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Faint young Sun and the carbon cycle: implication for the Proterozoic global glaciations

Eiichi Tajika *

Department of Earth and Planetary Science, Graduate School of Science, University of Tokyo, 7-3-1 Hongo, Bunkyo-ku, Tokyo 113-0033, Japan

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Abstract

The Earth may have been globally ice-covered several times during the Proterozoic. While the Neoproterozoic and the Paleoproterozoic glaciations may have been 'snowball' Earth events, there is no evidence for such glaciation during the Phanerozoic. It might be hypothesized that a dimmer Sun earlier in Earth's history may have made the Earth more susceptible to global glaciation. In this paper, the roles of solar flux and soil biological activity in the carbon cycle and the climate during the Proterozoic are investigated using a simple carbon geochemical cycle model with a one-dimensional energy balance climate model. The results indicate, perhaps counterintuitively, that the Proterozoic Earth, with its dimmer Sun, was not more susceptible to 'snowball glaciation'. Metamorphic and volcanic CO_2 fluxes accumulate in the atmosphere and ocean until such time that those inputs are balanced by silicate weathering followed by carbonate precipitation and net organic carbon burial. Because of the dependence of weathering rates on climatic conditions, changes in geologic CO_2 inputs have a large influence on climatic conditions. In contrast, slow variation in solar flux has relatively little long-term impact on climate, because of large compensating changes in atmospheric CO_2 level. A reduction in CO_2 inputs lowers atmospheric CO_2 level, which finally initiates global glaciation. The atmospheric CO_2 level at the critical condition for a globally ice-covered state would have been high during the Proterozoic. However, roughly the same amount of CO_2 flux reduction is required for both the Proterozoic and the Phanerozoic. This is essentially because the temperatures at the critical condition are very low, hence the silicate weathering rate (which should balance with a net CO_2 input rate in a steady state) is also very low, regardless of the variation in solar flux. Furthermore, the effect of the lower solar flux on the CO_2 input rate at the critical condition would have been largely canceled by a lower efficiency of the silicate weathering rate due to lower soil biological activity during the Proterozoic. As a result, CO_2 flux conditions for initiating the global glaciation may be similar during both the Proterozoic and the Phanerozoic. Therefore, the explanation for the susceptibility of the Proterozoic Earth to 'snowball' conditions cannot hinge simply on the dimmer Sun; we must look to other differences in behaviors of the carbon cycle and the climate between these two ages.

Keywords: snowball Earth; faint young Sun; carbon cycle; silicate weathering

1. Introduction

Low-latitude ice sheets suggested by paleomag-

* Tel.: +81-3-5841-4516; Fax: +81-3-5841-4569.
E-mail address: tajika@eps.s.u-tokyo.ac.jp (E. Tajika).

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netic studies of glaciogenic deposits during the Neoproterozoic are interpreted as the results of global glaciations [1,2]. This 'snowball' Earth hypothesis may explain several unusual features observed in the Neoproterozoic glaciations, including the evidence for low-latitude ice sheets, the existence of iron formation and cap carbonate, and the negative excursion of the carbon isotopic composition of seawater [1–4].

According to studies of energy balance climate models (EBMs), the globally ice-covered state is one of the stable climate states of the Earth (e.g., [5–7]). A decrease in the partial pressure of carbon dioxide would be responsible for 'phase change' of the climate system of the Earth from the partially ice-covered to the globally ice-covered state [8,9]. Recent modeling studies show that there may be another solution of the climate system in which the Earth is not globally ice-covered, but has an equatorial belt of open water [10,11]. This solution, called 'slushball' Earth, may have the advantage of providing refugia for life during the Neoproterozoic glaciations [10], although it probably cannot explain the existence of iron formation and cap carbonate with unusual textures and low $\delta^{13}C$ values [4,12].

Recently, the Paleoproterozoic glacial deposits in the Transvaal Supergroup on the Kaapvaal craton in South Africa have been investigated, and the depositional latitude of Ongeluk lava, which is about 2.2 billion years old and conformably overlies the glaciogenic deposits, is estimated to be $11 \pm 5°$ [13]. Thus the snowball Earth event might have occurred during the Paleoproterozoic as well [13,14].

Here, it is interesting to note that there is evidence for low-latitude ice sheets in all major glaciations (the Huronian, the Sturtian, and the Marinoan–Varangar glaciations) during the Proterozoic (e.g., [3,13,15]), although such evidence has never been found in the glaciations during the Phanerozoic (e.g., [16]). Are there any reasons for the systematic difference in the glaciations between the Proterozoic and the Phanerozoic? Were there any conditions preferable for global glaciations that prevailed during the Proterozoic?

