Long-term stability of climate and global glaciations throughout the evolution of the Earth

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Earth's climate is considered to be stable on the order of $>10^6$ years, owing to a negative feedback mechanism in a carbon cycle system. However, any decrease in net input flux of CO_2 to the atmosphere-ocean system (i.e., volcanic-metamorphic CO_2 flux minus excess organic carbon burial flux) lowers the surface temperature and would eventually initiate global glaciation. The F_D–F_B^O diagram (F_D: the total CO_2 degassing flux, F_B^O: the organic carbon burial flux) is proposed as a measure of the susceptibility of the Earth to global glaciations. By using this diagram with the carbon fluxes estimated from a carbon cycle model during the Phanerozoic, the net input flux of CO_2 is found to have been very close to the critical condition for a global glaciation at the Late Paleozoic. During the Proterozoic, a carbon isotope mass balance model with this diagram shows that global glaciations occurred probably due to a decrease in the CO_2 degassing in addition to an increase in the organic carbon burial. Because the Sun becomes brighter as it evolves, the critical level of atmospheric CO_2 pressure to cause global glaciation will be lower than the critical CO_2 pressure for photosynthesis of C4 plants within 500 million years. At this point, the net input flux of CO_2 will be too large to cause global glaciations. Continuous volcanic-metamorphic activities (i.e., plate tectonics) may be one of the necessary conditions for the Earth and Earth-like planets in extrasolar planetary systems to keep liquid water and life over the timescales of planetary evolution.

Key words: Carbon cycle, global glaciation, snowball Earth, astrobiology.

1. Introduction

The climate of the Earth has been warm ($>0°C$) during most of its geological history (Fig. 1), as suggested from evidence for the continued presence of liquid water and life (Walker, 1982; Kasting, 1989). Such a long-term stability of the Earth's climate is explained by a negative feedback mechanism within the carbon cycle system (Walker *et al.*, 1981; Kasting, 1989; Kasting and Toon, 1989; Tajika and Matsui, 1990, 1992, 1993; Broecker and Sanyal, 1998). Paleomagnetic studies, however, revealed the evidence of low-latitude ice sheets during the Proterozoic (Evans *et al.*, 1997; Evans, 2000), which could be interpreted to be a result of global glaciations (Kirschvink, 1992, 2002; Hoffman *et al.*, 1998; Hoffman and Schrag, 2002).

Global glaciation may be caused by a decrease of the atmospheric CO_2 partial pressure (pCO_2) to a certain critical level (e.g., Caldeira and Kasting, 1992; Ikeda and Tajika, 1999), because the greenhouse effect of CO_2 plays a key role in keeping the Earth's climate warm (e.g., Walker *et al.*, 1981). The critical CO_2 level has been studied by using general circulation models (GCMs) combined with ice sheet and/or sea ice models under the conditions of reduced solar flux and paleocontinental distribution during the Neoproterozoic (e.g., Jenkins and Smith, 1999; Hyde *et al.*, 2000; Donnadieu *et al.*, 2004a).

There are, however, several possibilities which would decrease pCO_2 in the carbon cycle system. For example, an increase in the organic carbon burial may have decreased pCO_2, causing the global glaciations during the Neoproterozoic (Kaufman *et al.*, 1997; Hoffman *et al.*, 1998). Tajika (2004) showed that the reduction of volcanic-metamorphic activities would have been required to cause the global glaciations, in addition to the increased rate of organic carbon burial. In contrast, Donnadieu *et al.* (2004b) argued that the decrease of pCO_2 could have been caused by an increase in silicate weathering rate, owing to an increase in runoff and emplacement of flood basalt through the break-up of the supercontinent Rodinia.

It is, therefore, important to clarify the condition of the carbon cycle system to cause global glaciations. Because the susceptibility of the Earth to global glaciations may change with time, it is useful to estimate the generalized condition of the carbon cycle system to cause global glaciations over the Earth's evolution (from the past to the future). Such an attempt has never been made to date, but will be important for understanding the susceptibility of the Earth to global glaciations and also the evolution of Earth-like planets which will be found in extrasolar planetary systems in the near future.

