Life as a greenhouse climate regulator through Earth's history

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The atmospheric concentration of greenhouse gasses is influenced both by non-biotic geochemical processes, such as volcanism and weathering of silicate rocks, and by the photosynthetic biota. The geochemical regulator of climate would work also on a dead planet. Here we argue from a simple model that the biotic regulation of climate might be of bigger importance, and that life itself could be responsible for the habitability of Earth through its geological history. The model predicts the surface temperature and atmospheric concentrations of carbon dioxide, methane and oxygen as a function of the solar luminosity in good agreement with the geological record through the entire history of the Earth. The biological regulator substantially widens the habitable zone for possible extra-terrestrial

Earth-like life. With future increasing solar luminosity the Earth will enter a regime where two stable states, a live planet and a dead planet, are possible.

Introduction

The average temperature on the surface of the Earth has been above freezing through its geological history, as required for the development of life, despite an up to 30% fainter young Sun. The surface temperature is regulated by the amount of greenhouse warming from water vapor and other greenhouse gasses. These other greenhouse gasses are mainly carbon dioxide, but also methane, which could persist in an early oxygen-free atmosphere. Primitive life has indeed existed on Earth since early in its geological history. Stromatolites, which are fossils of photosynthesizing cyanobacteria are found as old as 3.5 Ga BP (giga-anni = billion years, BP = before present) (Schopf, 1993). Isotope fractionation in carbon found in 3.8 Ga old rocks form Isua, Greenland indicates a biological origin (Schidlowski, 1988; Rosing, 1999). This is perhaps within a few hundred million years the earliest time of stable planetary climate possible for a biota to evolve. Impacts in the Hadean eon (4.6 - 3.8 Ga BP) might have released enough energy to vaporize the entire world ocean and thus sterilizing the Earth if life should have existed at that time. (Kasting & Catling, 2003). The photosynthetic organisms gradually changed the chemical composition of the atmosphere to its present state, away from chemical equilibrium, by consuming CO_2 and producing O_2 (Lovelock & Watson, 1982). However, it was not until around the Archaean-Proterozoic transition, 2.3 Ga BP, that the atmospheric content of oxygen rose rapidly almost to its present level, making way for dominance of aerobic life forms. This transition is sometimes called the great oxygen event (GOE). The reason for the low oxygen level prior to 2.3 Ga BP, despite the presence of photosynthetic organisms, could be that Earth's surface was strongly reducing until free iron was depleted from the surface and ocean waters, as documented in the banded iron formations (Cloud, 1972). It was recently proposed that the archean atmosphere could exist in one of two possible steady states with low ($\sim 10^{-6}$ bar) or high ($> 10^{-2}$ bar) oxygen concentration (Goldblatt *et al.*, 2006). The reason for this should be the shielding by the ozone layer in a high oxygen concentration atmosphere of ultra-violet radiation needed for oxidizing methane. Here we shall assume the former simpler explanation for the great oxidation event by changing reduction state of the surface and ocean waters.

The existence of life and the presence of liquid water in most of Earth's geological history show that the climate has been relatively constant despite the 30 percent increase in the solar luminosity through that time. This deviation from an expected frozen early Earth is the "Faint young Sun paradox". The enigma has been proposed resolved by an early enhanced greenhouse warming due to higher concentration of atmospheric CO_2 (Kasting *et al.*, 1984; Kiehl & Dickinson, 1987). The atmospheric CO_2 concentration is governed by the balance between sources and sinks. On the abiotic Earth this would be the geochemical cycle of carbon exchange with the crust through volcanism (source) and silicate rock weathering (sink).

However, the fossil soil sediment record, paleosols, constrain the early atmospheres CO_2 content to less than about 100 PAL (present atmospheric level)

(Rye *et al.*, 1995). This amount of atmospheric CO_2 is not enough for greenhouse warming to compensate for the faint young Sun. The deficit could be accounted for by an additional greenhouse warming from methane, which can only exist abundantly in the oxygen free early atmosphere (Kasting, 2005; Catling *et al.*, 2001).

The idea that life contributes to the regulation of the geochemistry and climate is not new (Hutchinson, 1954; Holland, n.d.), it was stated in a strong form in the Gaia hypothesis (Watson & Lovelock, 1983), where the biota is seen as self-regulating in order to maintain optimal environmental conditions. The primary evidence in support of the hypothesis is the observation that Earth's atmosphere is in a state far from chemical equilibrium. This is an indication in support of the hypothesis, but not a proof. On a dead planet with active tectonics the geochemistry, driven by the heat flow from the planetary core, is also maintained away from chemical equilibrium. For a critical discussion of the gaia hypothesis and its testability see (Kirchner, 2002).