One of the most different factors between the Proterozoic and the Phanerozoic is the solar flux. According to the standard model of solar evolution [17], the luminosity of the Sun increases with time. The solar luminosity is estimated to be 83% and 94% of the present-day value at the Huronian glaciations during the Paleoproterozoic (2.4 Ga) and at the Sturtian glaciations during the Neoproterozoic (0.75 Ga), respectively (Fig. 1). Because the climate of the Earth depends largely on the energy balance between the solar incident flux and the outgoing infrared radiation, this factor may have promoted cooler conditions during the Proterozoic. It is, however, noted that there is no certain record of glaciation during the mid-Proterozoic, although evidence of possible glacial deposits has been found in north-central Siberia and north Scotland in this interval (e.g., [18]). The mid-Proterozoic glacial gap between 2.2 and 0.9 Ga suggests that the lower solar flux alone cannot be the reason for initiation of global glaciations. However, the lower solar flux during the Proterozoic could have provided a tendency or background condition in which global glaciations can be caused easily once the climate becomes cold. Because of this, the Earth may have been more susceptible to global glaciation during the Proterozoic.

Another factor which differs greatly between the Proterozoic and the Phanerozoic may be the efficiency of silicate weathering on land surface due to soil biological activity [19–22]. Plants accelerate the uptake of CO_2 during weathering by a variety of mechanisms including the secretion of organic acids by roots and associated symbiotic microflora, the recirculation of water by transpiration and accelerated rainfall, and the retention of soil by roots from removal due to erosion [19–22]. Because there were no terrestrial higher plants before the Silurian, the efficiency of the silicate weathering rate due to the soil biological activity should have been much lower during the Proterozoic than the present [19–22]. It must have affected the level of atmospheric CO_2 through the carbon cycle.

In this study, the roles of solar flux and soil biological activity in the carbon cycle and the climate during the Proterozoic are investigated using a simple carbon geochemical cycle model combined with a one-dimensional (1-D) EBM. The

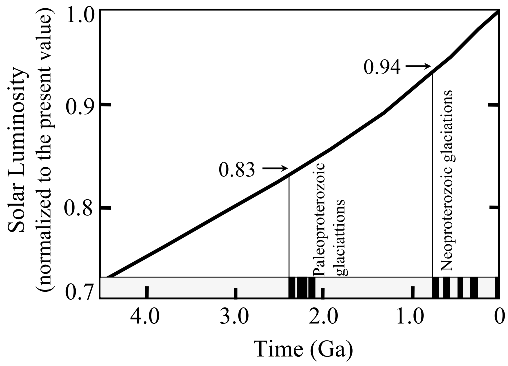

Fig. 1. Increase in solar luminosity with time. The solar evolution model is based on Gough [17]. Major glacial events are also shown as black bars.

possibility of the lower solar flux during the Proterozoic as a cause of the systematic difference in the glaciations between the Proterozoic and the Phanerozoic will be discussed.

2. Models

The atmospheric CO_2 level on a geologic timescale is controlled by the carbon cycle (e.g., [21]). The mass balance equation of carbon in the atmosphere–ocean system is fully expressed as follows [23]:

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

where M_{AO} = amount of carbon in the atmosphere–ocean system, F_V = degassing rate of mantle CO_2 via volcanism, F_M = degassing rate of CO_2 due to metamorphism–volcanism at the subduction zone, F_W = weathering rate, and F_B = burial rate. Subscripts r, h, and s represent volcanism at mid-ocean ridge, hot spot, and subduction zone, respectively. Superscripts C and O represent carbonate carbon and organic carbon, respectively.

In the above equation, the growth rate of the carbonate carbon reservoir ($= F_B^C - F_W^C$) can equate with the rate of uptake of atmospheric CO_2 via the weathering of Ca–Mg silicates followed by precipitation of carbonates (F_W^S; hereafter referred to as the silicate weathering rate) when the mass balance of calcium ion in the ocean is assumed to be in a steady state [21]. We integrate CO_2 degassing rates into the total CO_2 degassing rate F_D ($= F_{V,r} + F_{V,h} + F_{V,s} + F_M^C + F_M^O$). We also assume that the atmosphere–ocean system is in a steady state (that is, $dM_{AO}/dt = 0$) for the time scale of more than 10^5 years (residence time of carbon within the atmosphere–ocean system). Then the above equation is rewritten as:

$$0 = F_D + F_W^O - F_B^O - F_W^S \quad (2)$$

Among these fluxes, only the silicate weathering rate (F_W^S) is known to have dependences on temperature (T) and partial pressure of CO_2 (pCO_2) (e.g., [21,24–26]). Therefore, Eq. 2 can be simplified as:

$$F_{in} = F_W^S\,(T,\ pCO_2) \quad (3)$$

where F_{in} is defined as the net CO_2 flux into the atmosphere–ocean system expressed as follows:

$$F_{in} = F_D + F_W^O - F_B^O \quad (4)$$

As shown in Eq. 3, the silicate weathering rate represents the net CO_2 removal flux from the atmosphere–ocean system and should balance with the net CO_2 input flux in a steady state [27]. The total flux of CO_2 consumption due to silicate weathering today has been estimated to be $(11.5–23.0) \times 10^{12}$ mol C/yr (e.g., [28–30]). The net CO_2 input flux F_{in} should, however, equate actually with the net CO_2 consumption rate via the weathering of Ca–Mg silicates followed by carbonate precipitation (half of the flux of CO_2 consumption due to the weathering of Ca–Mg silicates). In this study, a value of 6.65×10^{12} mol/year [21] is assumed for F_{in} ($= F_W^S$) at present.

Silicate weathering rate is given as a function of temperature and CO_2 partial pressure:

$$F_W^S \propto f_E \cdot (pCO_2/pCO_2^*)^n \exp(-E/RT) \quad (5)$$

where n is the exponent of pCO_2 dependence, E is activation energy, and R is the gas constant (e.g., [19,21,24]). The factor $(pCO_2/pCO_2^*)^n$ represents the response of the weathering rate against pCO_2 changes. The ratio pCO_2/pCO_2^* is assumed to be a ratio of soil pCO_2 to the present value, $pCO_{2,soil}/pCO^*_{2,soil}$, for the post-Silurian case (after the emergence of vascular land plants [31]). For the case of today (the post-Silurian case), we adopt the formulation developed by Volk [31] as follows:

$$\frac{pCO_{2,soil}}{pCO^*_{2,soil}} = \frac{\Pi}{\Pi^*}\left(\frac{1-pCO^*_{2,atm}}{pCO^*_{2,soil}}\right) + \frac{pCO_{2,atm}}{pCO^*_{2,soil}} \quad (6)$$

$$\Pi = \Pi_{max}\frac{pCO_{2,atm}-pCO_{2,min}}{pCO_{2,1/2}+(pCO_{2,atm}-pCO_{2,min})} \quad (7)$$

where Π is the total (above-ground+below-ground) productivity, Π_{max} is the maximum productivity, $pCO_{2,1/2}$ is the value of pCO_2 at which $\Pi = 0.5\Pi_{max}$, and $pCO_{2,min}$ is the value of pCO_2 at which the rate of carbon fixation just balances the photorespiration [31]. These equations are based on the Michaelis–Menten formulation which is generally useful in describing the effect of CO_2 on plant growth [31]. For the pre-Silurian case (before the emergence of vascular land plants), the ratio pCO_2/pCO_2^* is assumed to be the ratio of atmospheric pCO_2 to the present value, $pCO_{2,atm}/pCO^*_{2,atm}$ [19,21,24,32–34]. For Ca–Mg silicates exposed on the continents of an abiotic Earth, n is probably between 0.3 and 0.4 [19]. For simplicity, a value for n of 0.3 [19,24,31–34] is used in this study. Note that this dependence is originally derived from early laboratory experiments, but it is not clear what the dependence really is or even if it does exist. It is, however, noted that, even when we assume $n = 0$, the conclusion of this study does not change (see Section 3.3). The activation energy of chemical dissolution of various silicates is estimated to range from 10 to 20 kcal/mol based on laboratory experiments [25], although it has not been constrained well from field studies (e.g., [35]). In this paper, a value for E of 15 kcal/mol [21] is used as a standard. The use of different values for E would modify the results, but does not change the conclusion.

The factor f_E represents the efficiency of the weathering rate depending on soil biological activity due to land plants and microbes [21]. The biota accelerates chemical weathering by stabilizing soil which is a reservoir of high surface area and acts as a medium for acid attack, producing organic and inorganic acids, and contributing to physical weathering through microfracturing of mineral grains [19,21]. Because there were no terrestrial higher plants before the Silurian, the efficiency of the silicate weathering rate due to soil biological activity should have been lower during the Proterozoic than the present (that is, $f_E < 1$ during the Proterozoic, whereas $f_E = 1$ at present). According to Schwartzman and Volk [19], chemical weathering under biotic conditions is significantly enhanced relative to abiotic continental crust exposed to similar rainfall, temperature, and pCO_2 conditions. The enhancement factor of weathering at present relative to that of an abiotic Earth would be on the order of, at least, 100 to > 1000 [19]. However, microbial and possibly lichen colonization of the continents in the Precambrian probably resulted in a considerable increase in chemical weathering relative to abiotic conditions [19]. The factor of soil biological activity before the Silurian may be estimated from the weathering rate in unvegetated regions on the present Earth [21]. For example, Berner [20,21] argued that, based on analysis of water chemical composition in a region of the southern Swiss Alps [36], the silicate mineral weathering rate decreases by a factor of about seven in unvegetated regions compared to vegetated regions of the same watersheds. Therefore, a value for f_E of 0.15 [21] is used in this study for the Proterozoic. Note that if we assume microbial and lichen colonization of the continents during the Proterozoic, we may have to assume another response function against pCO_2. But, because it is not clear how to formulate a function for soils with microorganisms (e.g., [21]), we simply assume a ratio of atmospheric $pCO_{2,atm}$ during the Proterozoic.