In this study, the general condition of the carbon flux required for the initiation of global glaciation is estimated by analyzing a simple climate model with a carbon cycle system. We propose the F_D–F_B^O diagram (F_D: the total CO_2 degassing flux, F_B^O: the organic carbon burial flux)
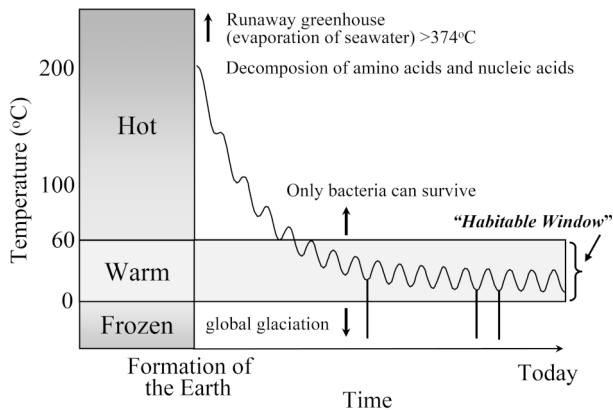
Fig. 1. Climate evolution of the Earth. The climate of the Earth may have been warm (>0°C) and suitable for life during the most of the geological history, although global glaciations have occurred at least three times during the Proterozoic (e.g., Kirschvink, 1992, 2002; Hoffman *et al.*, 1998; Hoffman and Schrag, 2002). The early Earth could have been "hot" (>60°C) if the land masses were small (e.g., Tajika and Matsui, 1990, 1992, 1993) and there were no microbial activity on the land surface (e.g., Schwartzman and Volk, 1989).
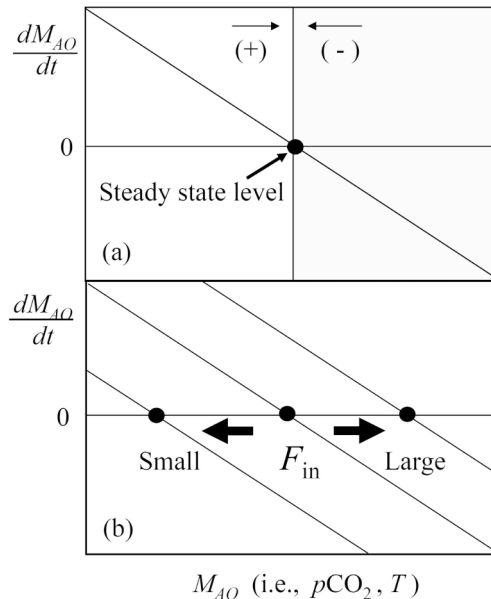

Fig. 2. Concept of the negative feedback mechanism for surface temperature of the Earth due to carbon cycle (Walker *et al.*, 1981). (a) Amount of CO_2 in the atmosphere-ocean system. M_{AO} (thus, pCO_2 or T), changes so that M_{AO} may be in a steady state. When M_{AO} is larger than the steady-state level, a warmer climate would remove extra amount of CO_2 owing to silicate weathering followed by carbonate precipitation in the ocean. When M_{AO} is smaller, CO_2 would accumulate in the atmosphere-ocean system owing to CO_2 degassing via volcanism-metamorphism. (b) The steady-state level of M_{AO} should change according to the net CO_2 input flux F_{in} into the atmosphere-ocean system. When F_{in} is large, the steady-state level becomes large (climate becomes warm). When F_{in} is small, the steady-state level becomes small (climate becomes cool).

as a measure of the susceptibility of the Earth to global glaciations. By using this diagram and carbon cycle models during the Phanerozoic and the Neoproterozoic, variations of carbon fluxes during these two geological ages are discussed in the context of global glaciations. Finally, the susceptibility of the Earth to global glaciations in its evolution is discussed.

2. Long-term Climate Stability and Global Glaciations

A mechanism to stabilize the Earth's climate on geological time scales (more than 10^6 years) resides in the carbon cycle system (Walker *et al.*, 1981; Berner *et al.*, 1983). The dominant process for the atmospheric CO_2 consumption is the weathering of silicate minerals, followed by the precipitation of carbonates in the ocean. Silicate weathering rate (F_W^S) depends strongly on temperature (T) and pCO_2 (e.g., Walker *et al.*, 1981; Schwartzman and Volk, 1989), that is,

$$F_W^S \propto f_E \cdot f_B(T, pCO_2) \tag{1}$$

where f_E and f_B represent the soil biological activity (= 1 at present) and the weathering feedback function, respectively. Mass balance of carbon in the atmosphere-ocean system is represented simply by

$$\frac{dM_{AO}}{dt} = F_{in} - F_W^S(T, pCO_2) \tag{2}$$

where M_{AO} is the amount of CO_2 in the atmosphere-ocean system, and F_{in} is the net CO_2 input flux via volcanism-metamorphism (e.g., Walker *et al.*, 1981).

