

UNDERSTANDING LIFE FROM A THERMODYNAMIC EARTH SYSTEM PERSPECTIVE

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ABSTRACT

The Earth's biosphere – the sum of all life – as well as the Earth system itself are two highly dissipative, thermodynamic systems that are driven by low entropy solar radiation and that produce high entropy "waste". Their dissipative activities are constrained by the exchange of energy of different entropies, but also by the material transport processes within these systems. The strength of material transport links the dissipative activity of the Earth system as a whole with the material exchange of the biosphere, such as the exchange of carbon dioxide that is critical to maintain life. Here, I show how the thermodynamic limit of material transport within the Earth's atmosphere imposes a limit to the exchange of carbon dioxide for terrestrial photosynthesis. This limit is not fixed, but is modulated by the biosphere through its effects on surface absorption and on the atmospheric composition. These effects are illustrated by a simple model of atmospheric transport and biotic activity. This tight interplay between the limits of physical transport on biotic activity and the effects of biotic activity on the Earth system emphasizes the need to approach and understand life in its Earth system context in which thermodynamics defines the limits of the overall dissipative activities of both systems.

THERMODYNAMICS OF LIFE ON EARTH

It is well recognized that life can be seen as a thermodynamic dissipative process that is fueled by low entropy energy and produces high entropy waste [1; 2]. Boltzmann [1] already described in the late 19th century that the

general struggle for existence of living organisms is therefore not a struggle for the basic materials – these materials are abundantly available for organisms in air, water and soil – ... but a struggle for entropy, which through the transformations of energy from the hot sun to the cold Earth becomes available.

Likewise, the Earth system as a whole is a thermodynamic, dissipative system that is fueled by the absorption of low entropy solar radiation and produces high entropy terrestrial radiation. This correspondence was noted by Lovelock [3; 4], who popularized the notion that the Earth is like a living organism. This comparison was not made on the basis of a biological definition of life, but rather from the recognition that both, life and the Earth system are highly dissipative systems that are maintained in states far from thermodynamic equilibrium.

Yet, the notion that life and the Earth system are dissipative systems by itself does not provide a constraint on how dissipative these systems are. Yet, thermodynamics provides information about limits as well and the factors that shape these limits. A prime example for such a limit is the Carnot limit of a heat engine, which describes how much heat can be converted into mechanical work by the engine which can later be dissipated. To evaluate such limits for life and for the Earth system, we need to view these systems in terms of their environmental setting. The Earth system is driven by radiative exchange,

so the question regarding the limit of dissipative activity of the Earth system relates to the thermodynamic limit of how much free energy can be generated from the radiative forcing. Life is embedded within the functioning of the Earth system, and it is subjected to thermodynamic limits regarding the conversion of sunlight into chemical energy, but also to limits regarding the transport and exchange of the basic materials that are required during the process of chemical energy generation and that are taken up from (and released to) the abiotic environment. The chemical transformation associated with life leaves an imprint in the environment, most notably in terms of the atmospheric composition, which has likely changed drastically during the history of the Earth system due to life [5; 6] and which would affect the radiative exchanges within the system. Hence, we gain a view of life and the Earth system as two, closely connected thermodynamic systems with reciprocal roles, with Earth system functioning shaping a habitable environment that favors life and with the effects of life shaping the Earth's environment.

This interplay between the Earth system and the biosphere – the sum of all life – is illustrated in Fig. 1. Both, the abiotic processes of the Earth system as well as the biosphere are driven mostly by the absorption of low entropy solar radiation. Solar radiation is absorbed at the Earth's surface, and differences in absorption and emission provide the gradients to drive abiotic processes, such as the generation of motion or the evaporation of water which represent the dissipative activity of the Earth system (arrow A in Fig. 1). Most of the biotic activity is driven directly or indirectly by photosynthesis, which utilizes a fraction of the absorbed solar radiation in converting carbon dioxide into carbohydrates (arrow B in Fig. 1). In both cases, the absorbed solar energy is eventually reemitted to space, but it is emitted at a much lower radiative temperature, so that the emitted ra-

