1991; Tajika 1998).

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Climate change during the Cretaceous has been studied by using mathematical models of the carbon cycle (Berner *et al.* 1983; Berner & Barron 1984; Kasting 1984; Lasaga *et al.* 1985; Volk 1987; Berner 1990, 1991, 1992a,b, 1994, 1997; Caldeira & Rampino 1991; Tajika 1998). The global ocean crust production rate was probably 1.8 times the present-day value between 120 and 100 Ma (Larson 1991). Therefore the mantle CO_2 degassing rate was expected to be quite large at that time (Caldeira & Rampino 1991; Tajika 1998).

In contrast, the Cretaceous is characterized by unusually widespread distribution of 'black shales' in both deep and shallow marine settings (Arthur *et al.* 1990). Black shales are sequences of variable lithology containing numerous beds with organic carbon contents of > 1 wt%. Time envelopes during which black shale deposition was particularly prevalent have been termed 'oceanic anoxic events' (OAE). There are three major periods of apparently heavier carbon isotopic composition and greater organic carbon accumulation corresponding to OAE during the Cretaceous (Arthur *et al.* 1985, 1990). A large quantity of CO_2 could have been removed from the atmosphere–ocean system as organic carbon during the OAE.

Therefore it is important to estimate the effects of these two factors (higher rates of the mantle CO_2 degassing and organic carbon burial) on the CO_2 level and carbon cycle during the Cretaceous.

Tajika (1998) constructed a model that includes the effects of these two factors, and estimated variations of CO_2 level during the last 150 million years. Relative contributions of various forcing factors to the climate change were analyzed by using high-resolution input data of the carbon isotopic composition of marine limestone and of the crustal production rates. The results were generally in good agreement with the previous estimates of paleo-CO_2 levels and paleoclimate inferred from the geological, biogeochemical, and paleontological records. In particular, it appeared that the mid-Cretaceous was very warm because of a high CO_2 level that was maintained by the enhanced CO_2 degassing rate due to the increased mantle plume activity and sea-floor spreading rate at that time, although the enhanced organic carbon burial would have decreased CO_2 level.

In the present paper, based on the study of Tajika (1998), several aspects of the carbon cycle and climate change during the Cretaceous will be examined and discussed. With geological and geochemical input data, the carbon biogeochemical cycle model used in Tajika (1998) is slightly improved and is applied to the problems concerning the characteristics of the carbon cycle during the Cretaceous, such as (i) high organic carbon burial rate; (ii) enhanced mantle plume activities; and (iii) effects of paleogeography on the warm, equable Cretaceous climate.

NUMERICAL MODELS

The carbon cycle model used in the present study is shown in Fig. 1, which is essentially the same as that of Tajika (1998). The outline of the model is described below.

Mass balance equations considered in this study are as follows.

$$\frac{dM_{AO}}{dt} = F_{D,r} + F_{D,h} + F_{D,s} + F_M^C + F_M^O + F_W^C + F_W^O - F_B^C - F_B^O = 0 \qquad (1)$$

$$\frac{dM_{sed}^C}{dt} = F_B^C - F_W^C - F_M^C - F_R^C \qquad (2)$$

$$\frac{dM_{sed}^O}{dt} = F_B^O - F_W^O - F_M^O - F_R^O \qquad (3)$$

$$\frac{dM_{ocean}^{Ca}}{dt} = F_W^S + F_W^C - F_B^C = 0 \qquad (4)$$

$$\frac{d\delta_{AO} M_{AO}}{dt} = M_{AO}\frac{d\delta_{AO}}{dt} = \delta_M(F_{D,r} + F_{D,h} + F_{D,s}) + \delta_{sed}^C(F_M^C + F_W^C) + \delta_{sed}^O(F_M^O + F_W^O) - \delta_{AO}F_B^C - (\delta_{AO} - \Delta)F_B^O \qquad (5)$$

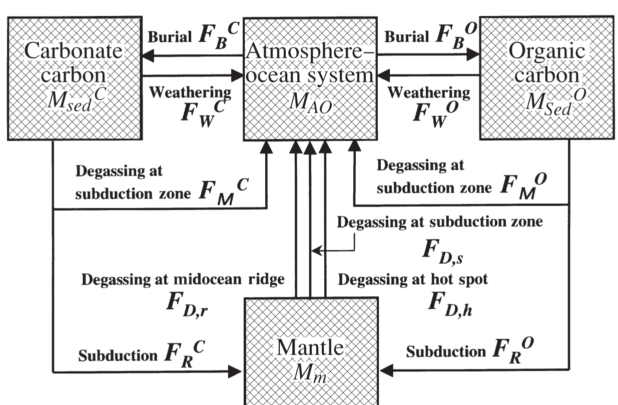

Fig. 1 The carbon cycle model considered in the present study. Boxes and arrows represent chemical reservoirs and fluxes. See text for symbols and flux expressions.

$$\frac{d\delta_{sed}^{C} M_{sed}^{C}}{dt} = \delta_{AO} F_{B}^{C} - \delta_{sed}^{C}(F_{W}^{C} + F_{M}^{C} + F_{R}^{C}) \quad (6)$$

$$\frac{d\delta_{sed}^{O} M_{sed}^{O}}{dt} = (\delta_{AO} - \Delta) F_{B}^{O} - \delta_{sed}^{O}(F_{W}^{O} + F_{M}^{O} + F_{R}^{O}) \quad (7)$$