Figure 2 shows the schematic behaviors of the solution of Eq. (2). We assume here that the average flux of CO_2 degassing via volcanism-metamorphism does not change on the order of 10^6 years. If the surface temperature increases for any reason (a certain perturbation or a short-term event), then CO_2 would be removed from the atmosphere-ocean system by increased rate of silicate weathering, followed

by carbonate precipitation in the ocean. Similarly, if the surface temperature decreases, CO_2 would accumulate in the atmosphere-ocean system through the decreased rate of silicate weathering. In either case, deviation from the steady state would be damped, and the system tends to maintain a steady state (Fig. 2(a)). Because of this negative feedback mechanism, the Earth's climate may have been suitable for life throughout the geological history despite variations of climate factors, such as solar luminosity (Walker *et al.*, 1981; Kasting and Toon, 1989; Tajika and Matsui, 1990).

It is, however, noted that several boundary conditions to control F_{in} and the efficiency of F_W^S should have changed with time. Such factors include volcanic-metamorphic activities, organic carbon budget (imbalance of weathering and burial of organic carbon), soil biological activity (Walker *et al.*, 1981; Berner *et al.*, 1983; Lasaga *et al.*, 1985; Schwartzman and Volk, 1989; Tajika and Matsui, 1992, 1993; Tajika, 1998, 1999; Berner, 1994).

The boundary conditions change the steady-state levels of T and pCO_2 (Fig. 2(b)). When F_{in} is large, steady-state levels become large (climate becomes warm), and climate becomes cool for small F_{in}. If F_{in} were below a certain critical level, the Earth would be globally ice-covered. Therefore, global glaciations are caused even if the system has a mechanism to stabilize the climate.

Then, in order to understand the physical aspects of global glaciation, we summarize behaviors of the climate system obtained from one-dimensional energy balance cli-
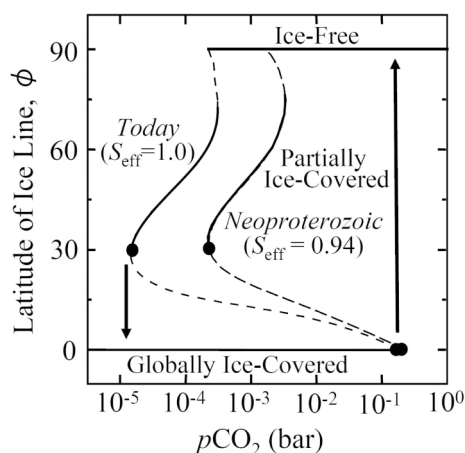
Fig. 3. The steady-state solutions of 1-D EBM for two cases of the normalized solar flux S_{eff} (1 for today and 0.94 for Neoproterozoic). Solid lines, dashed lines, and solid circles represent stable solutions, unstable solutions, and critical conditions (where stable solution disappears), respectively. Arrows represent climate jump due to climate instability for the case of $S_{eff} = 1$.
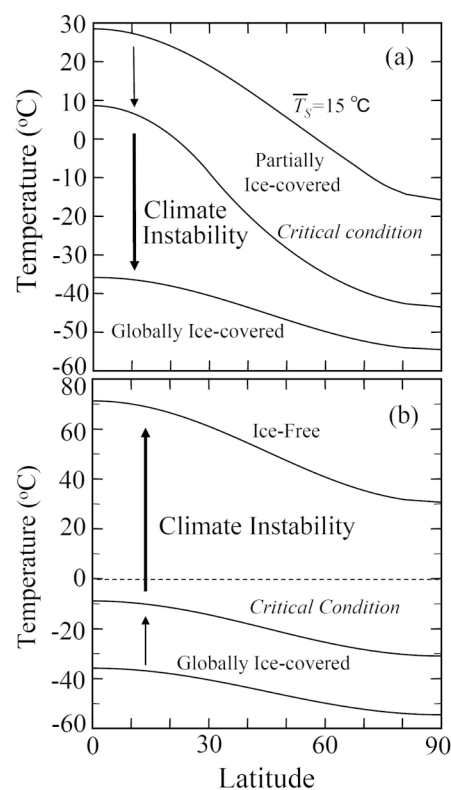