In the simplest form the biotic regulation works through the control of the atmospheric gas concentrations which determines the surface temperature by greenhouse gas warming. This in turn feeds back on living conditions for the biota. In the following we present a minimal model describing this dynamics of a living planet, se figure (1). The conceptual model consists of three parts; the radiative energy balance of the surface determined by the greenhouse gas concentrations, the growth of the biota determined by the surface temperature and the chemical composition of the atmosphere determined by the biota.

The energy balance

The mean surface temperature of the planet is determined by the balance between incoming solar radiation R_i and outgoing long wave radiation R_o(Bodyko, 1969; Sellers, 1969). The change of the temperature T is determined by the difference $R_\mathrm{i} - R_\mathrm{o}$ in incoming and outgoing radiation according to the energy balance;

$$c\frac{\mathrm{d}T}{\mathrm{d}t} = R_\mathrm{i} - R_\mathrm{o} = (1-\alpha)S - \sigma(T - T_g)^4, \tag{1}$$

where c is the heat capacity, α the planetary albedo and $S = \tilde{S}/4$ is a quarter of the solar constant. (The quarter comes from the ratio of the cross-sectional area to the surface area of the sphere). Finally, σ is the Stefan-Boltzmann constant. The energy balance depends, among other factors, on the atmospheric greenhouse warming and the planetary albedo. The planetary albedo is not a constant factor; it depends on the state of the climate itself, through both non-biotic factors and biotic factors. Non-biotic factor: The feedback of clouds on temperature is very complicated. It depends on the height in the atmosphere where the clouds are situated, the state of the atmosphere surrounding the clouds and the optical properties of the specific cloud. The clouds cool by reflecting the incoming radiation and they heat by trapping the outgoing radiation. The net effect of clouds in the present climate is a cooling of about 20 W/m^2 (Ramanathan *et al.*, 1989). Ice and snow on the surface unambiguously cool by reflecting the incoming short-wave radiation, thus the amount of clouds and of ice and snow strongly influences the planetary albedo. In geological perspective, where the extent and location of continents change, the difference in albedo between land and ocean also influences the planetary albedo.

Biotic factors: The land covered with plants will have a lower albedo than bare land, partly reflecting the fact that plants utilize the incoming radiation for photosynthesis. The effect on planetary albedo from marine life is somewhat smaller and opposite; plankton reflects more sunlight than clear bluewater. Another effect of the marine biota is the emission of demethyl sulfide leading to aerosols in the atmosphere, which could enhance the cloud cover, and thus increase the planetary albedo (Hegg *et al.*, 1990). Simple modeling of the effect of biota on the planetary albedo can be found elsewhere (Svirezhev & von Bloh, 1997). Here we shall neither be concerned with the biologically induced changes in planetary albedo or the ice-albedo feedback and glaciation.

The greenhouse effect

The parameter T_g in equation (1) expresses the atmospheric greenhouse effect. The effective black-body temperature $T_{\text{eff}} \equiv T - T_g$ of the planet is the temperature at the level in the atmosphere from where the outgoing long-wave radiation is emitted. This level is approximately the height of one optical thickness at the long-wave band seen from space. Depending on greenhouse gasses and clouds, the level of outgoing radiation is around 3 kilometers above the surface. The difference between the black-body temperature T_{eff} and the surface temperature T is the greenhouse warming T_g (or cooling for $T_g < 0$). On Earth the atmosphere is transparent to the sunlight which thus heats the surface. This, in turn, heats the atmosphere from below. The lower atmosphere (the troposphere) thus experiences a positive lapse rate (defined as minus temperature change with height). The lapse rate depends in a complicated way on the static stability, release of latent heat by water vapor condensation and the atmospheric dynamics. In the present climatic conditions the lapse rate is of the order of the dry adiabat of 10 K/km, thus the greenhouse effect on Earth is approximately 3 km×10 K/km=30K. Without the greenhouse effect there would be no liquid water at the surface of the Earth. The atmospheric greenhouse effect, the change in cloudiness and other factors are all expressed through the greenhouse function T_g in the model.

The dependence of T_g on the concentrations of CO_2 and CH_4 is obtained form a one-dimensional radiative-convective model (Kasting *et al.*, 2001). In this model the surface temperature is calculated from the energy balance of the Archean (2.8 Ga BP) with a 20 % reduction of the solar luminosity. A third order polynomial in the logarithms of the atmospheric concentrations of pCO_2 and pCH_4 in a constant 1-bar atmosphere approximates the equilibrium surface temperature. The fitting function is $T = T_{eff} + T_g = k_1x^3 + k_2y^3 + k_3x^2y + k_4xy^2 + k_5x^2 + k_6y^2 + k_7xy + k_8x + k_9y + k_10$, where $x = \log_{10} pCO_2$, $y = \log_{10} pCH_4$ and $k_1, ..., k_{10}) = (-0.75420.2625 - 0.9025 - 0.4325 - 9.10881.3738 - 9.6900 - 28.1973 - 4.8311301.0699)$. The fit is valid within ±1K in the range 10^{-5} bar $< pCO_2 < 10^{-2}$ bar, 10^{-5} bar $< pCO_2 < 10^{-2}$ bar (Kasting *et al.*, 2001).