where M_{AO} = amount of carbon in the atmosphere–ocean system, M_{sed} = amount of carbon in the crustal sediments, $M_{ocean}{}^{Ca}$ = amount of calcium ion in the ocean, $F_{D,r}$ = degassing rate of mantle CO_2 due to mid-ocean ridge volcanism, $F_{D,h}$ = degassing rate of mantle CO_2 due to hot spot volcanism, $F_{D,s}$ = degassing rate of mantle CO_2 due to subduction volcanism, F_M = degassing rate of CO_2 due to metamorphism-volcanism at subduction zone, F_W = weathering rate, F_B = burial rate, F_R = regassing rate, $\delta = \delta^{13}C$ value $(=[\{(^{13}C/^{12}C)_{sample}/(^{13}C/^{12}C)_{standard})\} - 1] \times 1000(‰))$ and Δ = carbon isotope fractionation factor through the photosynthetic process. Subscripts AO and M represent the atmosphere–ocean system and the mantle, respectively. Superscripts C, O, and S represent carbonate carbon, organic carbon, and silicate, respectively. Carbon in the atmosphere–ocean system (M_{AO}) is assumed to be in a steady state because the residence time of this quantity is short ($<10^6$ years) compared with the time scale considered in the present study. The amount of calcium ion in the ocean (M_{ocean}^{Ca}) is also assumed to be in a steady state; that is, the same amount of calcium supplied to the ocean is assumed to be precipitated as carbonates (Berner 1991, 1994).

The model of Tajika (1998) is slightly modified: (i) the term $d\delta_{AO}C_{AO}/dt$ is set to be $C_{AO}d\delta_{AO}/dt$ (because dM_{AO}/dt is assumed to be zero) for accuracy and consistency of the model; and (ii) ~10% of the present-day volcanic flux of CO_2 at the subduction zone is assumed to be derived from the mantle CO_2 ($F_{D,s}$), which has a mantle carbon isotopic composition, and the other 90% is derived from metamorphism of carbonates and organic carbon ($F_M{}^C$ and $F_M{}^O$), as suggested by Sano and Williams (1996). The metamorphic CO_2 fluxes are separated by assuming a steady state in the present-day carbon cycle system. These modifications will have influences on the numerical results for some specific periods during the Cretaceous.

Mass fluxes in the aforementioned equations are expressed as follows:

$$F_{D,r}(t) = f_{SR}(t) F_{D,r}^{*} \quad (8)$$

$$F_{D,h}(t) = f_{h}(t) F_{D,h}^{*} \quad (9)$$

$$F_{D,s}(t) = f_{SR}(t) F_{D,s}^{*} \quad (10)$$

$$F_{M}^{C}(t) = k_{M}^{C} f_{SR}(t) f_{C}(t) M_{sed}^{C} \quad (11)$$

$$F_{M}^{O}(t) = k_{M}^{O} f_{SR}(t) M_{sed}^{O} \quad (12)$$

$$F_{R}^{C}(t) = k_{R}^{C} f_{SR}(t) f_{C}(t) M_{sed}^{C} \quad (13)$$

$$F_{R}^{O}(t) = k_{R}^{O} f_{SR}(t) M_{sed}^{O} \quad (14)$$

$$F_{W}^{S}(t) = (1-\beta) f_{B}(t) f_{E}(t) f_{D}(t)^{0.65} F_{W}^{S*} \quad (15)$$

$$F_{W}^{C}(t) = k_{W}^{C} f_{BB}(t) f_{LA}(t) f_{E}(t) f_{D}(t) M_{sed}^{C}(t) \quad (16)$$

$$F_{W}^{O}(t) = k_{W}^{O} f_{D}(t) M_{sed}^{O}(t) \quad (17)$$

where f_{SR} = sea-floor spreading rate, f_h = production rate of oceanic plateau basalt, f_C = precipitation factor (dependence of degassing rate on the relative proportions of carbonates on shallow platforms and in the deep sea), f_E = soil biological activity factor, f_D = river runoff factor due to changes in paleogeography, f_{LA} = carbonate land area factor, f_B = feedback function for silicate weathering, and f_{BB} = feedback function for carbonate weathering. Superscript * represents the present-day value. All these functions are normalized to the present-day values (i.e. $fi = 1$ at present). Expressions of functions fi are basically adopted from Berner (1991, 1994). Coefficients and other constants are: $F_{D,m}{}^* = 2 \times 10^{18}$ mol/Ma; $F_{D,h}{}^* = 0.17 \times 10^{18}$ mol/Ma; $F_{D,s}{}^* = 0.58 \times 10^{18}$ mol/Ma; $F_W{}^{S*} = 6.7 \times 10^{18}$ mol/Ma; $k_W{}^C = 0.0026\ Ma^{-1}$; $k_W{}^O = 0.0030$ mol/Ma; $k_M{}^C = 9.939 \times 10^{-4}\ Ma^{-1}$; $k_M{}^O = 1.868 \times 10^{-4}\ Ma^{-1}$; $k_R{}^C = 2.9316 \times 10^{-4}\ Ma^{-1}$; $k_R{}^O = 5.6336 \times 10^{-4}\ Ma^{-1}$; $M_{sed}{}^C(0) = 5000 \times 10^{18}$ mol; $M_{sed}{}^O(0) = 1250 \times 10^{18}$ mol; $\delta_{AO}(0) = 1.0‰$; $\delta^C{}_{sed}(0) = 1.5‰$; $\delta^O{}_{sed}(0) = -23.5‰$; $\delta_M(t) = -5.0‰$; $\Delta = 25‰$ (Berner 1987, 1991, 1994; Marty & Jambon 1987; Schidlowski 1988).