Fig. 4. Changes in the latitudinal temperature distribution through the snowball Earth event. (a) Change in the temperature distribution from a warm climate (the globally-averaged surface temperature is 15°C) to the critical condition, and to the globally ice-covered state. (b) Change in the temperature distribution from the globally ice-covered state to the critical condition, and to the ice-free state.

mate models (1-D EBMs). The 1-D EBM considers energy balance among the solar incident flux, the latitudinal heat transport, and the outgoing infrared radiation at each latitude, and ice cap with high surface albedo is assumed to extend from poles to lower latitudes (e.g., North *et al.*, 1981). When we assume the outgoing radiation as functions of pCO_2 and T (Caldeira and Kasting, 1992; Ikeda and Tajika, 1999), the surface temperature at each latitude and the latitude of the ice line (ice cap edge) against pCO_2 can be obtained (Figs. 3 and 4).

Figure 3 shows typical results for steady-state solutions of latitude of ice line against pCO_2 for two cases of the normalized solar flux (1.0 for today and 0.94 for the Neoproterozoic). It is well known that there are three kinds of stable climate states: the ice-free branch, the partially ice-covered branch, and the globally ice-covered branch (North *et al.*, 1981), as represented by solid curves in Fig. 3.

As pCO_2 decreases from the partially ice-covered branch, the ice cap extends to lower latitudes. When pCO_2 reaches a critical value and the ice line reaches $\approx 30°$, the Earth falls into the globally ice-covered branch, owing to the climate instability (North *et al.*, 1981). Figure 4 shows changes in the surface temperature distribution (a) from the partially ice-covered to the globally ice-covered branch and (b) from the globally ice-covered to the ice-free branch, obtained from the same model. The surface temperature decreases drastically through the climate instability (Fig. 4(a)), because the planetary albedo becomes very high.

The global glaciation would not continue because CO_2 is released via volcanism-metamorphism and accumulates on the order of 0.1 bar in the atmosphere in several million years on the globally glaciated Earth (Kirschvink, 1992; Hoffman *et al.*, 1998). Then, the globally ice-covered branch disappears, and the climate shifts to the ice-free branch (Figs. 3 and 4(b)).

In these respects, we conclude that the concept of the long-term climate stability and the occurrence of global

glaciations do not contradict with each other.

3. Critical Condition of CO_2 Flux

We now estimate the critical condition of the CO_2 flux in the carbon cycle system to cause the climate instability described in the previous chapter.

When we consider more detailed mass balance of carbon in the atmosphere-ocean system than Eq. (2), we should describe the net CO_2 input flux to the atmosphere-ocean system, F_{in}, as follows (see Tajika (2003)):

$$F_{in} = F_D + F_W^O - F_B^O \qquad (3)$$

where F_D is the total degassing rate of CO_2 via volcanism-metamorphism, F_W^O is the weathering rate of organic carbon, and F_B^O is the burial rate of organic carbon. Note that we regarded $F_{in} \approx F_D$ (i.e., $F_W^O \approx F_B^O$) in the previous chapter.

For the time scale of more than 10^5 years, we can assume that the atmosphere-ocean system is in a steady state. Then, Eq. (2) is rewritten by

$$F_{in} = F_W^S(T, pCO_2) \qquad (4)$$

The critical value of F_{in} to cause the climate instability may be estimated from Eq. (4) with the pCO_2 level and the latitudinal temperature distribution at the critical condition (Fig. 4(a)) by integrating the weathering rate from the equator to the ice line as a function of temperature at
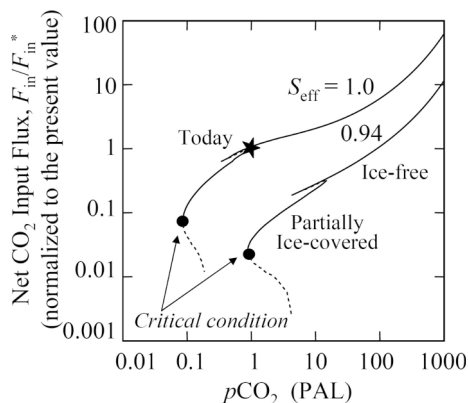
Fig. 5. The net CO_2 input flux (normalized to the present rate) and the atmospheric CO_2 level (in PAL = present atmospheric level) for the normalized solar flux S_{eff} = 1.0 (today) and 0.94 (Neoproterozoic). The critical condition is represented by a solid circle for each case.
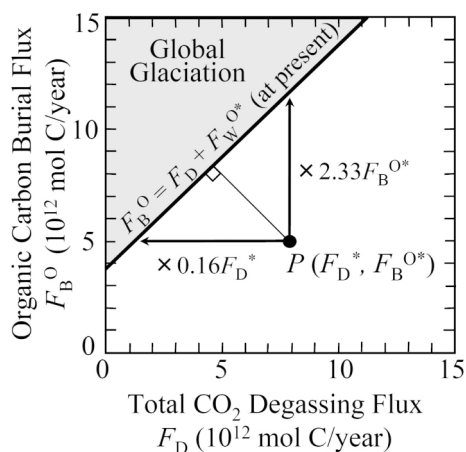