There are other important factors which affect the silicate weathering rate. For example, river runoff, which also has a dependence on temperature, should affect the silicate weathering rate [21,24,29]. However, it is not considered here for simplicity. Similarly, other factors (such as river runoff due to changes in paleogeography, land area, uplift and physical erosion, and lithology) are not considered here for simplicity. These assumptions do not change the conclusion of this study qualitatively.

Because the surface temperature is controlled by the greenhouse effect of CO_2, the atmospheric CO_2 level should be controlled by variation of F_{in} according to Eq. 3. For example, an increase in the net CO_2 input rate results in an increase in the silicate weathering rate through an increase in the atmospheric CO_2 level (hence warmer climate). On the other hand, a decrease in the net CO_2 input rate results in a decrease in the silicate weathering rate through a decrease in the atmospheric CO_2 level (hence colder climate).

Behaviors of the climate system of the Earth can be understood by using a 1-D EBM which considers energy balance among the solar incident flux, the latitudinal heat transport, and the outgoing infrared radiation. An ice cap with high surface albedo is assumed to expand from the pole to lower latitudes. In order to obtain the response of the climate system to variations of the atmospheric CO_2 partial pressure, a 1-D EBM with CO_2-dependent outgoing radiation [8,9] is adopted in this study. This model is based on the Budyko–Sellers-type EBM, which employs diffusive-type heat transport and discontinuous albedo at the ice cap edge [5–7], and takes into account the dependence of pCO_2 on the outgoing infrared radiation (for a detailed description of the model, see [8,9]). The following model parameter values are used as a standard: ice albedo = 0.62 and land/ocean albedo = 0.3 [7]. The diffusion coefficient for latitudinal heat transport is assumed to be 0.455 $W/m^2/K$ which is determined from obtaining the present condition (solar constant = 1, ice line = 0.95, pCO_2 = 300 ppm). From this model, we can estimate latitudinal temperature distribution as a function of the effective solar constant (S) and pCO_2.

The carbon cycle model is combined with the 1-D EBM to estimate variations of the global silicate weathering rate due to variations of the latitudinal distribution of the surface temperature. For simplicity, continents and oceans are assumed here to distribute uniformly from the pole to the equator as an ideal case. Silicate weathering is assumed to occur as long as the surface temperature is above 0°C. The global silicate weathering rate is estimated by integrating weathering rates from the pole to the equator in all latitude bands using the latitudinal temperature distribution obtained from the 1-D EBM.

3. Results and discussion

3.1. Effect of faint young Sun on the climate

Fig. 2 shows the results of steady-state solutions of the 1-D EBM for the standard case. Because the luminosity of the Sun becomes larger with time, Fig. 2 shows the results of three cases for the effective solar constant (normalized to the present-day value) S of 1.0 (today), 0.94 (Neoproterozoic, about 750 Ma), and 0.83 (Paleoproterozoic, about 2400 Ma). It is well known from the previous studies on 1-D EBMs (e.g., [5–9]) that stable climate states can be classified into three branches: an ice-free branch, a partially ice-covered branch, and a globally ice-covered branch (Fig. 2). As pCO_2 decreases from the partially ice-covered branch, the ice line advances to lower

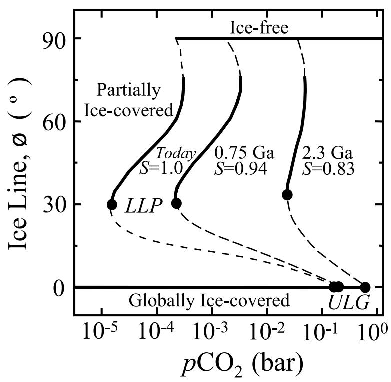

Fig. 2. Standard results of steady-state solutions of a 1-D EBM for S = 1.0 (today), 0.94 (Neoproterozoic), and 0.83 (Paleoproterozoic). Solid circles represent the lower limit of a partially ice-covered branch (LLP) and the upper limit of a globally ice-covered branch (ULG).

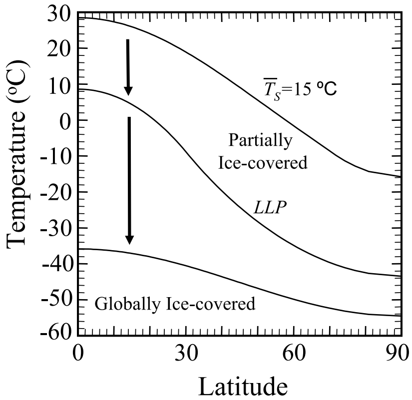

Fig. 3. Temperature distribution from a warm climate (the globally averaged surface temperature $\overline{T}$ is 15°C) to the critical condition (LLP), and to the globally ice-covered state.

latitudes. When pCO_2 decreases to some critical value and the ice line reaches about 30°, the stable solution disappears. Then, the Earth will fall into the globally ice-covered branch owing to a climate jump derived from ice albedo feedback (so-called large ice cap instability). The CO_2 levels at the critical condition where the partially ice-covered branch disappears (represented by LLP, the lower limit of partial ice cover) are one and three orders of magnitude higher for $S = 0.94$ and 0.83, respectively, than that for $S = 1.0$ (Fig. 2).

The globally ice-covered Earth has very low surface temperatures (Fig. 3), and is stable with respect to energy balance. This is because the net solar incident flux which is small owing to high surface albedo balances the outgoing radiation which is also small owing to low surface temperatures (Fig. 3). Even when pCO_2 increases slightly, the equator is still under freezing condition, thus the Earth remains ice-covered globally, keeping the high surface albedo. In order to escape from the globally ice-covered branch, pCO_2 should increase to on the order of 0.1 bar (Fig. 2) via volcanic degassing of CO_2. Then, the climate system becomes unstable and will jump to the ice-free branch.

The results depend on model parameters, such as surface albedo and thermal diffusion coefficient for latitudinal heat transport (e.g., [9]), and also on climate models (e.g., [37–39]). However, basic features of the solutions described above may not change. Because a parameter study of the EBM is not the main purpose of this paper, and also because the conclusion of this paper does not change qualitatively, the following discussion will be based on the standard results.

Fig. 4a shows the results of temporal variations of the atmospheric CO_2 level required for maintaining a constant globally averaged surface temperature (15, 10, 5, and 0°C, and LLP) over the last 3.0 billion years against the increase in solar flux (Fig. 1). Fig. 4b shows a close-up of variations of pCO_2 for the last 1.0 billion years. In

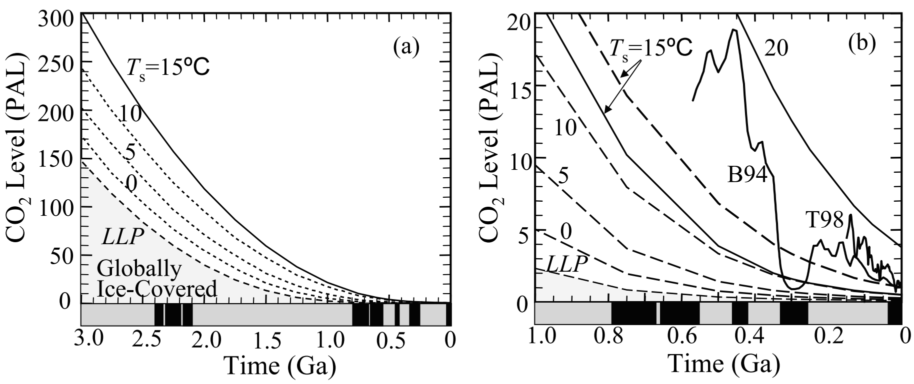

Fig. 4. The CO_2 levels for maintaining a constant globally averaged surface temperature (15, 10, 5, 0°C, and LLP) through time. The solid curve represents the ice-free branch, and the broken curve represents the partially ice-covered branch. Black bars represent major glaciations. (a) Estimated CO_2 levels for the standard case through the past 3.0 billion years (PAL = present atmospheric level). (b) Estimated CO_2 levels for the standard case through the past 1.0 billion years. Estimates of variations of pCO_2 using carbon cycle models (B94 = Berner [21]; T98 = Tajika [26]) are also shown for reference.

reality, pCO_2 has fluctuated with time, which results in climate change over a geologic time scale (e.g., [21,26]). In Fig. 4b, variations of CO_2 levels estimated from carbon cycle models [21,26] are also shown for reference. As shown by these figures, because the solar flux was lower in the past, the atmospheric CO_2 level required for maintaining a constant temperature was higher in the past than it is today. This tendency of pCO_2 variation, that is, the CO_2 level has decreased with time, is expected from a response of the carbon cycle system to the increase in solar flux with time [24,32–34,40].

In the case of $S = 1.0$, the CO_2 level at the critical condition (LLP) is one order of magnitude lower than the present level (Fig. 2). On the contrary, in the case of $S = 0.83$, the critical CO_2 level is only one half to one third of the CO_2 level for a warm climate similar to the present (Figs. 2 and 4a). It is therefore suggested that the critical condition could have been achieved more easily during the Proterozoic than the Phanerozoic owing to small relative changes of the atmospheric CO_2 level, although we cannot conclude it because the absolute amount to be changed is much larger in the Paleoproterozoic. We therefore need to understand the controlling mechanism for variations of CO_2 levels.

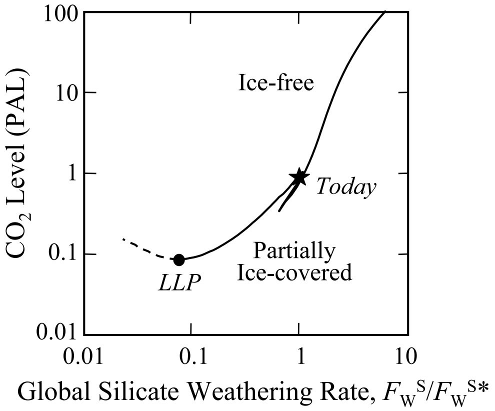

Fig. 5. Relation between the global silicate weathering rate (= the net CO_2 input rate in a steady state; normalized to the present rate) and the CO_2 level under the present solar luminosity.

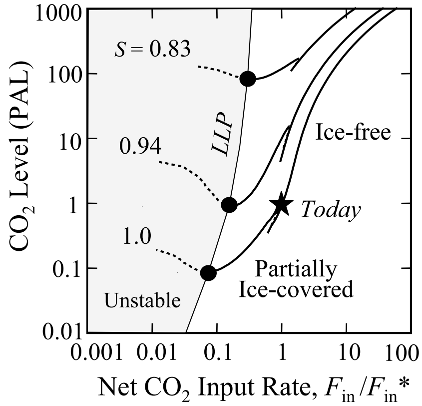

Fig. 6. The atmospheric CO_2 levels as a function of the net CO_2 input rate (normalized to the present rate) for $S = 1.0$, 0.94, and 0.83. The flux at the critical condition is represented by a solid circle for each case. The effect of a change in soil biological activity is not considered.

3.2. *Effect of faint young Sun on the carbon cycle*

The CO_2 level of the atmosphere is controlled by the carbon cycle. In the long-term carbon cycle, the surface temperature and thus the atmospheric CO_2 level is determined through Eq. 3. Fig. 5 shows a relation between the CO_2 level and the global silicate weathering rate under the present solar constant ($S = 1$), which is obtained by integrating the weathering rate from the pole to the equator as a function of temperature at each latitude. As the net CO_2 input rate decreases, pCO_2 and the temperatures should lower, thus the global silicate weathering rate decreases. At the critical condition (LLP), the global silicate weathering rate is very small (Fig. 5) because the climate is very cold (Fig. 3), irrespective of the values of the model parameters.

The CO_2 levels at the critical condition for $S = 1.0$, 0.94, and 0.83 as a function of the net CO_2 input rate are compared in Fig. 6 (the present value for the soil biological activity, that is, $f_E = 1$, is assumed here for all cases). It is noted that the silicate weathering rate responds to variations of the net CO_2 input rate (F_{in}) through the carbon cycle, and should balance with F_{in} in a steady state. Thus, the case for $S = 1$ in Fig. 6 is equivalent to the curve in Fig. 5. As shown in Fig. 6, the net CO_2 input rates at the critical condition are higher in the past than today. This is owing to the dependence of pCO_2 on the silicate weathering rate. Because the atmospheric CO_2 levels were higher in the past owing to the lower solar flux, the silicate weathering rate may have been higher even when the temperatures were the same. This means that the net CO_2 input rate may not have to decrease as low as the present critical level in order to achieve LLP during the Proterozoic. If it were the case, the critical condition could have been achieved more easily in the past, because smaller changes in the CO_2 fluxes could have resulted in the global glaciation during the Proterozoic.

In any case, however, the net CO_2 input rate required for achieving the critical condition should be low because the temperatures are essentially low at the critical condition.

3.3. *Effect of soil biological activity on the carbon cycle*

When we consider the difference in soil biological activity for the silicate weathering rate between the Proterozoic and the Phanerozoic in addition to the solar flux, the results should change. Fig. 7 shows the net CO_2 input rate required for maintaining a constant temperature (15, 10, 5, 0°C, and LLP) when changes of soil biological activity in addition to the solar flux change are considered. In this figure, the soil biological activity is changed from 0.15 to 0.75 due to the rise of vascular land plants in the Silurian–Devonian, and is also changed from 0.75 to 1.0 due to the rise of angiosperms in the Cretaceous [21]. It is clearly shown that the net CO_2 input rate required for maintaining a constant temperature should have changed with time because of changes both in solar flux and in soil biological activity with time (Fig. 7). The tendency of a decrease in the net CO_2 input rate to maintain a constant temperature with time is the result of the increase in solar flux, but sudden increases at around 350 and 130 Ma are the results of increases in soil biological activity. In other words, if the net CO_2 input rate is constant through time, the surface temperature should have increased owing to the increase in solar flux, but decreased abruptly

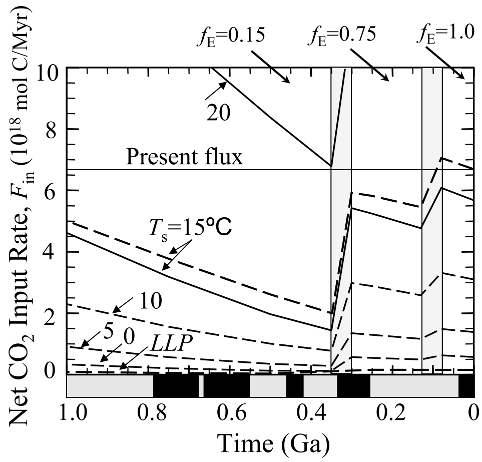

Fig. 7. Net CO_2 input rate required for maintaining a specific temperature (20, 15, 5, 0°C, and LLP) during the past 1.0 billion years. The effect of variations of soil biological activity (f_E) due to evolution of terrestrial higher plants on the silicate weathering rate in addition to the solar luminosity change is considered. The solid curve represents the ice-free branch, and the broken curve represents the partially ice-covered branch.

at around the times of the rise of vascular land plants and of angiosperms. In fact, there is a possibility that this may have been one of the causes for the Late Paleozoic (Gondwana) glaciation at around 300 Ma [20,41].

Fig. 8 shows the atmospheric CO_2 levels at the critical condition for $S = 1.0$, 0.94, and 0.83 as a function of net CO_2 input rate for the case where change in the soil biological activity in addition to the solar flux change is considered. As shown in this figure, the net CO_2 input rates at the critical condition during the Proterozoic are estimated to be similar to or even lower than the present. This is because the efficiency of the silicate weathering rate on the land surface must have been very low during the Proterozoic owing to the low soil biological activity. Therefore, in order to achieve the critical condition (LLP), a similar or even lower flux of net CO_2 input is required during the Proterozoic compared to the present flux. The influence of the lower solar flux during the Proterozoic is cancelled out completely by the low efficiency of the silicate weathering rate due to low soil biological activity at that time (Figs. 6 and 8). It is therefore suggested that the condition for achiev-

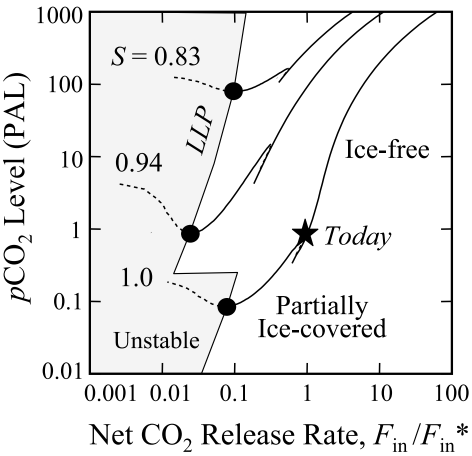

Fig. 8. Atmospheric CO_2 levels as a function of the net CO_2 input rate (normalized to the present rate) for $S=1.0$, 0.94, and 0.83. The flux at the critical condition is represented by a solid circle for each case. The effect of change in the soil biological activity in addition to the solar luminosity change is considered.

ing LLP, that is, the condition for the net CO_2 input rate to achieve the critical condition, is basically the same between the Proterozoic and the Phanerozoic. From a viewpoint of the carbon cycle, the difficulty in initiation of a snowball Earth event may not be different between these two ages, irrespective of the different solar fluxes. Note that even if the silicate weathering rate actually does not depend on pCO_2 (i.e., $n=0$), this conclusion does not change. In that case, the differences between the results which only consider the different condition of solar flux (Fig. 5) should disappear, but the lower soil biological activity during the Proterozoic should decrease the LLP (that is, the net CO_2 input fluxes at the critical condition during the Proterozoic are much lower than those obtained here). Therefore, if it were the case, the conclusion obtained here should be strengthened.

Then, what is the cause for global glaciations? There are three possibilities for the net CO_2 input rate F_{in} ($=F_D+F_W^O-F_B^O$) to decrease to the critical level: (1) a decrease in the degassing rate (F_D), (2) a decrease in the weathering rate of organic carbon (F_W^O), and/or (3) an increase in the burial rate of organic carbon (F_B^O). There is evidence for increases in organic carbon burial before the Neoproterozoic glaciations [2,42]. Sedimentary environments may have been very different from the present one. Fragmentation of a supercontinent, Rodinia, could have provided a large area of continental margins at which organic carbon is buried effectively, and may have resulted in the CO_2 drawdown [2]. The carbon cycle during the Proterozoic might have been affected also by the position of the supercontinent: in order to decrease the CO_2 level by silicate weathering in the last stage of cooling, it is required for the continental land masses to be located in the middle to low latitudes (e.g., [1,37,43]). If there were no land masses in the middle to low latitudes, silicate weathering on the land surface would not take place, thus CO_2 would not decrease any more (e.g., Fig. 5). In that case, it will be difficult for global glaciation to occur [37,43]. In this respect, the uniform distribution of land masses assumed in this study may satisfy this condition, although land masses at high latitudes are not important in the last stage of cooling (because silicate weathering cannot occur at high latitudes when the surface is covered with ice). On the other hand, variations in the CO_2 degassing rate due to the Wilson cycle may also be important, because there might have been periods when the CO_2 degassing rate was very low. In such a case, the net CO_2 input rate could decrease to a very low level, achieving the critical condition on the order of 10^5 years [27,44].

In any case, the reason for the apparent systematic difference in the glaciations between the Proterozoic and Phanerozoic remains to be determined. However, because the susceptibility to global glaciation during the Phanerozoic may be the same as that during the Proterozoic, there would be special conditions or perturbations which have occurred during the Proterozoic, but, at least, not yet during the Phanerozoic. Breakup of low-latitude supercontinents may have played an important role in the initiation of global glaciation [1,2,43], or collapse of methane atmospheres during the Paleoproterozoic and the Neoproterozoic might have caused the snowball Earth events [14,43]. Or there might have been periods of large reduction in volcanic activities, which could have been the cause of CO_2

decrease [27,44]. It is, at least, clear that we must look to possibilities other than the difference in solar flux.

4. Conclusion

It might be hypothesized that a dimmer Sun might make Earth more prone to global glaciation. For example, if CO_2 degassing from volcanoes and metamorphism were to diminish, it might be thought that the Earth would more likely fall into global glaciation with a dimmer Proterozoic Sun than a brighter Phanerozoic Sun. However, this study indicates this hypothesis may be wrong.

The Proterozoic Earth should have had a high level of atmospheric CO_2 as a response of the carbonate–silicate geochemical cycle to lower solar irradiance. The CO_2 level for initiation of global glaciation would also have been higher in the past. Despite increasing solar luminosity, the critical net CO_2 input flux below which global glaciation would be initiated may have been essentially constant for the past 2.5 billion years or more. This is because the temperatures at the onset of global glaciation must be low, hence the silicate weathering rate (which should balance the net CO_2 input rate in a steady state) should be also very low, regardless of the variation in solar flux.

The dimmer Sun would tend to allow global glaciation to be initiated with a greater CO_2 flux. However, with the near absence of terrestrial higher plants during the Proterozoic, much more atmospheric CO_2 and higher temperatures would have been necessary to weather enough silicate rock to consume the metamorphic and volcanic CO_2. These two effects tend to work in opposite directions, such that the propensity of CO_2 flux reductions to cause global glaciation seems no greater in the Proterozoic than in the Phanerozoic. This means that the susceptibility to global glaciation may not be different between the Proterozoic and the Phanerozoic.

Thus, the reason for the systematic difference in the glaciations between the Proterozoic and the Phanerozoic lies somewhere other than in the differential response of the climate and carbonate–silicate cycle to diminished geologic CO_2 inputs under conditions of reduced solar luminosity. Special conditions, such as breakup of low-latitude supercontinents, collapse of methane atmospheres, or large reductions in volcanic activities, might have caused the snowball Earth events during the Proterozoic.

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