A parameter β represents the contribution of the Ca dissolution from the eroded Himalayas–Tibetan Plateau materials into the ocean to the worldwide total supply of Ca due to the global chemical weathering today (Tajika 1998). Formation and uplift of the Himalayas–Tibetan Plateau is considered to have affected the global weathering rate since 40 Ma. Although the calculation will be done only for the last 150–60 million years in the present study, the parameter β should be considered because the global weathering rate before 40 Ma is determined from this parameter and the

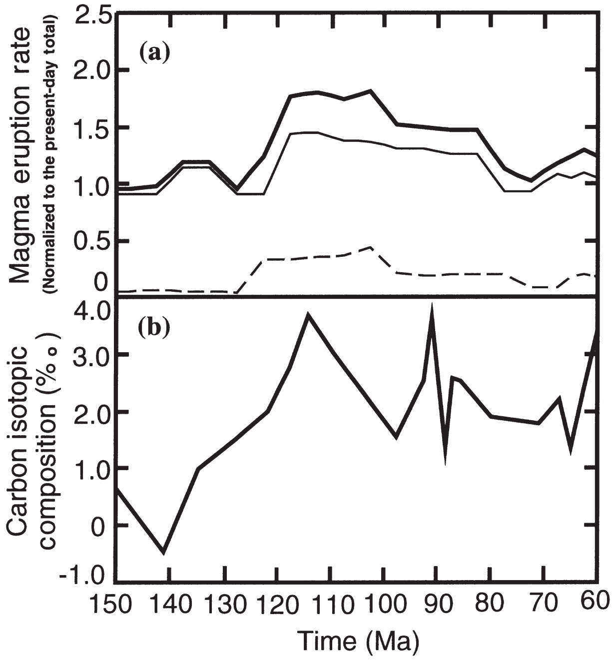

Fig. 2 Model input data used in the present study. (a) Magma eruption rates (Larson 1991). (▬), total (except Tethys); (—), mid-ocean ridges; (- - -), oceanic plateaus. (b) Carbon isotopic composition of marine carbonate (Arthur *et al.* 1985).

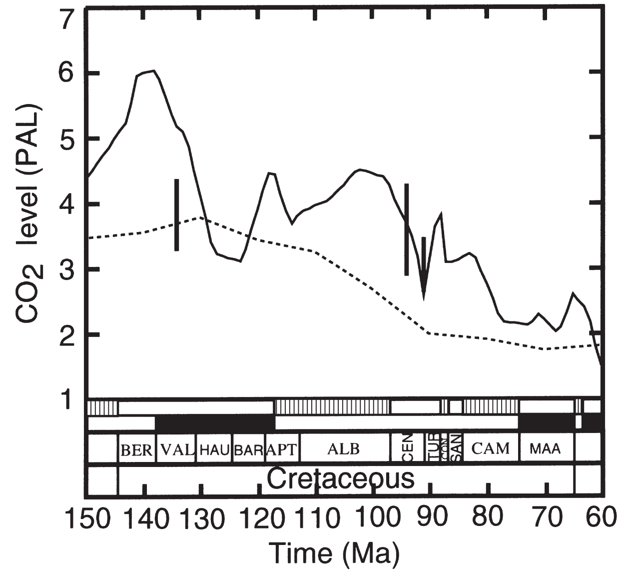

Fig. 3 Result of the temporal variation of CO_2 level (PAL, present atmospheric level) for (—) the standard case. The result of (- - -) Berner (1994) is also shown for comparison. Vertical bars are biogeochemical estimates of paleo-CO_2 level (Freeman & Hays 1992). Paleoclimate (▥/■ warm/cool) indicated from geological records are also shown (Frakes *et al.* 1992).

present-day value of global weathering rate. We adopt $\beta = 0.45$ for the standard case in the present study (Tajika 1998).

Advent and diversification of angiosperm–deciduous ecosystem during the Early–middle Cretaceous (Taylor & Hickey 1990) could have affected the global weathering rate (Knoll & James 1987; Volk 1989; Berner 1991, 1994). To reflect this, the soil biological activity factor (f_E) is assumed to increase from 0.75 at 130 Ma to 1.0 at 80 Ma (Berner 1991, 1994).

We adopt estimates of production rates of the mid-oceanic ridge basalt and the oceanic plateau basalt (Larson 1991), and the carbon isotopic composition of pelagic marine limestone (Arthur *et al.* 1985) as input data to the model (f_{SR}, f_h, and δ_{AO}; Fig. 2). The fractionation factor Δ is set to be 25‰ over the Cretaceous.

In order to estimate the global mean surface temperature by greenhouse effect of atmospheric CO_2 and H_2O, the formulation of Caldeira and Kasting (1992), which was proposed by compiling the results of one-dimensional radiative–convective climate models, is used in the present study. However, another simple formulation will be adopted to discuss the influences of paleogeographic forcing factors on the climate in the later section.

Calculations are performed by using a forward scheme. Initial conditions are given by the standard results of Tajika (1998).

RESULTS FOR STANDARD CASE

At first, we summarize the numerical results of the standard case for the climate change during the Cretaceous. These results are essentially the same as those obtained by Tajika (1998). Figure 3 shows the temporal variations of the atmospheric CO_2 level during the Cretaceous period. The result of the previous study (Berner 1994) is also shown for comparison. It is noted that Berner (1994) studied the carbon cycle during the Phanerozoic by using the model which considered the organic carbon subcycle and the CO_2 degassing as due to metamorphism–volcanism at the subduction zone, but the resolution of input data was lower and the mantle CO_2 degassing processes were not considered explicitly (the total carbon outgassing flux is, however, consistently expressed by the process of CO_2 degassing due to metamorphism–volcanism). It is also noted that Caldeira and Rampino (1991) studied the carbon cycle during the Cretaceous by using the model which considered the mantle CO_2 degassing with the same input data adopted in the present study (Larson 1991), but they did not consider the organic carbon subcycle.