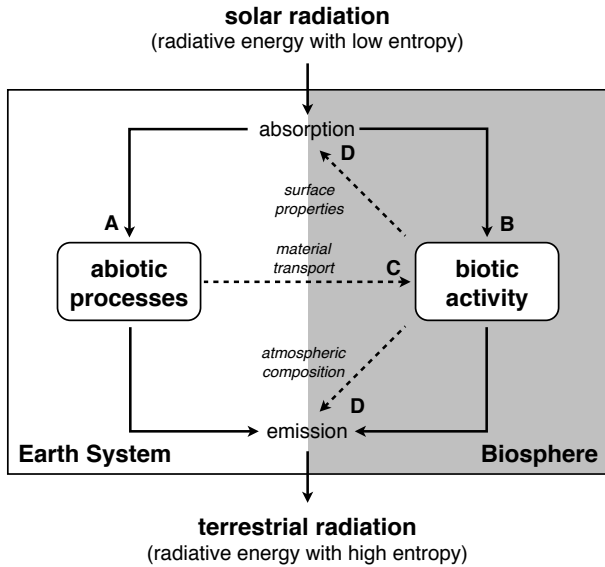


Figure 1. Schematic diagram to understand life in the thermodynamic context of the whole Earth system.

diation has a much higher radiative entropy. This difference in radiative temperatures between the absorbed solar radiation and the emitted terrestrial radiation provides the difference in entropy that fuels both, the Earth system and its biosphere.

To link biotic activity and Earth system functioning, we first note that biotic activity also requires basic building materials, particularly carbon dioxide from which organic biomass is being formed from. Carbon dioxide is taken up from the atmosphere (for the terrestrial biosphere, which I will focus on here), or from the ocean (for the marine biosphere), and in both cases the physical environment provides the means to transport carbon dioxide to the biosphere (arrow C). The ability to transport depends on the intensity by which motion can be generated within the Earth system from the planetary forcing of solar radiation, and this generation rate is thermodynamically constrained. Hence, the material supply for biotic activity is one factor by which the Earth system imposes a constraint on biotic activity (among other factors, such as temperature or water availability on land).

The arrows D in Fig. 1 describe the effects of biotic activity on the Earth system by the mass exchange of basic materials (as mentioned above). Two aspects of this modification directly relate to the radiative forcing of the Earth system. The first aspect relates to the presence of photosynthetic tissues that typically increase the absorption of solar radiation at the surface, which can, for instance, easily be noted on land where the presence of vegetation darkens the surface. The second aspect is more subtle and involves alterations of the atmospheric composition, for instance in terms of the concentration of carbon dioxide, methane, and molecular oxygen. The atmospheric composition alters the radiative properties of the atmosphere in terms of the concentration of greenhouse gases and thereby affects the transfer of terrestrial radiation and the ability of the system to emit radiation. Hence, the two effects associated with arrows D have quite profound effects for the physical functioning of the Earth system.

The goal of this contribution is to illustrate the interplay between life and Earth shown in Fig. 1 with a simple, yet quantitative model of the Earth system and biotic activity and to quantify the thermodynamic limits of both systems as well as their

coupling. I first illustrate the thermodynamic limit on mass exchange between the surface and the atmosphere associated with convection. This intensity of mass exchange is then related to the limitation imposed by the environment on the transport of basic materials for photosynthesis. The consequences of biotic activity are then discussed in terms of altering the atmospheric composition, which in turn affects the strength of the atmospheric greenhouse effect and the thermodynamic limits. The implications for the understanding of life in a thermodynamic Earth system context are then summarized.

TRANSPORT LIMITS IN THE EARTH SYSTEM

The transport and exchange of mass within the Earth system is strongly constrained by thermodynamic limits. To demonstrate these limits, I set up a simple model of atmospheric convection in the following, which is based on [7; 8]. This model considers the Earth's surface with a temperature T_s and the atmosphere with a temperature T_a as a thermodynamic system made up of two heat reservoirs. The system is forced by the heating associated with the absorption of solar radiation at the surface, J_{sw} , and by the cooling associated with the emission of terrestrial radiation from the atmosphere, J_{lw} . Its steady state is considered in which $J_{sw} = J_{lw}$. The surface and the atmosphere are coupled by a flux of radiative exchange, $J_{s,a}$, as well as the sensible and latent heat fluxes, J_{sh} and J_{lh} . These latter fluxes are directly linked to atmospheric motion and thus to the ability of the atmosphere to transport mass, with the latent heat flux linked to the strength of the hydrologic cycle.

