Thickness of tropical ice and photosynthesis on a snowball Earth

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Abstract. On a completely ice-covered "snowball" Earth the thickness of ice in the tropical regions would be limited by the sunlight penetrating into the ice cover and by the latent heat flux generated by freezing at the ice bottom — the freezing rate would balance the sublimation rate from the top of the ice cover. Heat transfer models of the perennially ice-covered Antarctic dry valley lakes applied to the snowball Earth indicate that the tropical ice cover would have a thickness of 10 m or less with a corresponding transmissivity of > 0.1%. This light level is adequate for photosynthesis and could explain the survival of the eukaryotic algae.

Introduction

Based on geological evidence it has been suggested [Kirschvink, 1992; Hoffman et al., 1998; Hoffman and Schrag, 2000] that the Earth was completely ice-covered during the Neoproterozoic (about 550-829 Myr ago). Negative carbon isotope anomalies in carbonate rocks are consistent with a virtual elimination of oceanic photosynthesis due to a thick covering of ice [Hoffman et al., 1998; Hoffman and Schrag, 2000]. It has been previously assumed that the ice cover would be of order a kilometer thick, limited only by geothermal heat flow [Gaidos et al., 1999]. Climate models [Caldeira and Kasting, 1992] and geological data [Hoffman et al., 1998; Hoffman and Schrag, 2000] both suggest that ice covered conditions would persist for about 5 Myrs. Given these conditions it is puzzling therefore that many eukaryotic algae (including red, green, and chromophytic algae) survived this event [Knoll, 1992; Williams et al., 1998]. Gaidos et al. [1999] discuss the challenge of survival under a kilometer thick ice cover.

It may be possible to resolve this paradox if in the tropical regions the ice cover was not nearly as thick as supposed. Evidence for this comes from the Antarctic dry valley lakes [Parker et al., 1982; McKay et al., 1985]. Here the mean annual temperature is -20°C, and the depth to geothermal melting is over 300 m. Yet the thickness of the ice cover on large closed lakes in these valleys is only about 5 m [McKay et al., 1985]. The explanation lies in the fact that the main heat flow through these ice covers is due to sunlight penetrating the ice cover and to latent heat released by freezing at the bottom of the ice cover. Geothermal heat is less than either of these terms by a factor of 50 [McKay et al., 1985]. In steady state the freezing rate balances ablation which removes about 35 cm yr⁻¹ from the lake ice surfaces [Hendersen et al., 1965; McKay et al., 1985; Clow et al., 1988]. This mass loss is balanced by meltwater inflow when air temperatures do rise above freezing for brief periods [Clow et al., 1988].

Assuming that radical changes in the obliquity of the Earth were not the cause of the Neoproterozoic glaciations, then, even given that the solar constant was about 6% less during this time, the average sunlight incident on the surface of the tropical ice would still be about 4 times that reaching the Antarctic dry valley lakes. Sunlight and latent heat release would also limit the thickness of the tropical ice on a snowball Earth. If radical obliquity changes were involved then the presence of equatorial glaciations is not indicative of global ice cover and the paradox of the Neoproterozoic glaciations is removed [Williams et al., 1998].

If the equatorial regions of the snowball Earth did receive more radiation than the polar caps then some transport of heat from equator to pole would result. The circulation driven by this equator-to-pole transport might be associated with precipitation at the equator itself but in the tropical desert zones, caused by descending motion in the general circulation, there would be a net sublimation of ice from the sea surface. In steady state the ice thickness is constant and therefore the removal of ice from the top of the tropical ice cover is balanced by the freezing of water at the bottom of the ice cover. The latent heat removal necessary for this freezing to occur must be accomplished by a steeper thermal gradient in the ice cover. The requirement to carry out the solar energy deposited within the ice cover as well as the latent heat removal necessary for freezing to balance ablation determines the temperature gradient in the ice cover and limits the thickness of the ice.

Energy balance model

The energy balance equation for a thick ice cover can be solved for the equilibrium thickness of the ice [McKay et al., 1985]. Energy inputs to the system include geothermal heat flow, solar energy and latent heat released when water freezes at the bottom of the cover. Assuming steady state conditions and averaging over the year the only loss of energy from the ice cover is by conduction upward from the warm ice-water interface to the cold atmosphere, and so:

\[ k \frac{dT}{dz} = S(z) + L + F_g \]  

where \( k \) is the thermal conductivity, \( T \) is the mean annual temperature, \( z \) is the depth below the ice surface, \( S(z) \) is the net solar flux penetrating below level \( z \), \( F_g \) is the geothermal heat flux, and \( L \) is the rate of latent heat release given by \( L = v \rho \lambda \), where \( v \) is the abla-
tion rate, \( \rho \) is the density of ice, and \( l \) is the latent heat of sublimation of ice. Energy transport by the upward motion of the ice has been ignored. Following McKay et al. [1985] the attenuation of sunlight in the ice cover is expressed with an exponential. Thus,
\[
S(z) = (1-a)(1-r)S_0 \exp(-Z/h),
\]
where \( a \) is the albedo of the ice averaged over the solar spectrum, \( r \) represents a fraction of the ice surface that is covered with dark absorbing material such as sand or silt, \( S_0 \) is the solar radiation, and \( h \) is the effective attenuation coefficient which includes the cosine of the averaged solar zenith angle. The thermal conductivity of ice is approximated by \( k = b/T - c \), where \( b \) and \( c \) are constants given by 780 W m\(^{-1}\) K\(^{-1}\) and 0.615 W m\(^{-1}\) K\(^{-1}\), respectively. Using these relations and integrating Eq. 1 from the ice-water interface to the surface gives [McKay et al., 1985]:

\[
\frac{Z}{vpl + F_g} - \frac{c(T_s - T_e) - b \ln(T_e/T_o)}{vpl + F_g} = S_0 h(1-a)(1-r)(1-\exp(-Z/h))
\]

where \( T_e \) is the temperature of ice in equilibrium with sea water, \( T_o \) is the yearly averaged temperature at the surface (both in K), and \( Z \) is the equilibrium thickness of the ice cover. For very cold conditions \((T_e < -20^\circ C)\) the non-constancy of the thermal conductivity is significant.

Results

In this paper we are interested in determining if the ice cover was too thick to allow photosynthesis in the water below. The transmissivity of the photosynthetically active radiation (PAR, 400 - 700 nm) through the ice cover is shown in Figure 1. This computation is based on the multi-layer radiative scattering model of McKay et al. [1994] developed for the Antarctic dry valley lakes. To model ice covers thicker than those found in the dry valleys additional thickness was added to the bottom of the ice cover with optical properties assumed to be the same as the lowest layer of the ice cover. This layer has formed by slow freezing of water at the ice-water interface and this ice is relatively clear [McKay et al., 1994]. A low ablation/freezing rate would be expected to result in clear ice rather than the cloudy ice that forms in seasonal sea ice. Freezing in sea ice creates myriad brine channels as the salt is excluded from the ice in the freezing process. We assume here that for a thick ice cover the freezing rate is so slow \((<10 \text{ cm yr}^{-1})\) that the excluded salt is carried away without forming brine channels resulting in clearer ice similar to lake ice. However for rapid freezing such as is associated with thin ice covers this is not the case. We are not attempting to model these thin ice cases for which light availability is not in question. Permanent ice covers similar to those in the dry valleys cover sea water in the flords of the Bungar Hills Oasis in Antarctica [Andersen et al., 1995].

This clarity of the ice accounts for the result seen in Figure 1, that light levels above \(10^{-5}\) persist for thicknesses up to 30 m. Photosynthesis is reported for light levels as low as \(5 \times 10^{-4}\) of full sunlight [Berner and Evenari, 1978; Littler et al., 1985; Robinson et al., 1995]. From Figure 1, it is apparent that ice cover thickness of more than about 30 m are not interesting since no significant light penetrates these ice covers.

The thickness of ice on the snowball Earth has been computed using Eq. (2). Figure 2 illustrates the dependence of the ice thickness on the ablation rate for surface temperatures of \(-30^\circ C\), \(-40^\circ C\), and \(-50^\circ C\), for values of \( a = 0.7\), \( r = 0.1\), and \( h = 0.8\) m similar for the brightest ice from the dry valley lakes in Antarctica [McKay et al., 1994]. In addition, for \(-40^\circ C\), values of \( a = 0.8\) and 0.6; \( h = 0.7\) and 0.9 m are shown as dashed and dotted lines. The albedo, \( a\), and the surface dust loading, \( r\), appear in the equation in mathematically identical ways. They both reduce the amount of light entering into the ice cover. For this reason only variations in \( a\) have been explored. Our nominal values of albedo (0.7) and dust loading (0.1) correspond to a pure albedo value of 0.73.

Interestingly, for temperatures near \(-30^\circ C\), the ice thickness is less than a few meters for all values of the ablation — the thickness is set by sunlight alone. The
importance of sunlight penetrating into the ice cover can be seen by comparing the curves for -40°C; reducing the sunlight either by increasing $a$ or decreasing $h$ significantly increases the thickness of the ice cover.

The albedo is a key parameter in Eq. (2) and Figure 3 shows the thickness of the ice versus the albedo for temperatures of -50°C and -30°C for values of the parameters $a$, $r$, and $h$ as in Figure 2. Various ablation rates are shown. For the Antarctic dry valley lakes McKay et al. [1994] found maximum albedo for clear ice regions of 0.6 in the visible. Calderia and Kasting [1992] use a similar value for the albedo of an ice-covered Earth. Over bare sea ice, Allison et al. [1993] reported albedo values as low as 0.2. The albedo of sea ice that is free of snow is typically 0.6 or less in the PAR [Allison et al., 1993; Greenfell and Perovich, 1984; Brandt et al. 1999] with solar averaged albedo typically 0.5 or less. Clean, thick ice may have a higher albedo than the Antarctic lakes and we use 0.7 as our nominal value. The presence of snow can increase the albedo to high values, 0.8 or above [Brandt et al. 1999]. Snow accumulation is ignored since we are considering the regions in the tropics in which descending air creates dry conditions. Figure 3 also shows that for temperatures warmer than -30°C, and albedos less than 0.75, sunlight alone is sufficient to limit the ice thickness.

It is interesting to express the results as a function of mean temperature since this, rather than ablation rate, is what is usually predicted by climate models. The dependence of the ablation rate with temperature is dominated by the vapor pressure of ice [Clew et al., 1988; Moore et al., 1995] and is determined in these calculations by scaling from the measured value in the Antarctic dry valley (35 cm yr$^{-1}$ at -20°C). Thus at -40°C the ablation rate is 4 cm yr$^{-1}$ and is 0.4 cm yr$^{-1}$ at -60°C. The results for the nominal case are shown in Figure 4 as the solid curve labeled with the value of the albedo (0.7). Interestingly, the ice thickness falls to low values for temperatures warmer than about -40°C. However, an important difference between the dry valley lakes and the tropical regions of the snowball Earth is that the lakes are surrounded by dry land and water vapor that evaporates is readily carried off. On the snowball Earth some of the water vapor may remain and reduce evaporation rates.

In the results presented in Figures 2, 3, and 4, only ice covers thicker than 5 m are accurately calculated because the model averages over the diurnal and annual cycles and is therefore only applicable to thick ice covers. Ice cover thickness less than 5 m imply a partly open sea surface with variable ice thickness and allow sufficient light penetration to support photosynthesis.

Discussion

The mean equatorial temperature for a snowball Earth is uncertain. Jenkins and Smith [2000] suggest a global mean temperature of -52°C corresponding to an ice thickness of 60 m and an associated light level of $10^{-6}$, even for an albedo of 0.6. However, this is a global mean temperature and the equatorial temperature is predicted to be closer to -30°C — too warm to form a persistent optically thick ice cover even if the ice albedo was 0.8.

This analysis suggests that even if Earth was largely covered with ice during the Neoproterozoic there would likely remain some regions in the tropics that would be covered with ice thin enough ($<10$ m) to allow for photosynthesis. Since the total flux of sunlight available for primary productivity would be greatly reduced compared to ice-free conditions, this conclusion is not at odds with the geological data which implies that, globally, the proportion of organic carbon to total carbon burial changed from almost 0.5 before the glaciation to virtually zero during it [Kirschvink, 1992]. Locally, however, photosynthesis through thin ice could have allowed for the survival of photosynthetic marine algae and bacteria in a reduced, but complete, trophic chain. Life would then have been well poised to recover from this event.

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References


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