In Fig. 3, the biogeochemical estimates of paleo-CO_2 levels by Freeman and Hays (1992) are also shown. They estimated paleo-CO_2 partial pressures by a correlation between the fractionation of carbon isotope during photosynthetic fixation of CO_2 from sedimentary porphyrins, and concentrations of dissolved CO_2 estimated from chemical equilibria in the dissolved inorganic carbon system.

The short-term variations of CO_2 level will result in the short-term climate change during the Cretaceous. Figure 3 also shows the climate (warming and cooling) in each age, suggested from the geological record (i.e. oxygen isotope data, various fossil floras and faunas, and sedimentary facies) (Frakes *et al.* 1992). As shown in Fig. 3, the result of the variations of CO_2 level (hence the surface temperature) is approximately consistent with the pattern of paleoclimate change inferred from the geological record. The CO_2 level at the mid-Cretaceous is estimated to be ~4–5 times the present atmospheric level (PAL), corresponding to a surface temperature of ~20–21°C.

Figure 4 shows the results of the cases where specific forcings to the carbon cycle system are given to understand the climate change during the Cretaceous. The 'biological forcing case' is defined here as $f_{SR}=f_h=f_D=f_{LA}=1$ (all the tectonic forcing factors are constant at the present-day value). Similarly, the 'tectonic forcing case' is defined as $f_C=f_E=1$, $\delta_{AO}=1‰$, and $\Delta=21.8‰$ (all the biological forcing factors are constant at the present-day value). As shown in Fig. 4, the tectonic forcing generally results in warming, whereas the biological forcing results in cooling during the Cretaceous. For example, during the mid-Cretaceous, the biological forcing decreased the CO_2 level by 20–70% and the tectonic forcing increased it by 80–90% relative to the CO_2 level for the standard case. The net result (= the standard case) is an intermediate, but nevertheless the mid-Cretaceous is very warm because of the strong effects of the tectonic forcing at that time. The solar forcing (long-term increase in the solar luminosity) has little effect on the CO_2 level because the solar luminosity was only 1.0% lower at 120 Ma than it is today. These results suggest that the warm, equable Cretaceous resulted from tectonic forcing such as the enhanced CO_2 degassing rates from the mantle, and at subduction zone due to the increased magma eruption and sea-floor spreading rates at that time, although the enhanced organic carbon burial rate may have decreased the CO_2 level to some extent from the level expected for the tectonic forcing case.

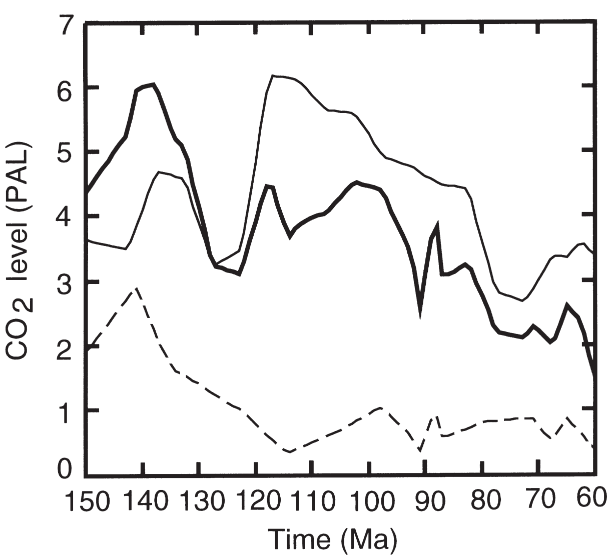

Fig. 4 The results of the temporal variation of CO_2 level for the cases of (- - -) biological forcing ($f_{SR}=f_h=f_D=f_{LA}=1$) and (—) tectonic forcing ($f_C=f_E=1$, $\delta_{AO}=1‰$, $\Delta=21.8‰$). (**—**), standard case.

ORGANIC CARBON BURIAL

The mid-Cretaceous sediments are characterized by the occurrence of thick sequences of organic carbon-rich shales called black shales. There are three major episodes of OAE during the Cretaceous (Arthur *et al.* 1985, 1990); OAE 1 (Aptian–Albian), OAE 2 (Cenomanian–Turonian), and OAE 3 (Coniacian–early Campanian). Large quantities of CO_2 could have been removed from the atmosphere–ocean system as the form of organic carbon during these periods, although the reason why the organic carbon burial rate was so large has been a matter of debate (e.g. slower rate of deep-water formation and consequent deep-water anoxia, unusually high primary productivity in surface water, or an increased organic carbon flux of terrestrial higher plant materials).

In order to compare the organic carbon burial rate for the Cretaceous with that for the Cenozoic, Fig. 5 shows the estimate of organic carbon burial rate during the last 150 Ma obtained from the present study for the Cretaceous, and from the model of Tajika (1998) for the Cenozoic. Organic carbon burial rates for the Cretaceous seem to be generally larger than those for the Cenozoic: averaged values are 6.1×10^{12} mol/Ma for the Cenozoic (0–65 Ma), 6.6×10^{12} mol/Ma for the Cretaceous (65–145 Ma), and, in particular, 7.7×10^{12} mol/Ma

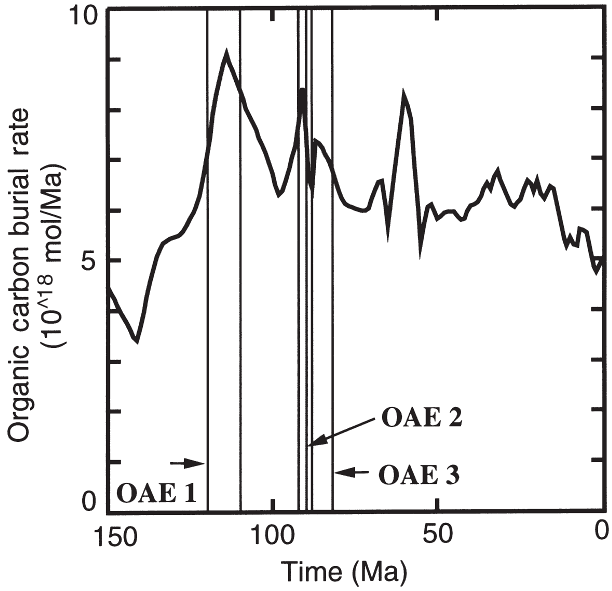

Fig. 5 The result of the temporal variation of organic carbon burial rate during the last 150 Ma. The results are estimated from the present study for the Cretaceous and from the model of Tajika (1998) for the Cenozoic. Major oceanic anoxic events (OAE) are also shown.