To derive the thermodynamic constraints of material transport, we consider the limit to the rate by which kinetic energy can be generated within the atmosphere. This rate, G , is set by the Carnot limit for dry convection which is associated with the sensible heat flux, J_{sh} , and the temperature difference, $T_s - T_a$:

$$G = J_{sh} \cdot \frac{T_s - T_a}{T_s} \quad (1)$$

The two terms, J_{sh} and $T_s - T_a$, are constrained by the energy balances of the system, with a greater flux J_{sh} corresponding to a smaller temperature difference $T_s - T_a$. This trade-off is derived from the explicit consideration of the energy balances. Using this energy balance constraint then yields a maximum power limit G_{max} that is associated with an optimum mass exchange between the surface and the atmosphere that is characterized by an optimum vertical exchange velocity w_{opt} .

In the following, several simplifying assumptions are being made to derive a relatively simple, but realistic analytical solution. A wet surface is considered, i.e. that the evaporation rate is not limited by water availability. The atmosphere is assumed to absorb all of the emitted terrestrial radiation from the surface. As will be seen below, these considerations are quite reasonable for present-day conditions and yield estimates by the model that compare well with observations.

Energy balance constraints

The surface energy balance of the system is given by

$$0 = J_{sw} - J_{s,a} - J_{sh} - J_{lh} \quad (2)$$

The corresponding energy balance of the atmosphere is given by

$$0 = J_{s,a} + J_{sh} + J_{lh} - J_{lw} \quad (3)$$

In these equations, J_{sw} represents the forcing of the system, and the steady state requires that $J_{lw} = J_{sw}$. Since the radiative temperature is fixed by the global energy balance, $J_{sw} = \sigma T_a^4$ (with σ being the Stefan-Boltzmann constant), the atmospheric temperature is fixed at $T_a = (J_{sw}/\sigma)^{1/4}$. The radiative exchange flux between the surface and the atmosphere is expressed in a linearized approximation by $J_{s,a} = k_r(T_s - T_a)$ with $k_r = 4\sigma T_s^3/(1 + 0.75\tau)$ and τ being the longwave optical depth of the atmosphere. The sensible heat flux is expressed as $J_{sh} = c_p \rho w (T_s - T_a)$ with heat capacity c_p and air density ρ . The latent heat flux for an open water surface is written as $J_{lh} = \lambda w (q_s - q_a) \approx s/\gamma \cdot J_{sh}$ with λ being the latent heat of vaporization, q_s and q_a being the specific humidities of near surface and atmospheric air, $s = de_{sat}/dT \approx s_0 \exp[19.83 - 5417/T_s]/T_s^2$ is the slope of the saturation vapor pressure evaluated at the reference temperature T_s (with $s_0 \approx 3.3 \cdot 10^6$ Pa K), and γ is the psychrometric constant. These formulations are typical formulations of atmospheric heat fluxes in meteorology, and the details of the formulations can be found in [8].

Maximum power limit

The expression for the Carnot limit with these formulations of the heat fluxes depends on the forcing, J_{sw} , a series of physical and radiative parameters (such as heat capacity, air density, the psychrometric constant, the slope of the saturation vapor pressure curve, and optical thickness), and on the vertical exchange velocity within the atmosphere, w , which is a yet unconstrained variable:

$$G = \frac{c_p \rho w}{T_s (k_r + c_p \rho w (\gamma + s) / \gamma)^2} \cdot J_{sw}^2 \quad (4)$$

We can constrain the value of w by assuming that the generation rate G is maximized with respect to w , that is, that the generation of motion is maximized and operates at the thermodynamic limit within the system. When we neglect the slight dependence of T_s (because variations in T_s are relatively small compared to the mean), we can derive an analytic expression for the maximum generation rate, G_{max} :

$$G_{max} = \frac{\gamma}{\gamma + s} \cdot \frac{J_{sw}^2}{2k_r T_s} \quad (5)$$

with associated partitioning of heat fluxes of

$$J_{s,a,opt} = \frac{J_{sw}}{2} \quad J_{sh,opt} = \frac{\gamma}{\gamma + s} \cdot \frac{J_{sw}}{2} \quad J_{lh,opt} = \frac{s}{\gamma + s} \cdot \frac{J_{sw}}{2} \quad (6)$$

Note that this state of maximum power associated with convection is closely related to a state of Maximum Entropy Production (MEP), which is a general hypothesis that complex thermodynamic systems are maintained in steady states at which entropy production is maximized ([9; 10; 11; 12; 13; 14; 15]). The generation of kinetic energy equals its frictional dissipation