The logarithmic dependence is a consequence of saturation of absorption bands. Using this, and the reduced solar luminosity of the Archean, T_g is obtained from (1) as a function of the greenhouse gas concentrations.

Modelling the biota

The biota B is modelled by a simple logistic growth equation (Bailey, 1974; Carter & Prince, 1981),

$$\frac{dB}{dt} = U(T)B(1-B) - \lambda B. \tag{2}$$

The function $U(T)$ represents the environmental growth conditions, while λ is a death rate. The growth condition depends on the temperature T, where it is assumed that life would be suppressed by to low temperatures (freezing) or to high temperatures (boiling). We shall assume an optimal temperature T_o and a range of temperatures $[T_o - \Delta T, T_o + \Delta T]$ suitable for life, such that $U(T)$ has a parabolic shape,

$$U(T) = U_0\big(1 - [\frac{T - T_o}{\Delta T}]\big)^2, \text{for } T_o - \Delta T \leq T \leq T_o + \Delta T, \tag{3}$$

and is zero elsewhere. This model, representing the biota in just one variable B is the most crude and simple possible. When modeling the bio-geo-chemistry involved with life, we shall for simplicity assume that the activity of an-aerobic as well as photosynthesizing biota is represented by the variable B.

Modelling the bio-geo-chemistry

Here we shall construct a simple model of the equilibrium concentrations of carbon dioxide CO_2, methane CH_4 and oxygen O_2 in the atmosphere as a function of the surface temperature T and size of the biota B. Through biotic and non-biotic processes these gasses (or the atoms constituting the molecules) are interchanged with the crust, the ocean and other atmospheric gasses. On the time scale of the geochemical cycle the dissolved oceanic CO_2 is in diffusive balance with the free atmospheric CO_2. The biological sink depends on the conditions, such as the acidity of the deep ocean, for sedimentation of dead organic matter. These conditions might have varied but are, due to the relatively young age of the ocean floor, largely unknown through Earth's geological history.

The feedback to the climate from the biota works through various mechanisms. The geochemical CO_2 cycle is influenced by enhanced weathering by vascular plants (Walker *et al.*, 1981; Lenton & von Bloh, 2001). The planetary albedo (Svirezhev & von Bloh, 1997) and the hydrological cycle (Oke, 1992) also depend on the biota.

The atmospheric mixing rations μ_{CO_2}, μ_{CH_4} and μ_{O_2} will be denoted C, M, O for carbon dioxide, methane and molecular oxygen respectively. The atmospheric inventory in these three gasses is given by:

$$\begin{aligned}
\frac{dC}{dt} &= -\big(1-\beta\frac{O+\delta/2}{\delta+O}\big)\frac{C}{\eta+C}N+V-WC+\xi MO^2 &\quad (4)\\
\frac{dM}{dt} &= \beta\frac{\delta/2}{\delta+O}\frac{C}{\eta+C}N+V_M-\xi MO^2 &\quad (5)\\
\frac{dO}{dt} &= \big(1-\beta\frac{O}{\delta+O}\big)\frac{C}{\eta+C}N+S_W-FO-2\xi MO^2 &\quad (6)
\end{aligned}$$

The parameter $N = (NPP) * B$ is the net primary production by the photosynthesizing biota. (NPP) is a scaling factor corresponding to burial of 1 PAL (present atmospheric level) of CO_2 by a global biota in 25 Myr. The fraction $C/(\eta + C)$ models by Michaelis-Menten kinetics the dependence of photosynthesis on the availability of CO_2 in the atmosphere. The parameter

η determines when photosynthesis is limited by the (low) atmospheric CO_2 concentration.