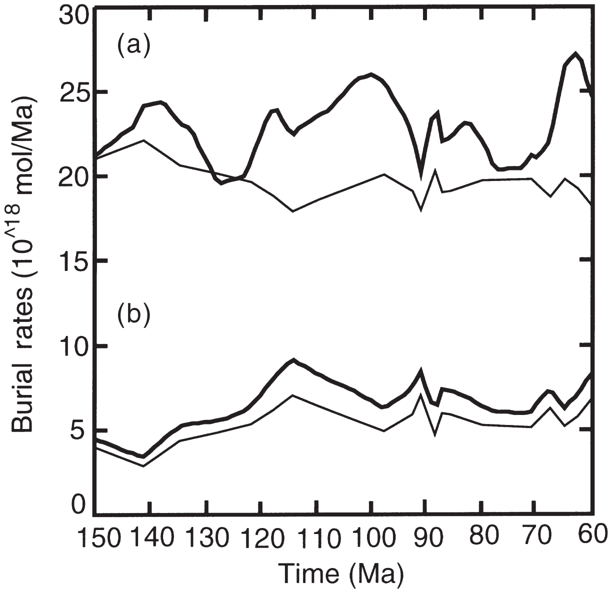

Fig. 6 The results of the temporal variations of (a) carbonate carbon burial rate and (b) organic carbon burial rate estimated from (━) present study and (—) a simple carbon isotope mass balance model (see text).

for the mid-Cretaceous (80–120 Ma). The result of the organic carbon burial rate for the Cretaceous is characterized by the major three peaks corresponding to the major OAE (Fig. 5). During the major OAE, the organic carbon burial rates would have been 1.5–1.8 times the present-day value.

As shown in Fig. 5, there is another episode of quite a large burial rate at the late Paleocene (*ca* 60 Ma). According to Shackleton and Hall (1984), the carbon isotopic composition of marine carbonates is estimated to be 3.37–3.94‰ at 59–60 Ma. This Paleocene event is recorded in the ^{13}C content of both planktonic and benthic foraminiferal tests, and it is suggested that at the time when δ^{13}C values were the most positive, the surface-to-deep water δ^{13}C gradient was at its maximum (Shackleton & Hall 1984). Such a heavy carbon isotopic composition implies the removal of light carbon isotopes from the surface ocean due to burial of a large quantity of organic carbon. Shackleton (1987) suggested that this event may be partly analogous to similar events associated with the time of black shale deposition during the Cretaceous.

The organic carbon burial rate during the Cretaceous is also estimated by Arthur *et al.* (1985) (it is noted that the carbon isotopic composition data proposed by Arthur *et al.* (1985) is used as the model input data in the present study). Arthur *et al.* (1990) describe the method used by Arthur *et al.* (1985) as 'a steady-state carbon-flux model assuming that the carbon flux from rivers and its carbon isotopic composition remained constant at present levels (44×10^{14} g/year; δ^{13}C = –5‰) and with variable δ^{13}C of pelagic carbonate'. The result of such a simple carbon isotope mass balance model is recalculated and compared with the result of the present study (Fig. 6). As shown in Fig. 6, these two estimates seem to be quite similar to each other except for their absolute values (it is noted that the organic carbon burial rate at present is the same for both cases in this figure, although Arthur *et al.* (1985) used a much higher present-day value and estimated generally higher values than those obtained here). The results of the present study are ~20% larger than that of the simple model. This implies that the pattern of variation of the organic carbon burial rate will strongly depend on the model input data of carbon isotopic composition, although the absolute value may depend on other factors.

In contrast to the organic carbon burial rate, the estimates of the carbonate carbon burial rate by these two methods are quite different, as shown in Fig. 6. The result of a simple model clearly reflects the input data of carbon isotopic composition. Because river inputs of carbon and calcium to the ocean due to chemical weathering must largely depend on the CO_2 level and the global surface temperature, both the pattern and the absolute values of the carbonate carbon burial rate should be different between the results of two different methods (Fig. 6). In fact, the result of the carbonate carbon burial rate during the Cretaceous, estimated from the full model, seems to be generally larger than it is today because of the higher CO_2 level and the higher surface temperature for that period.

From these results, it appears that a simple carbon isotope mass balance model would be useful for estimating the pattern of the organic carbon burial rate, but may not be appropriate for its absolute value, and cannot estimate the carbonate carbon burial rate.

MANTLE PLUME DEGASSING CASE

The mid-Cretaceous is period of large sea-floor spreading rate and mantle plume activity. Larson (1991) estimated a 50–75% increase in the ocean crust formation rate during 100–120 Ma compared with the present-day value. According to Larson (1991), the magma eruption rate due to mid-ocean ridge volcanism was ~50% larger, and that due to mantle plume activity was ~six times the present-day value. These estimates imply that the degassing rate of CO_2 from the mantle could have also increased during the mid-Cretaceous. Assuming the same carbon content in the source materials of magma for the upper mantle and for the mantle plume, the CO_2 degassing rate from the mantle plume is ~1/10 of that at mid-ocean ridge at present. Figure 7 shows the results of system analysis for the standard case of the carbon cycle model. As expected from the magma eruption rates, effects of mantle CO_2 degassing are large during the mid-Cretaceous; the 15–20% of CO_2 level is contributed by CO_2 degassing at mid-ocean ridge and hot spot. Degassing of CO_2 due only to the eruption of oceanic plateau basalt at that time increases the CO_2 level by 5–10% for the same period. The enhanced total CO_2 degassing rate due to the enhanced magma eruption and sea-floor spreading rates contributes to the elevated atmospheric CO_2 level by 35–60% at that time.