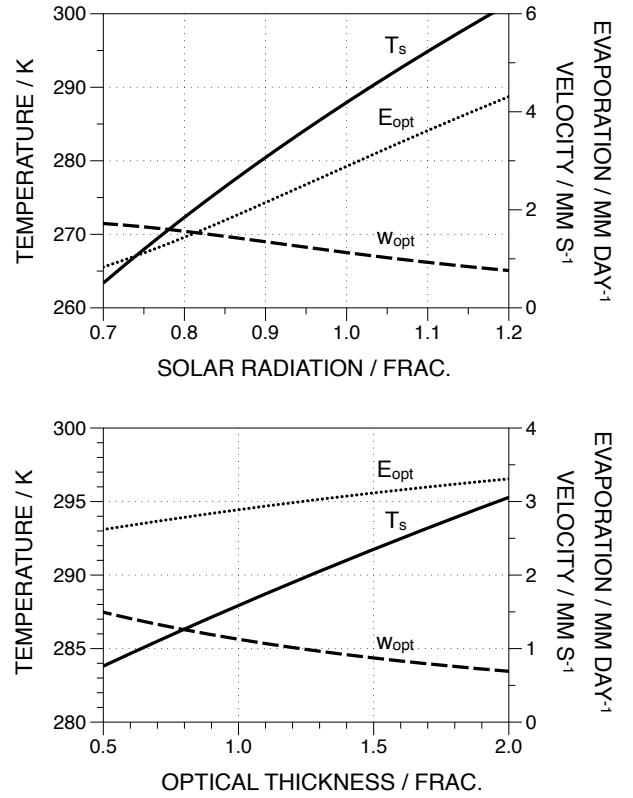


Figure 2. Sensitivity of the thermodynamic limit of convective exchange within the atmosphere, represented by the optimum exchange velocity w_{opt} , to radiative forcing and the associated surface climate in terms of surface temperature T_s and evaporation E_{opt} . The panels show the sensitivities to (top) solar radiative heating, J_{sw} , and (bottom) optical thickness, τ , both expressed as a fraction of today's reference value.

in steady state, i.e. $G = D$, so that a maximization of the generation rate then corresponds to a maximization of dissipation. If this dissipation occurs at the cold, atmospheric temperature, T_a , then the entropy production, σ_{sh} , associated with this frictional dissipation is given by:

$$\sigma_{sh} = \frac{G}{T_a} = J_{sh} \cdot \left(\frac{1}{T_a} - \frac{1}{T_s} \right) \quad (7)$$

using $G = D$ and eqn. 1 from above. Since T_a is fixed by the planetary energy balance with $T_a = (J_{sw}/\sigma)^{1/4}$, the maximization of G corresponds to a maximization of σ_{sh} .

This state of maximum power (and maximum dissipation) relates back to the motivation of this contribution (arrow A in Fig. 1) in that it is this state which characterizes the maximum dissipative activity of the Earth system in terms of atmospheric motion.

Climate sensitivity at maximum power

The properties at a state of maximum convective power can now be associated with climatic conditions at the surface. The above maximization is associated with an optimum vertical exchange velocity w_{opt} at the surface-atmosphere interface of

$$w_{opt} = \frac{\gamma}{\gamma + s} \cdot \frac{k_r}{c_p \rho} \quad (8)$$

With this maximization, the temperature difference is set, so that the surface temperature, T_s , is obtained from the energy balance and the atmospheric temperature, T_a :

$$T_{s,opt} = T_a + \frac{J_{sw}}{2k_r} = \left(\frac{J_{sw}}{\sigma} \right)^{1/4} + \frac{J_{sw}}{2k_r} \quad (9)$$

The associated strength of the hydrologic cycle, expressed by the flux of evaporation E (or precipitation, since $P = E$) is given by:

$$E_{opt} = \frac{s}{\gamma + s} \cdot \frac{J_{sw}}{2\lambda} \quad (10)$$

When these properties are evaluated for present-day conditions with $J_{sw} = 240 \text{ W m}^{-2}$ and $\tau = 0.65$, these expressions yield values of $T_s = 288 \text{ K}$, $w_{opt} = 1.1 \text{ mm s}^{-1}$, and $E_{opt} = 2.9 \text{ mm d}^{-1}$, which are very close to observed magnitudes of $w_{obs} \approx 1 \text{ mm s}^{-1}$ [16] and $E_{obs} \approx 2.7 \text{ mm d}^{-1}$ [17].

The sensitivity of these estimates to changes in the radiative forcing is shown in Fig. 2. The upper plot in the figure shows the sensitivity to absorbed solar radiation, J_{sw} , which characterizes the strength of the forcing of the system. This forcing is affected by the luminosity of the Sun, but also to some extent by the reflectivity of the Earth's atmosphere (e.g. clouds) and the surface (e.g. ice, vegetation, water), with the latter aspects not dealt with here. The lower plot in Fig. 2 shows the sensitivity to the optical thickness, τ , which describes the strength of the atmospheric greenhouse effect. This property does not affect the rate of surface heating, but rather the rate by which the surface cools through emission of terrestrial radiation.

The sensitivities of the estimates to these two radiative properties are qualitatively similar and consistent with sensitivities derived from complex climate models. Greater values of J_{sw} and τ both result in warmer surface temperatures, T_s , but for different reasons. In the first case, the warmer temperature results from a stronger solar forcing, while in the latter case, it results from a reduced cooling rate due to a stronger atmospheric greenhouse effect. Greater values of J_{sw} and τ also result in enhanced evaporation, E_{opt} , and lower values of the optimum vertical exchange velocity, w_{opt} . This lower value of w_{opt} results from the fact that at higher temperatures, s obtains a greater value, so that less vertical motion is needed to accomplish the turbulent heat exchange.

To summarize this section, we derived a Carnot-type limit for convective motion within the atmosphere from the radiative forcing and evaluated the climatic conditions associated with this maximum convective transport state. The maximum results from the strong interaction between the convective heat fluxes of sensible and latent heat, $J_{sh} + J_{lh}$, and the driving temperature difference, $T_s - T_a$, which decreases with greater convective heat fluxes due to the energy balance constraints. This trade-off is the same trade-off that is involved in studies of Maximum Entropy Production (MEP), although here this limit is interpreted by more conventional means in terms of the Carnot limit to mechanical power. This maximum convection state yields a realistic representation of the climate and characterizes the upper thermodynamic limit on mass exchange between the surface and the atmosphere. Next, this limit is evaluated regarding its implication for biotic activity at the surface.

TRANSPORT LIMITS AND BIOTIC ACTIVITY

The environmental limits on the photosynthetic rate are now being considered, as photosynthesis acts as the main driver for biotic activity on Earth. Two environmental limits are considered that directly relate to the processes considered in the previous section: the availability of light at the surface to drive the photochemistry associated with photosynthesis (arrow B in Fig. 1), and the ability of the atmosphere to transport carbon dioxide to the surface at which photosynthesis takes place (arrow C in Fig. 1). These two constraints are formulated in terms of a light-limited rate, $J_{bio,sw}$, and a flux-limited rate, J_{bio,CO_2} , of photosynthesis.

The light-limited rate, $J_{bio,sw}$, is linked to the absorption of solar radiation at the surface and is expressed as

$$J_{bio,sw} = \varepsilon \cdot J_{sw} \quad (11)$$

where ε is the light use efficiency. Since about 55% of solar radiation is photosynthetically active radiation, and it requires about 10 photons of wavelengths of 580 and 600nm to fix one molecule of carbon, the value of ε should be around $\varepsilon = 3.910^{-6} \text{ gC J}^{-1}$. For present-day conditions with $J_{sw} = 240 \text{ W m}^{-2}$, this yields a light-limited rate of about $J_{bio,sw} = 77 \mu\text{mol m}^{-2} \text{ s}^{-1}$. This rate is quite a bit higher than the observed maximum photosynthetic rate of around $50 \mu\text{mol m}^{-2} \text{ s}^{-1}$ [18] so that, overall, it is not the availability of light that limits biotic activity.

The flux-limited rate, J_{bio,CO_2} , reflects the limitation due to the atmospheric exchange of carbon dioxide between the atmosphere and the surface. It is expressed in terms of the vertical exchange velocity, w_{opt} , as well as the difference in carbon dioxide concentration

$$J_{bio,CO_2} = \rho w_{opt} \cdot (pCO_{2,a} - pCO_{2,s}) \quad (12)$$

where $\rho = 1.2 \text{ kg m}^{-3}$ is the air density and $pCO_{2,a}$ and $pCO_{2,s}$ are the mixing ratios of carbon dioxide within the atmosphere and at the surface. Since the CO_2 mixing ratio within the air space of leaves is about 70% of the atmospheric concentration, a value of $pCO_{2,s} = 0.7 pCO_{2,a}$ is used here, with a value of $pCO_{2,a} \approx 390 \text{ ppmv}$. With $w_{opt} = 1.1 \text{ mm s}^{-1}$, this yields a value of J_{bio,CO_2} for present day conditions of about $20 \mu\text{mol m}^{-2} \text{ s}^{-1}$. This is noticeably smaller than the observed maximum photosynthetic rate stated above, emphasizing the importance of this transport limitation to photosynthesis.