Fig. 6. The F_D–F_B^O diagram as a measure of the susceptibility of the Earth to global glaciations. The thick solid line represents the critical condition for the carbon fluxes to cause a climate instability to the globally ice-covered state. The solid circle P represents the flux condition today. Superscript $*$ represents the present-day value.
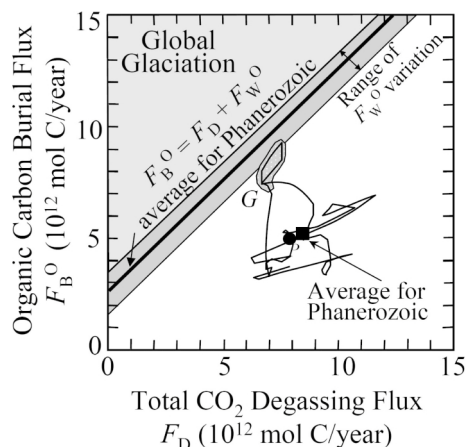

Fig. 7. Trajectory of temporal variation of CO_2 fluxes during the Phanerozoic estimated from a carbon cycle model (GEOCARB II: Berner, 1994). Hatched region G represents the conditions for the Late Paleozoic glaciation at around 300 Ma. Lines, solid circle, and solid square represent the critical condition, the flux condition today, and the average flux condition obtained from averaging the calculated results throughout the Phanerozoic, respectively. Note that the average condition during the Phanerozoic is very close to the condition today.

each latitude. Figure 5 shows the relation between F_{in} and the steady-state level of pCO_2. The critical value of F_{in} is estimated to be $\approx 5.0 \times 10^{11}$ mol/year at present and $\approx 1.5 \times 10^{11}$ mol/year at 750 million years ago (Ma), which are less than 1/10 and 1/40 of the present-day flux ($\approx 6.65 \times 10^{12}$ mol/year; Berner, 1994), respectively. The critical F_{in} should be such low values because F_W^S becomes very low under the climate at the critical condition.

We therefore approximate $F_{in}(= F_W^S) \approx 0$ at the critical condition. This approximation provides the *lower limit* of the critical condition regardless of the choice of models and model parameters, and does not depend on the precise value of the critical pCO_2 because the approximation $F_W^S \approx 0$ corresponds to the assumption $T < 0°C$ even at the equator. As a result, from Eq. (3), the critical condition is represented by

$$F_B^O = F_D + F_W^O. \qquad (5)$$

This relation can be expressed by a line in a F_D–F_B^O space, as shown in Fig. 6 (hereafter we call this "the F_D–F_B^O diagram").

The susceptibility of the Earth to global glaciations is measured by a distance between this line and the CO_2 flux condition. For example, the present CO_2 flux condition (F_D^* and F_B^{O*} where $*$ represents the present-day value) is represented by the solid circle P in Fig. 6. Then, in order to cause global glaciation, F_D should become $<1.25 \times 10^{12}$ mol/year ($= F_D^* - F_{in}^*$, or <0.16-fold the present-day value) when F_B^O and F_W^O remain constant. Similarly, F_B^O should become $>11.65 \times 10^{12}$ mol/year ($= F_B^{O*} + F_{in}^*$, or >2.33-fold the present-day value) when F_D and F_W^O remain constant. Note that the minimum distance between the line and P does not correspond to F_{in}^*, but to $F_{in}^*/\sqrt{2}$ in this diagram. This means that simultaneous variations of F_D and F_B^O affect the climate change effectively.

In the following, we evaluate the CO_2 flux condition in the Phanerozoic and the Neoproterozoic, respectively, and discuss the susceptibility of the Earth to global glaciations in these ages.