Recently, Marty and Tolstikhin (1998) proposed that the upper limit for the CO_2 flux from the mantle plume is estimated to be 3×10^{18} mol/Ma, which is comparable to, or even larger than, that from mid-ocean ridge (they estimated it to be $(1.8\pm0.7)\times10^{18}$ mol/Ma). This result is based on time-averaged magma production rates, estimated contributions of geochemical sources to plume magmatism, and estimates for ^{3}He content and

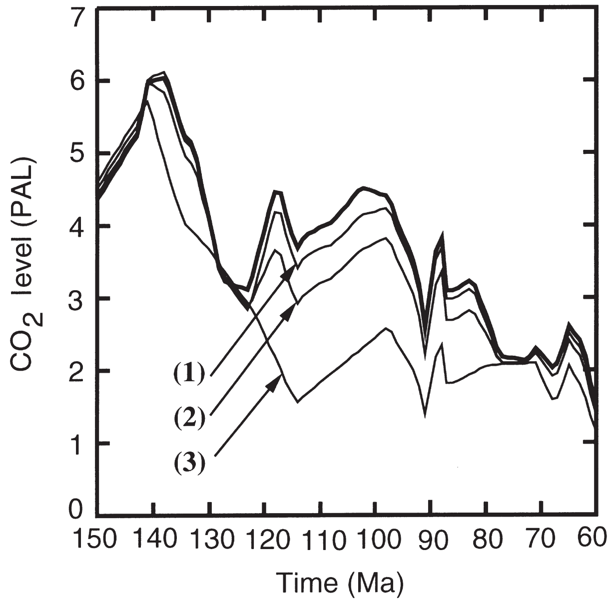

Fig. 7 The results of the temporal variations of CO_2 level for (▬) the standard case. (1) $F_{D,h}$ is constant at the present-day value. (2) $F_{D,r}$ and $F_{D,h}$ are constant at the present-day values. (3) $F_{D,r}$, $F_{D,h}$, $F_{D,s}$, F_M^C and F_M^O are constant at the present-day values. PAL, present atmospheric level.

C/^{3}He ratio of the mantle plume (Marty & Tolstikhin 1998). It is noted that the CO_2 degassing rate for the mantle plume ($F_{D,h}$) of 0.17×10^{18} mol/Ma is used in the standard case in the present study. If their estimates were the case, the CO_2 degassing rate during the mid-Cretaceous could have been much larger, and the climate could have been much warmer than that estimated for the standard case in the present study. We therefore examine CO_2 levels during the Cretaceous in the case for a large CO_2 degassing rate from mantle plume as an extreme case (referred to as 'the plume degassing case').

Assuming a steady state in the carbon cycle system today, the total degassing rate of CO_2, F_{input} ($=F_{D,r}+F_{D,h}+F_{D,s}+F_M^C+F_M^O$), should be equal to 7.95×10^{18} mol/Ma ($=F_W^S+F_B^O-F_W^O$) from equations (1) and (4). The assumption of steady state will be reasonable because the residence time of carbon in the atmosphere–ocean system is very short. In order to estimate the maximum effect of CO_2 degassing from the mantle plume on the mid-Cretaceous climate, mantle CO_2 degassing rates are assumed to be 3.0×10^{18} mol/Ma from the mantle plume, 1.3×10^{18} mol/Ma from the mid-ocean ridge, and zero from the subduction zone at present (others are metamorphic CO_2 outgassing fluxes from subduction zone).

Figure 8 shows the results of system analysis for the plume degassing case of the carbon cycle model.

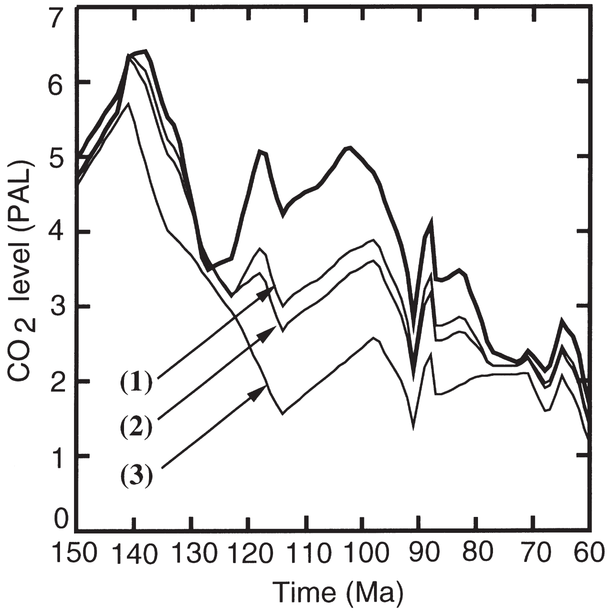

Fig. 8 The results of the temporal variations of CO_2 level for (▬) the mantle plume degassing case (see text). (1) $F_{D,h}$ is constant at the present-day value. (2) $F_{D,r}$ and $F_{D,h}$ are constant at the present-day values. (3) $F_{D,r}$, $F_{D,h}$, $F_{D,s}$, F_M^C and F_M^O are constant at the present-day values. PAL, present atmospheric level.