The sensitivity of both limitations to the radiative properties of absorbed solar radiation, J_{sw} , and longwave optical thickness, τ , are shown in Fig. 3. The light-limited rate increases with absorbed solar radiation, while it is insensitive to changes in τ . In contrast, the flux-limited rate decreases with both, J_{sw} and τ , due to the lower value of w_{opt} .

This example is, of course, formulated in a highly simplified way. There are several ways by which the biota, particularly vegetation on land, can alter the transport limit to some extent, thereby alleviating this constraint. For instance, vegetation can reduce the rate of transpiration by stomatal control, which would enhance the vertical exchange velocity by reducing the effect of s in the expression of w_{opt} (not shown here). A greater value of w_{opt} would then raise the flux-limited rate and allow for a greater rate of photosynthesis.

Nevertheless, the example demonstrates that atmospheric transport and the associated flux limitation for photosynthesis

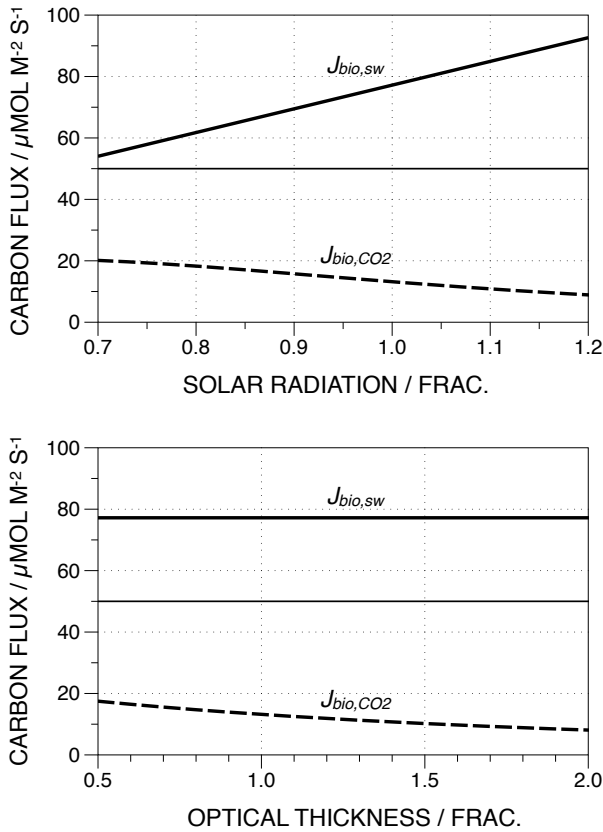


Figure 3. Sensitivity of the light-limited and flux-limited rates of photosynthesis ($J_{bio,sw}$, solid line and $J_{bio,CO2}$, dashed line) to (top) solar radiative absorption, J_{sw} , and (bottom) optical thickness, τ , both expressed as a fraction of today's reference value. The thin horizontal line indicates roughly the maximum observed rate of photosynthesis.

is more limiting than the availability of light. This limitation forms one aspect of the coupling between the dynamics within the abiotic part of the Earth system and the activity of the biosphere, as represented by arrow C in Fig. 1.

BIOTIC EFFECTS ON THE EARTH SYSTEM

The effects of biotic activity on the Earth system as shown by arrows D in Fig. 1 affect the values of J_{sw} through enhanced absorption by biomass and of τ through changes in atmospheric composition. Even though the first effect plays a large role on land, its global effect is relatively small, so that it is not considered here. The following scenario focuses on the second effect that involves changes in τ that are taken here as a result of biotic activity.

The scenario that is considered here is placed in the context of Earth system history. Geological indicators suggest that the Earth maintained an ice-free state and maintained surface temperatures within a relatively narrow range through most of its history although the sun was a lot fainter in the past, with about 70% of today's luminosity at 4.5 billion years ago. One common "solution" to this discrepancy is that the concentration of greenhouse gases may have been substantially higher in the past [19; 20], with changes in greenhouse gas concentrations being attributed to changes in biotic activity.

We now consider such a scenario in the context of the simple model developed here. The surface temperature T_s is prescribed to its present-day value, and this condition is used to derive the value of τ under the assumption of maximized convective

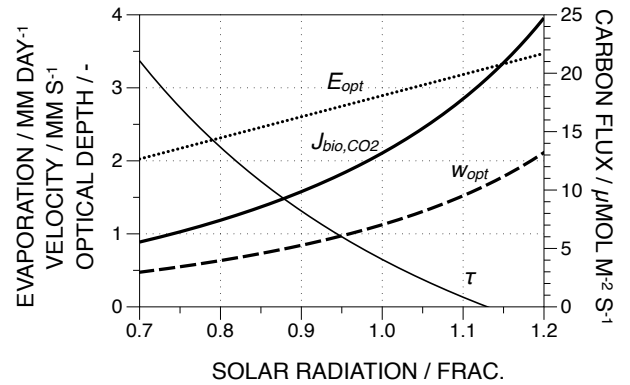


Figure 4. Implied changes in optical depth τ (thin solid line), evaporation E_{opt} (dotted line), vertical exchange velocity w_{opt} (dashed line), and flux-limited rate of carbon exchange, $J_{bio,CO2}$ (solid line) when a fixed surface temperature of $T_s = 288$ K is prescribed for the given variation in absorbed solar radiation, J_{sw} .

exchange. The sensitivity of this setting is then evaluated to changes in J_{sw} , which is shown in Fig. 4. The figure shows the decrease in τ with an increase in J_{sw} , which is consistent with the previous studies that argued for a stronger greenhouse effect to compensate for the lower values of solar luminosity in the past. The figure also shows a strengthening of the hydrologic cycle, shown by E_{opt} , and a stronger vertical exchange, w_{opt} , with greater values of J_{sw} , which result in a less-limiting rate $J_{bio,CO2}$.

This sensitivity shown in Fig. 4 can directly be related to the equations described above. The prescribed surface temperature of $T_s = 288$ K requires a smaller value of k_r for lower values of J_{sw} , which can be seen in eqn. 9. This smaller value of k_r is achieved by a greater value of τ . Hence, the value of τ is reduced with increased values of J_{sw} under the constraint of the prescribed surface temperature. The evaporation rate, E_{opt} , is directly proportional to J_{sw} (eqn. 10), so that the strength of the hydrologic cycle increases proportionally with the rate of absorption of solar radiation, J_{sw} . The increase in w_{opt} with J_{sw} reflects the increased value of k_r (cf. eqn. 8), which is due to the lower values of τ that are needed to maintain the prescribed surface temperature. This increase in w_{opt} then leads to the increase in $J_{bio,CO2}$ with J_{sw} .

Even though the sensitivity and the model considered here is highly simplified, it illustrates the important point that the dissipative activities of the Earth system and of the biosphere are not externally determined, but merely constrained. In the model, this constraint is represented by the magnitude of absorption of solar radiation, J_{sw} . Furthermore, these two systems strongly interact, with the radiative forcing and the value of τ affecting the flux-limited rate of photosynthesis, while greater values of biotic activity could result in a reduction of greenhouse gases, which could reduce τ . Hence, the interaction between life and the Earth system is likely to affect the magnitudes of their respective dissipative behaviors.

CONCLUSIONS

To conclude this study, the results presented here suggest that it is not primarily the struggle for light that limits the dissipative activity of the Earth's biosphere, but rather the ability to exchange materials. To formulate the essence of this contribution in a similar way to Boltzmann's quote that was presented at

the beginning of this paper, this study would suggest that the

general struggle for existence of living organisms is therefore not a struggle for light – this is abundantly available at the surface – but a struggle for transport of the basic materials, which through the transformations of energy from the hot sun to the cold Earth becomes available.

This perspective intimately links the abiotic transport characteristics of the Earth system to the essential resource requirements for life. If we want to better understand the role of life, we would need to view it as a component of the Earth system that is deeply embedded in its function and that is subjected to limits that are not just related to direct energy conversions, but also to environmental transport limitations. These limits are, however, not fixed, but are ameliorated by the consequences of life. Hence, this would seem to require a thermodynamic Earth system perspective to understand the role and consequences of life on Earth.

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