3.1 Phanerozoic

In order to evaluate the condition for CO_2 fluxes during the Phanerozoic, we adopt a carbon cycle model called GEOCARB (Berner, 1994). This model considers carbon mass balance and carbon isotope mass balance by parameterizing each geochemical process as functions of various factors, such as the seafloor spreading rate. Carbon isotopic composition of marine carbonate is used as boundary conditions. We recalculate the carbon cycle during the Phanerozoic with the GEOCARB model to evaluate the CO_2 flux condition during this time interval by plotting the results on the F_D–F_B^O diagram.

Figure 7 shows the resulting trajectory of carbon fluxes (F_D and F_B^O) during the Phanerozoic. The weathering rate of organic carbon (F_W^O) also varies, but within a range shown by a hatched region in Fig. 7. This figure clearly shows that the critical condition was never achieved during the Phanerozoic.

It should be noted, however, that the CO_2 fluxes (hatched region G in Fig. 7) might have been very close to the critical condition during the Late Paleozoic glaciation at about 300 Ma when F_W^O was very close to the Phanerozoic average (represented by thick line in Fig. 7).

In fact, based on studies of paleosols, pCO_2 is estimated to have been very low (as low as 150–200 ppm) at about 300 Ma (Mora *et al.*, 1996). Furthermore, based on studies of the distribution of ice-rafted debris, ice sheets could have advanced to $35°$ paleo-latitude in the Westphalian (in the Late Carboniferous) and Sakamarian (in the Early Permian) (Frakes and Francis, 1988; Frakes *et al.*, 1992). This value is close to the critical condition expected from EBMs (Fig. 3), supporting the idea that the Late Paleozoic glaciation may have been close to the critical condition for a global glaciation.

The large-scale glaciation of this period is considered to have been caused by a large drop in pCO_2 as a result of the rise of vascular plants which accelerated the rate of CO_2 uptake by weathering (Berner, 1993, 1997). These plants also provided a new source of bacterially resistant organic matter, such as lignin, removing CO_2 by the massive burial of sedimentary organic carbon during this period (Berner, 1993, 1997). For some unknown reason, however, the critical condition was not achieved at that time, and the Late Paleozoic glaciation did not develop into global glaciation.

3.2 Neoproterozoic

It has been argued that global glaciations occurred during the Proterozoic (Kirschvink, 1992; Hoffman *et al.*, 1998). If the Earth really experienced global glaciations during this period, how was the critical condition achieved?

One possible cause for global glaciations during the Proterozoic could be the significant reduction of the net CO_2 input flux, owing to an increase in the organic carbon burial and/or a decrease in the CO_2 degassing rate. According to studies of carbon isotopic composition of seawater during the Neoproterozoic, the average value of $\delta^{13}C$ was much higher than that for the Phanerozoic, and the $\delta^{13}C$ value increased up to $1‰$ before the glaciations (Kaufman *et al.*, 1977; Hayes *et al.*, 1999). This is interpreted as a result of the high rate of organic carbon burial, resulting in a large reduction of the net input flux of CO_2 and causing the global glaciations (Kaufman *et al.*, 1997; Hoffman *et al.*, 1998).

The above possibility was evaluated by Tajika (2004), using a carbon isotope mass balance model combined with the 1-D EBM and with the record of carbon isotopic composition of marine carbonates during the Neoproterozoic. Tajika (2004) found that an increase in the organic carbon burial alone cannot cause global glaciations. Here we recalculate the carbon cycle with the same model and evaluate the resulting carbon flux conditions in the F_D–F_B^O diagram (Fig. 8). Because we do not know the temporal variations of F_D and F_W^O during the Neoproterozoic, these parameters are assumed to be constant. We use several constant values for these parameters to evaluate the conditions for global glaciations.

The flux of the organic carbon burial increased at the timing of positive excursions of $\delta^{13}C$ during the Neoproterozoic. The increase in the organic carbon burial alone cannot, however, cause global glaciations when the present values
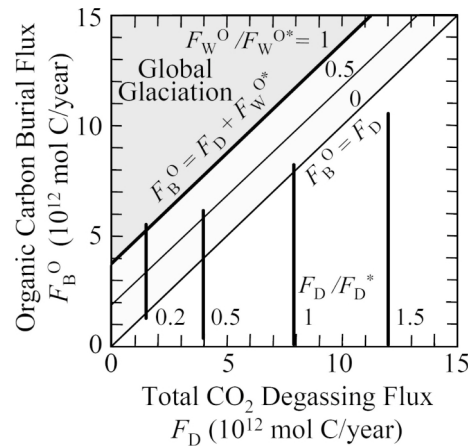

Fig. 8. Temporal variations of the organic carbon burial flux, F_B^O, through the Neoproterozoic estimated from a carbon isotope mass balance model with carbon isotope data (Hayes *et al.*, 1999). Weathering rate of organic carbon, F_W^O, and the total CO_2 degassing flux, F_D, are given as constant values in the model.

of F_D and F_W^O are assumed in the model (Fig. 8): the average surface temperature varies within the range of a very warm condition in this case (Fig. 9). This is because there were no vascular land plants before the Silurian—hence the soil biological activity, f_E, was low (<1), and so the climate should have been warm in order to balance the silicate weathering rate with a large net CO_2 input rate (see Eqs. (1) and (4)).