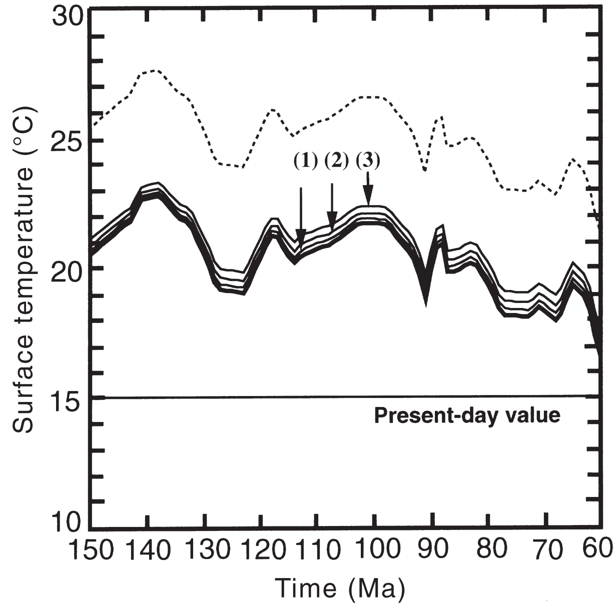

Fig. 9 The result of the temporal variation of the globally averaged surface temperature. (▬), standard case (T_S^{std}); (---), $T_S^{std}(t)$ + 4.8°C. Others are the cases for ΔT_{pg} to be (1) 1.5°C; (2) 3.0°C; and (3) 4.8°C.

In this case, ~ 10% higher CO_2 levels compared with the standard case are obtained during 120–100 Ma. The CO_2 degassing at mid-ocean ridge and hot spot contribute 30–35% relative to the CO_2 level during this period. As reflected in the large CO_2 degassing rate from mantle plume, degassing of CO_2 due only to the eruption of oceanic plateau basalt at that time increases the CO_2 level by 25–30% for the same period. It is, however, noted that the enhanced total CO_2 degassing rate due to the enhanced magma eruption and sea-floor spreading rates is contributed by the elevated atmospheric CO_2 level by 35–60% at that time, which is almost the same as that for the standard case.

As a result, in the mantle plume degassing case, CO_2 levels predicted for the mid-Cretaceous would increase only slightly from that estimated for the standard case because of the constraint on the carbon cycle system to be in a steady state at present. However, the causes for enhanced CO_2 levels during this mid-Cretaceous period would be quite different from those expected for the standard case.

OTHER CLIMATIC FORCING FACTORS

A warm, equable Cretaceous is originally deduced from the extensive latitudinal distribution of floras, faunas and sedimentary indicators that are presently restricted to lower latitudes. Barron (1983) summarized geological, geochemical, and paleontological data, and concluded that the equator-to-pole temperature contrast was 17–26°C compared with 41°C at present, and the globally averaged surface temperature during the mid-Cretaceous would be 6–14°C warmer than that at present, although it may have been much cooler than previously believed (Sellwood *et al.* 1994). Figure 9 shows the result of temperature variation estimated from the carbon cycle model in the present study. During the mid-Cretaceous, the globally averaged surface temperature is estimated to be 5–6°C warmer than that at present, which is consistent with the lower estimate of paleotemperature. Because the estimate of surface temperature in the present study is only from greenhouse effects of atmospheric CO_2 and water vapor, and also because there may be other climatic forcing factors which could have contributed to the warm climate, a higher estimate for paleotemperature may be explained by a combination of the effects of several climatic forcing factors.

Barron and Washington (1984) investigated the role of paleogeography in the climate for the mid-Cretaceous through a series of sensitivity studies by using a general circulation model (GCM). The

Cretaceous is a large geographical contrast from the present, with substantially different continental positions and elevations and with a 20% reduced land area due to high global sea-level (Barron & Washington 1984). They demonstrated that paleogeography was a substantial climatic forcing factor, which could explain the 4.8°C increase in globally averaged surface temperature compared with a present-day control case. Then, combined with the greenhouse effect estimated in the present study, an ~10°C increase in surface temperature might be expected (Fig. 9).

It is, however, noted that the surface temperature should be determined as a consequence of a steady state in the carbon cycle system. Therefore, the surface temperature would not change greatly from that estimated in the present study, even when other climatic forcing factors are considered. When other climatic forcing factors (e.g. paleogeography) are added to the model, the surface temperature, T_S, will be expressed as follows:

$$T_S = T_{eff} + \Delta T_{gh} + \Delta T_{pg} \tag{18}$$

where T_{eff} represents the effective temperature of the Earth (~255°K). ΔT_{gh} and ΔT_{pg} represent a temperature increase due to the greenhouse effect and due to the paleogeography effect, respectively. As described earlier, the relation

$$F_W{}^S = F_{input} + F_W{}^O - F_B{}^O \tag{19}$$

is obtained from the mass balance equations. If we assume that the fluxes appearing in the right-hand side of this equation F_{input}, $F_W{}^O$, and $F_B{}^O$ do not depend on the surface temperature, the silicate weathering rate ($F_W{}^S$) under the specific condition at the specific period should be determined uniquely in the carbon cycle system. Then, the feedback function of silicate weathering rate $f_B(P_{CO_2}, T_S)$, which represents the dependence of the silicate weathering rate on the CO_2 partial pressure and the surface temperature, should be determined. Therefore, the CO_2 partial pressure and the surface temperature should also be determined. As a result, if the globally averaged surface temperature could have increased at some extent by paleogeographic forcing, the greenhouse effect of CO_2, and thus the CO_2 level, should be lower than the values estimated for the standard case in the present study in order to be the same silicate weathering rate for the case without paleogeographic forcing (Figs 9,10).