The fraction β of the net primary production is returned to the atmosphere by respiration. This respiration can be either an-aerobic ($2[CH_2O] \rightarrow CO_2 + CH_4$) if the oxygen concentration is below the pasteur point δ or aerobic ($[CH_2O] + O_2 \rightarrow CO_2 + H_2O$) if the oxygen concentration is above the pasteur point. Here we indiscriminantly denote organic molecules in the biota by $[CH_2O]$. The respiration is modelled as two fractions, with net reaction:

$$[\mathrm{CH_2O}] \rightarrow \frac{1}{2}(\mathrm{CO_2}+\mathrm{CH_4})\frac{\delta}{\delta+O}(\text{an-aerobic})+(\mathrm{CO_2}+\mathrm{H_2O}-\mathrm{O_2})\frac{O}{\delta+O}(\text{aerobic}). \tag{7}$$

The photosynthesis and these two respiration processes account for the first terms on the right hand sides of the three equations. Other source terms in the equations are: V: Volcanic outgassing of CO_2. V_M: Outgassing of methane, mainly from mid-oceanic ridges. Other sink terms are: $W \times C$: Weathering of silicate rocks by carbonated rain. For simplicity the weathering rate W is assumed to be independent of temperature. This is only valid if the temperature is not so low that the hydrological cycle is significantly reduced. S_W: Photo-dissociation of water and subsequent escape of hydrogen to space. $F \times O$: Surface oxidation of crustal iron, and oxidation of oceanic dissolved iron. The last terms on the right hand sides represents oxidation of methane in the atmosphere ($CH_4 + 2O_2 \rightarrow CO_2 + H_2O$). Methane oxidation can be parametrized based on various photochemical models (Goldblatt *et al.*, 2006), here we shall simply take ξ to be a constant fitting parameter.

The values for the various parameters are given in the table. The processes represented in equations (4)–(6) are shown in the cartoon in figure (2). The steady state solutions (setting the left hand sides to zero in the equations) determine the composition of the atmosphere as a function of B. This closes the equations of the life-climate model.

The steady states of the life-climate system

The steady states for the system can easily be found by expressing $B(T)$ as a function of T from equation (2), and $T(B)$ as a function of B using equation (1), in both cases taking the left hand side to be zero. The steady states of the system are the crossing points of the two curves, either a single global steady state or two stable points separated by an unstable point. The curve $T(B)$ depends on the solar insolation, thus changing the solar insolation leads to the shift of the $T(B)$ curves (dashed curves in figure 3).

Using the weaker solar insolation back in time (Gough, 1981) the surface temperature as a function of time is shown in figure 4. With the present day level of greenhouse gasses, the Earth's surface would be frozen, even without the ice-albedo feedback around 1.5 Ga BP. Assuming the initial greenhouse concentration at 100 PAL CO_2 and 5 PAL CH_4 the surface is approximately 17K warmer at 4.6 Ga BP (dashed curve) than it would be with the present atmospheric greenhouse gas concentration. Around 3.5 Ga BP the temperature becomes suitable for development of the biota, which in turn keeps the surface temperature remarkably constant around 280 K with a slight increase for the past 500 Ma.

The effect of the photosynthesizing biosphere on the surface temperature is comparable to the albedo control by the biology in the Daisyworld model (Watson & Lovelock, 1983). The habitability range as a function of solar insolation is strongly enlarged and the sensitivity of the surface temperature to the change in insolation is decreased by the greenhouse feedback, $(\partial T/\partial S)_{\text{biotic}} < (\partial T/\partial S)_{\text{abiotic}}$. In this loose sense the bio-climate system can be said to be homeostatic.

The evolution of the chemical constituents in the atmosphere through Earth's history is shown in figure (5). The size of the atmosphere is assumed to have been constant at 1 bar. Around 3.8 Ga BP conditions for early photosynthetic life arose and the oxygen level began increasing, driving down the atmospheric methane. Around 2.4 Ga BP the oxygen level rapidly increased due to the depletion of free iron in the oceans. Banded iron formations prior to 2.4 Ga BP contains perhaps 20 times the present amount of oxygen in the atmosphere. The evolution of oxygen compares well with the constraints imposed by the geological records (red shaded area (Kasting, 1993)). The upper limit of 100 PAL concentration of CO_2 constrained by paleosols (Rye *et al.*, 1995) is indicated by the blue bar. The present atmospheric level of methane is much higher than predicted. This is due to modern biological processes not included in the model.

We can now focus on the response of the system to changes in the external forcing by the solar luminosity. By plotting the steady state temperatures as functions of the solar luminosity, normalized to present days value, (S/S_0), the bifurcation diagram shown in figure 6 is obtained. The thin curve represents the equilibrium surface temperature on the dead planet as a function of the solar insolation. The heavy curve is the equilibrium surface temperature with biology cooling the planet by lowering the CO_2 level of the atmosphere. The dashed curve is unstable steady states. When the solar insolation is increased by 49%, the biosphere is no longer capable of cooling the surface to habitable conditions and the system undergoes a bifurcation into the dead planet state. For solar insolation more than 27 % higher than today the dead planet state is a stable state in contrast to earlier conditions. This means that for a strong perturbation of the atmospheric CO_2 the system would on the early Earth return to the green planet state, while in the future such a perturbation could result in a permanently dead planet.