When the CO_2 degassing rate is assumed to be low ($<1/4$), however, the condition for global glaciation can be achieved at an extensive burial rate of organic carbon (Figs. 8 and 9). Such a decrease in the CO_2 degassing rate could have been resulted from a reduction of volcanic-metamorphic activities or a decrease in the average carbonate content in the subducting deep-sea sediments, owing to the changes in distribution (shallow/deep) of carbonate precipitation in the ocean. Lower weathering rate of organic carbon due to a lower level of oxygen content in the atmosphere during the Neoproterozoic than those of today (e.g., Canfield, 2005) might have also provided a favorable condition for the initiation of global glaciations (Fig. 8).

A recent study on the U-Pb age of detrital zircons from river sands in three major rivers in North and South America reveals that a time interval between 800–600 Ma is characterized by one of the lowest levels of continental growth (that is, subduction volcanism) during the geological history (Fig. 10; Rino *et al.*, 2004). This suggests that the period of the Neoproterozoic global glaciations seems to coincide with the period of the lowest volcanic activity, at least, at subduction zones. Because a significant portion of CO_2 degassing occurs via subduction volcanism today (Sano and Williams, 1996; Marty and Tolstikhn, 1998), the lowest activity of subduction volcanism might have resulted in the reduction of global CO_2 input.

There was another period of the very low population of zircon grains between 2500–2300 Ma (Fig. 10; Rino *et al.*, 2004). In this period, a series of glaciations occurred, and there is additional evidence for low-latitude ice-sheet (i.e., evidence for global glaciation) obtained from the Transvaal
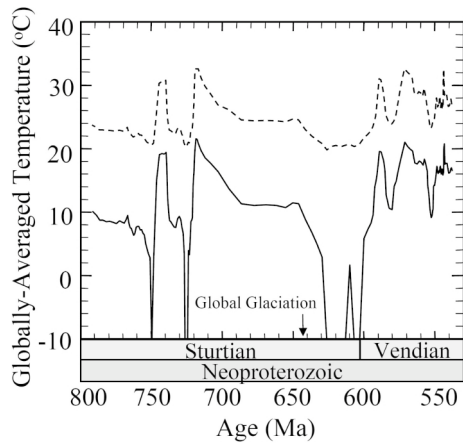
Fig. 9.   Temporal variations of globally-averaged surface temperature during the Neoproterozoic for the case of the present CO_2 degassing rate (dashed curve) and the case of the low CO_2 degassing rate (1/4 of the present rate; solid curve).
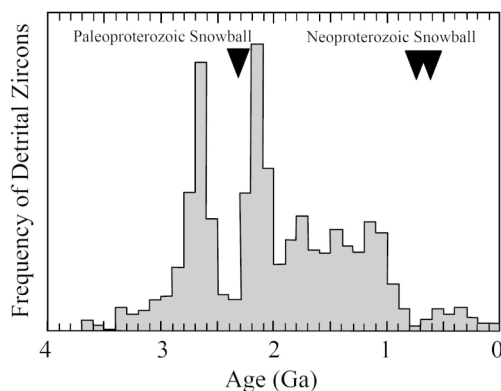

Fig. 10.   Age population of zircons in river sands of the North and South American continents (modified from Rino *et al.*, 2004).
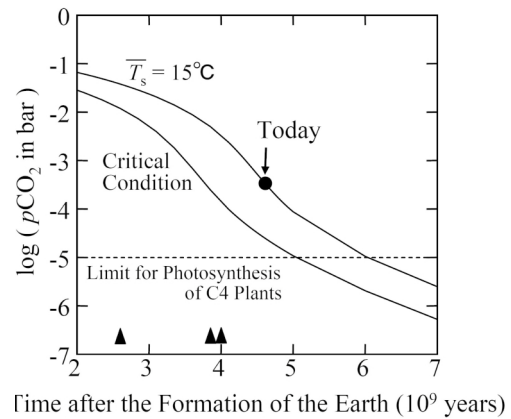