Figure 10 shows the results of CO_2 variations in

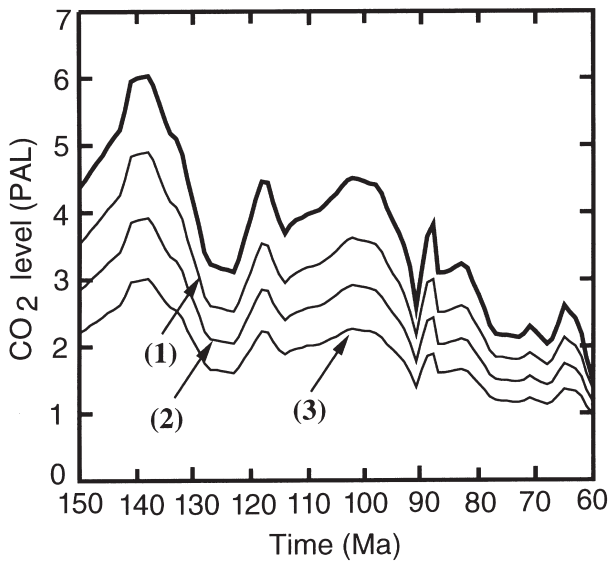

Fig. 10 The results of the temporal variations of the CO_2 levels. (—), standard case. Others are the cases for ΔT_{pg} to be (1) 1.5°C; (2) 3.0°C; and (3) 4.8°C.

the cases for different temperature increases by paleogeographic forcing. As described in the previous paragraph, CO_2 levels should be lower for a larger temperature increase due to paleogeographic forcing. However, as shown in Fig. 9, the surface temperature would not change greatly even for the maximum increase in the surface temperature due to paleogeography. Although the surface temperature may not be estimated by simple summation of ΔT_{gh} and ΔT_{pg}, the CO_2 level obtained in the present study would be an upper estimate under the specific boundary condition given in the present study, and it might be lower when there were other climatic forcing factors which may have a large effect on the surface temperature during the Cretaceous.

It is, however, noted that there are possibilities for the cases of the similar CO_2 level with the higher surface temperature. When the adjustable parameter β is given to be 0.75, variations of CO_2 level are very similar to, but the surface temperature is ~6°C higher than, those for the standard case. However, the value of $\beta = 0.75$ should be too large when we consider this parameter to be the contribution of the Ca dissolution from the eroded Himalayas–Tibetan Plateau materials into the ocean to the worldwide total supply of Ca due to the global chemical weathering today. Therefore this case seems to be unreasonable. Another possibility is that dependence of the silicate weathering rate on T_S and the CO_2 partial pressure might be much weaker than that used in the present study. In such a case, the similar CO_2 level but higher surface temperature might be possible. However, the feedback functions for the silicate weathering rate proposed so far (Walker *et al.* 1981; Volk 1987; Berner 1991, 1994) seem to be too strong to explain such a case.

CONCLUSIONS

The conclusions of this paper are summarized as follows.

(1) The variations of the atmospheric CO_2 level and carbon fluxes in the carbon cycle system during the Cretaceous are reconstructed by using high-resolution data of crustal production rates of the mid-oceanic ridge basalt and oceanic plateau basalt, and of the carbon isotopic composition of marine limestone. The model considered the enhanced magma eruption and organic carbon burial rates, both of which characterize the carbon cycle during the Cretaceous.

(2) The variations of CO_2 level (hence the surface temperature) are roughly consistent with the pattern of paleoclimate change inferred from the geological record. The CO_2 level at the mid-Cretaceous is estimated to be ~4–5 PAL, corresponding to the surface temperature of ~20–21°C.

(3) Tectonic forcing results in warming, whereas biological forcing results in cooling during the Cretaceous (that is, biological forcing decreased CO_2 level by 20–70%, and tectonic forcing increased it by 80–90% relative to the CO_2 level for the standard case). The warm, equable Cretaceous resulted from tectonic forcing such as the enhanced CO_2 degassing rates from the mantle and also at subduction zone due to the increased magma eruption and sea-floor spreading rates at that time, although the enhanced organic carbon burial rate may have a tendency to decrease CO_2 level to some extent from the level expected for the tectonic forcing case.

(4) Organic carbon burial rates for the Cretaceous seem to be generally larger than those for the Cenozoic. The result of the organic carbon burial rate for the Cretaceous is characterized by the three major peaks (~1.5–1.8 times the present-day rate) corresponding to the major OAE.

(5) It appears that a simple carbon isotope mass balance model would be useful for estimating the pattern of organic carbon burial rate, but it may not be appropriate for its absolute value, and it cannot estimate the carbonate carbon burial rate.

(6) In the plume degassing case, CO_2 levels that are ~10% higher compared with the standard case are obtained during 120–100 Ma. The CO_2 degassing at mid-ocean ridge and hot spot contributes 30–35%, and degassing due only to the eruption of oceanic plateau basalt contributes 25–30% relative to the CO_2 level during this period. The causes for enhanced CO_2 levels would be quite different from those obtained for the standard case.

(7) The surface temperature should be determined as a consequence of a steady state in the carbon cycle system. Therefore the surface temperature would not change greatly from that estimated in the present study, even when other climatic forcing factors, such as paleogeography, are considered. If the globally averaged surface temperature could have increased at some extent by paleogeographic forcing, the greenhouse effect of CO_2, and thus the CO_2 level, should be lower than the values estimated for the standard case. If the CO_2 levels are similar to, but the surface temperature should be higher than, those for the standard case in the present study, either the parameter β may be unreasonably large or the dependence of the silicate weathering rate on the CO_2 partial pressure and the surface temperature should be much weaker than those previously proposed.

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