Summary

Extra-solar Earth like planets are now observed and probably abundant in the galaxy. Conditions for development of life in a form similar to ours, relying on liquid water, defines a relatively narrow window in time span and distance from the star (Kasting & Catling, 2003). If a biosphere can initiate and subsequently influence the climate in a direction favorable for its existence, this range could potentially be much widened. The minimal model presented here of the coupling between the climate and the biosphere captures the overall development of the radiative balance, surface temperature and atmospheric composition through Earth's history. The biology plays a central role in the climate stability. The model includes the geochemical weathering cycle, however, the model suggests that the control by life of the greenhouse warming has a stronger effect in stabilizing the surface temperature. The biosphere increases the habitability as a function of the solar luminosity. In the far future the Earth will due to ever increasing solar luminosity enter a bi-stable regime, where both the living and the dead planet are possible steady states.

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FIGURE CAPTIONS

Fig. 1 The model describes the interdependence of the climate, represented by the surface temperature and the biosphere through the biotic regulation of atmosphric greenhouse gasses.

Fig. 2 The processes represented in equations (4)-(6) determines the atmospheric contents of CO_2, CH_4 and O_2 as functions of the size of the biota B. These processes and the sources and sinks are represented by the arrows. Not all molecules, such as water, involved in the processes are explicitly accounted for in the model. Water is abundant since open oceans have been present for at least the last 4 billion years. The parameters determining the strengths of the different processes, sinks and sources are listed in the table.

Fig. 3 The global biomass as a function of the global mean surface temperature has a maximum at the optimal temperature for habitability. The curves marked by the solar insolation expresses the temperature as a function of the biomass through the greenhouse effect of CO_2 consumed and sedimented by the photosynthetic biosphere. The zero line represents the dead planet. The steady states of the system are found where the curves cross.

Fig. 4 The steady state temperature as function of time, where the solar insolation evolves according to (Gough, 1981). The dashed curves are unstable steady states.

Fig. 5 The equilibrium atmospheric greenhouse gas concentrations depend on the solar insolation. The evolution of concentrations follows from the evolution of the insolation with time (Gough, 1981). Around 3.8 Ga BP conditions for early photosynthetic life arose and the oxygen level began increasing, driving down the atmospheric methane. Around 2.4 Ga BP the oxygen level rapidly increased due to the depletion of free iron in the oceans. Banded iron formations prior to 2.4 Ga BP contains perhaps 20 times the present amount of oxygen in the atmosphere. The horizontal red bar indicates the maximum initial CO_2 concentration constrained by the paleosol record. The red shaded area is the range of O_2 concentrations constrained by the geological records, adapted from (Kasting, 1993).

Fig. 6 The full bifurcation diagram for the bio-climate model as a function of the solar insolation. The heavy curves are the stable states. For $1.2S_0 < S < 1.6S_0$ there exist two stable states, a biotic – and an abiotic state.

Parameter	Value	Description
T_0	290K	Present days mean surface temperature
S_0	1370/4 W/m^2	Solar insolation
NPP	40 bar/Myr	Net primary production [est. 500 g m^{-2} yr^{-1}(Kicklighter *et al.*, 1999)]
V_m	0.002 bar/Myr	Mid-oceanic hot vent CH_4 emission (est. 5-20 Tg/yr (Kandilikar & McRae, 1995))
V	0.03 bar/Myr	Volcanic CO_2 emission (est. $(3.3 \pm 1.1) \times 10^{14}$ g/yr(Holland, 1978))
Sw	0.0041 bar/Myr	Photo dissociation of water/hydrogen escape (0.7-7.0 $\times 10^{12}$ mol/yr (Catling *et al.*, 2001))
W	1.07 Myr^{-1}	Weatherning rate (fitted to 5 PAL prebiotic CH_4)
β	0.99005	Fraction of organic carbon returning by decay (fitted)
ξ	3.7 $\times 10^9$ bar^{-2}Myr^{-1}	Reactivity of methane oxidation (fitted)
Fe	0.129 Myr^{-1}	Reactivity of reducing free iron (fitted)
δ	2 $\times 10^{-3}$ bar	Pasteur point, 0.01 PAL O_2 (Goldblatt *et al.*, 2006)
η	10^{-4} bar	Threshold for photosynthesis

Table 1: Parameter values used for the model of atmospheric chemistry. The atmospheric pressure is assumed constant through time. 1 bar = 1.77×10^{20} mol.