Fig. 11.   Temporal variations of the atmospheric CO_2 level for the critical condition and for a warm climate (globally-averaged surface temperature = 15°C) from 2.6 billion years before to 2.4 billion years after the present. Solid triangles represent the global glaciations during the Proterozoic. The horizontal line represents the critical CO_2 level for photosynthesis of C4 plants. Note that the climate model adopted here may not be accurate for low pCO_2 levels predicted for the future because radiative parameters were approximated with an error under 5 W/m² for 10^{-4} bar < pCO_2 < 2 bar (Caldeira and Kasting, 1992). Nevertheless, this result may represent a qualitative trend of the critical CO_2 level variation with time.

Supergroup on the Kaapvaal craton in South Africa (Evans *et al.*, 1997; Kirschvink *et al.*, 2000). Therefore, the Paleoproterozoic global glaciation could have been also explained by a low CO_2 degassing rate due to a reduction of volcanic-metamorphic activities.

4.   Susceptibility to Global Glaciations throughout the Earth's Evolution

According to studies of thermal evolution of the Earth, the potential temperature of the mantle has decreased through the Earth's history (e.g., Christensen, 1985). Then, an efficiency of magma generation, volcanism, and activity of plate tectonics (hence the CO_2 degassing via volcanism-metamorphism) may have decreased with time, suggesting that it might be more likely to cause global glaciations with time.

On the other hand, because the luminosity of the Sun increases as the Sun evolves (Gough, 1981), the critical level of CO_2 for the climate instability also changes with time (Fig. 5). Figure 11 shows temporal variations of the critical level of CO_2 and the atmospheric CO_2 level required for maintaining the average surface temperature at 15°C in the past and the future.

As the atmospheric CO_2 level decreases owing to the in-crease in the solar flux (e.g., Walker *et al.*, 1981), it will be under the limit for photosynthesis of C3 plants, and finally fall under the limit for photosynthesis of C4 plants (< 10 ppm). But before that, the critical CO_2 level (*not* the atmospheric CO_2 level) fall under the limit for photosynthesis of C4 plants after 500 million years from now (Fig. 11). Then, if a reduction of volcanic-metamorphic activities occurs and the atmospheric CO_2 level decreases to this limit, the net CO_2 input flux should become large because CO_2 cannot be fixed as organic carbon efficiently. In such a situation, even if CO_2 degassing rate becomes zero, global glaciation does not occur any more.

Although the susceptibility of the Earth to global glaciations in the evolution should depend on the competition of mantle cooling and solar luminosity enhancement, it will take a much longer time for the mantle activity to decrease drastically (e.g., the average mantle temperature would have decreased 150–250° for the last 3 billion years; Christensen, 1985). Therefore, before the mantle becomes cool enough to cause permanent global glaciation, the influence of the solar luminosity becomes important. As a consequence, the Earth will not be globally glaciated after 500 million years from now.

The Earth's climate could have been suitable for life during the most of the geological history because a certain level of volcanic-metamorphic activities due to spreading and subduction of oceanic plates has maintained sufficient net CO_2 input flux. As described earlier, even if the Earth was globally glaciated, the Earth could have escaped from such a regime, owing to an accumulation of CO_2 via volcanism-metamorphism.

In these respects, plate tectonics may be one of the necessary conditions for the Earth and Earth-like planets in extrasolar planetary systems to keep liquid water and life over the timescales of the planetary evolution.

5. Summary

Any decrease in the net input flux of CO_2 to the atmosphere-ocean system lowers the surface temperature and would eventually initiate global glaciation. The $F_D - F_B^O$ diagram is proposed for measuring the susceptibility of the Earth to global glaciations. By analyzing results from a carbon cycle model during the Phanerozoic, the net input flux of CO_2 may have been very close to the critical condition at the Late Paleozoic glaciation. During the Proterozoic, global glaciations occurred probably by reduction of the CO_2 degassing in addition to an increase in the organic carbon burial. It will be more unlikely for the Earth to be globally glaciated after 500 million years from now, because the Sun becomes brighter as it evolves. Continuous volcanic-metamorphic activities (i.e., plate tectonics) may be one of the necessary conditions for Earth-like planets in extrasolar planetary systems to keep liquid water and life on their surface over the timescales of the planetary evolution.

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