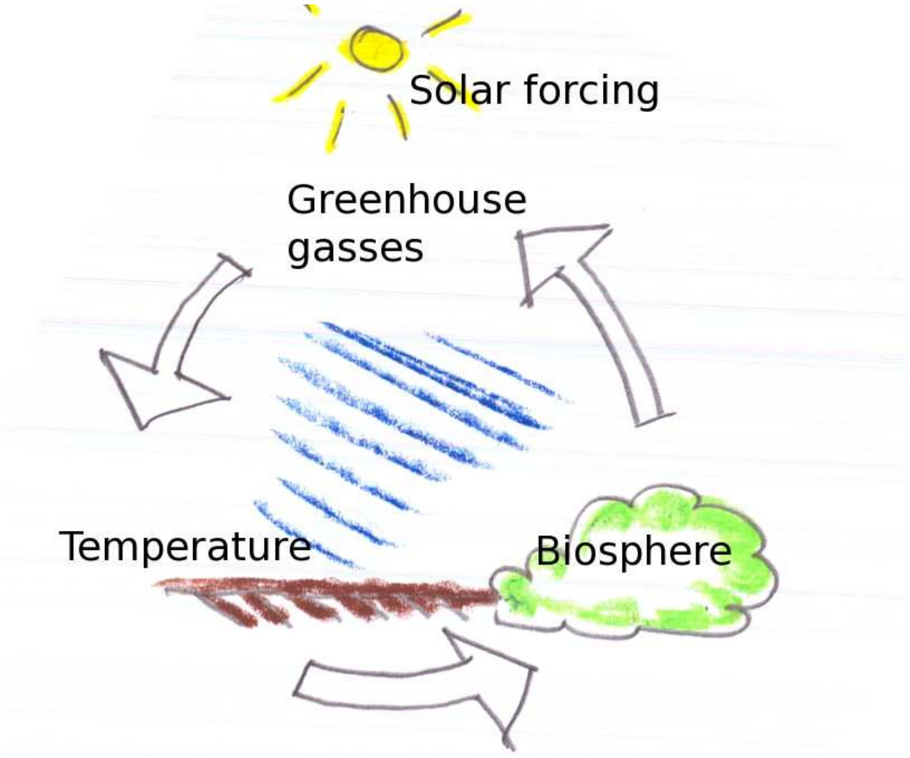


Figure 1:

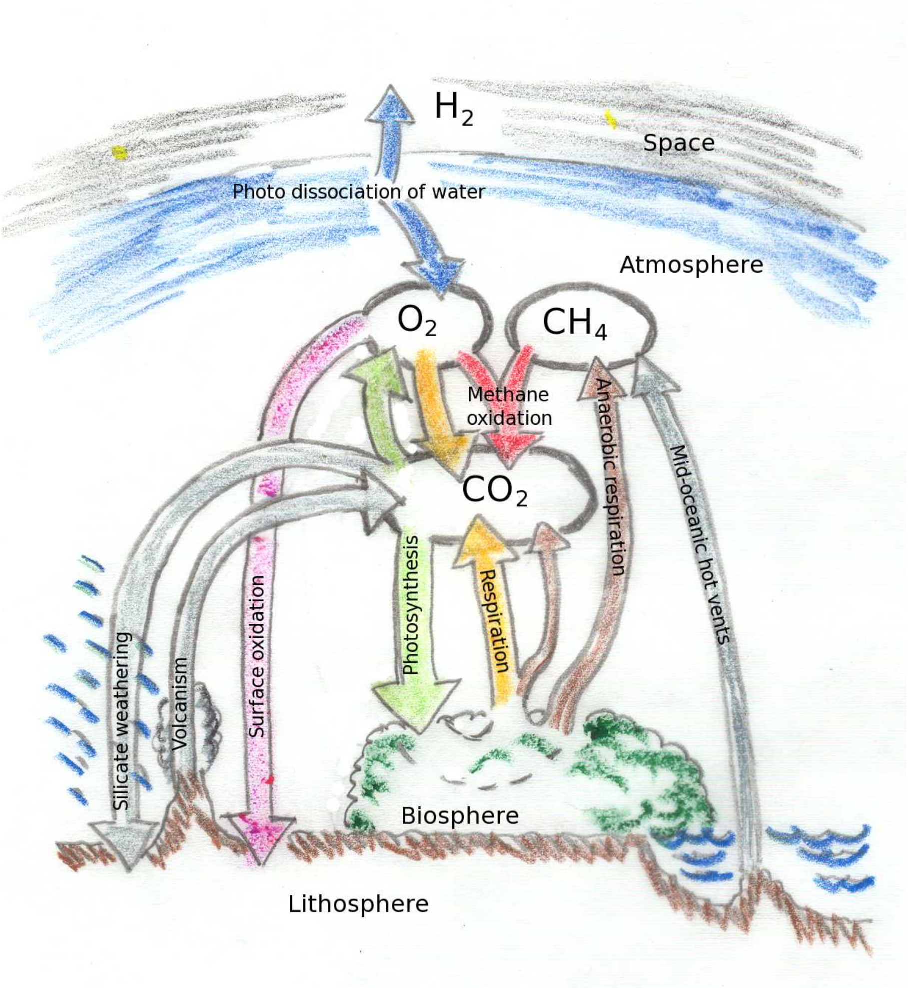
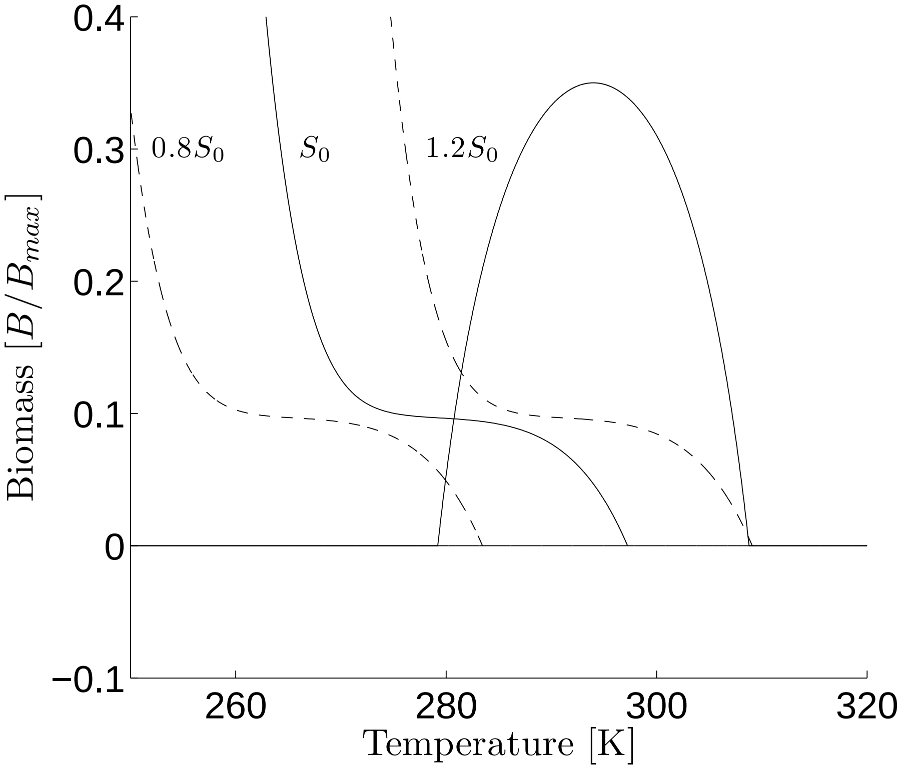

H_2
Space
Photo dissociation of water
Atmosphere
O_2
CH_4
Methane oxidation
CO_2
Anaerobic respiration
Mid-oceanic hot vents
Photosynthesis
Respiration
Surface oxidation
Volcanism
Silicate weathering
Biosphere
Lithosphere

Figure 2:



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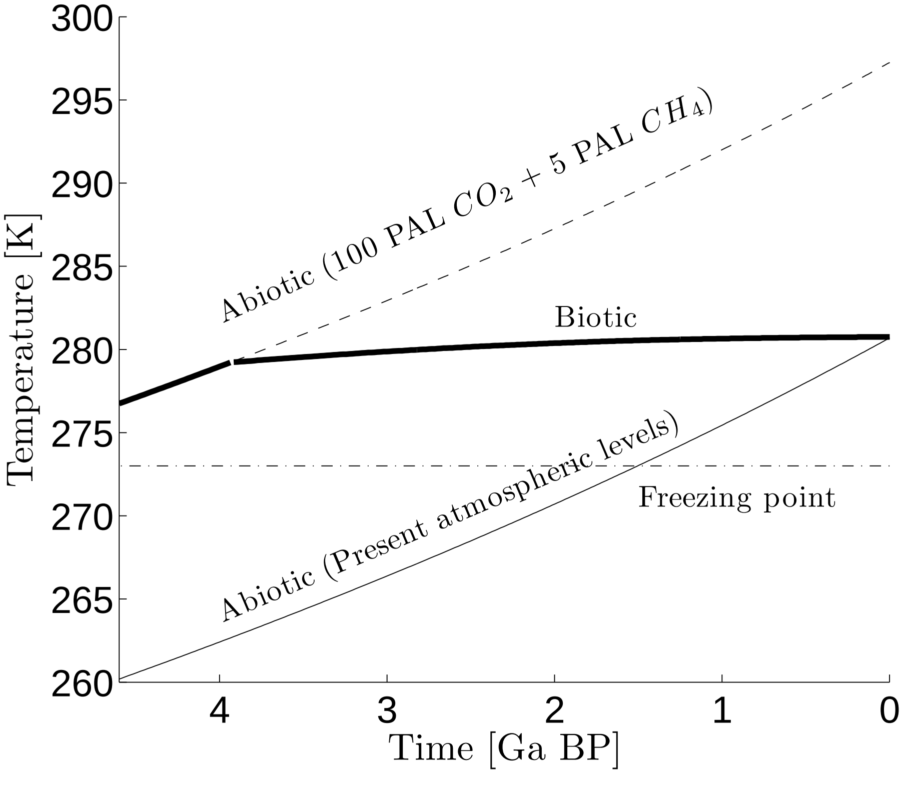
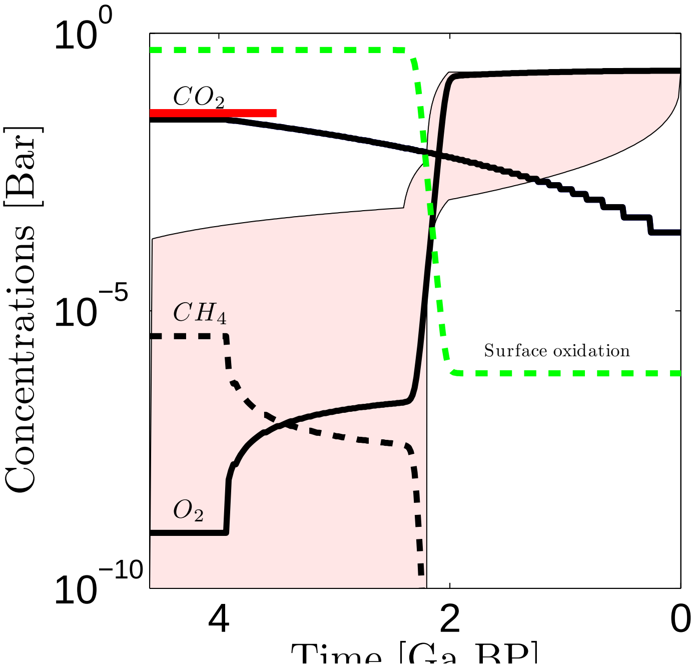


Figure 4:



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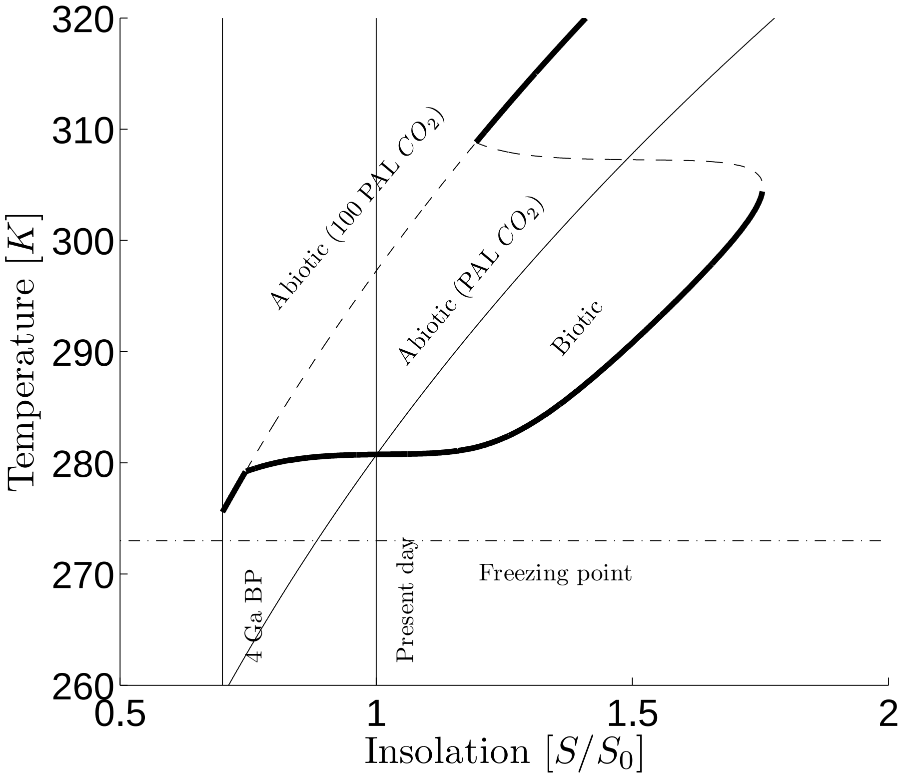

Figure 6: