Did Antarctic sea-ice expansion cause glacial CO₂ decline?

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[1] Recently, Stephens and Keeling [2000] have put forward an appealing theory for explaining the decrease in glacial atmospheric CO₂. They argue that a compact sea-ice cover extending southward of ~55° S trapped large amounts of CO₂ beneath the sea surface, thus accounting for the lower atmospheric concentrations. An atmosphere-ocean box model in which sea-ice area is prescribed allows them to simulate ~80% of the CO₂ drawdown. However, glacial CO₂ levels can be attained in their model only when the fraction of ice-covered area southward of the Antarctic Polar Front rises to 99–100%. We present simulations with a coupled sea ice-upper ocean model indicating that ice-area fractions so large might have not prevailed even under rather extreme glacial conditions. The combination of our glacial ice-coverage estimates with the ice area-CO₂ relation derived by Stephens and Keeling suggests that CO₂ sequestration under sea ice could account for at most 15–50% of the total glacial CO₂ decline. INDEX TERMS: 4267 Oceanography: General: Paleonography, 3344 Meteorology and Atmospheric Dynamics: Paleoclimatology, 4207 Oceanography: General: Arctic and Antarctic oceanography, 4255 Oceanography: General: Numerical modeling

1. Introduction

[2] Elucidating the causes of the glacial/interglacial atmospheric CO₂ cycles and their correlation with ice-volume and temperature changes [Petit et al., 1999] is proving a most elusive problem. Depletion of surface nutrients at high latitudes is a favoured mechanism, but it is unclear to what degree such depletion occurred, and how much drawdown resulted from it [Archer et al., 2000; Sigman and Boyle, 2000]. This has prompted some investigators to turn their attention toward physical processes that might exert a firmer control on atmospheric CO₂. It has recently been suggested that a drastic reduction in CO₂ outgassing over the Southern Ocean attending on a glacial expansion of the sea-ice cover could account for a substantial part of the 80–100 ppm decrease in CO₂ during the Last Glacial Maximum (LGM) [Stephens and Keeling, 2000] (hereafter SK).

[3] The theory of SK hinges upon the hypotheses that the bulk of the global deep water upwelling takes place southward of the Antarctic Polar Front (APF), and that sea ice covered most of the upwelling region during the LGM. An atmosphere-ocean box model incorporating these two premises allowed them to reproduce a 65-ppm reduction in CO₂. However, their simulations reveal that sea ice-driven regulation of CO₂ outgassing is effective only if the winter-time sea-ice area fraction southward of the APF (hereafter A) is large: CO₂ drawdowns above 40 ppm necessitate values of A ≥ 95%, and their LGM case stipulates A = 99.5%. The reason why sea ice is an inefficient barrier for air-ocean CO₂ exchanges lies in the following compensating mechanism. A positive ice-area anomaly causes both a decrease in atmospheric CO₂ and an increase in oceanic CO₂, leading to higher surface partial pressure differences, and hence to more vigorous CO₂ fluxes over the remaining open water area, thereby partly offsetting the effects of the initial perturbation. Because of this negative feedback, only a virtually locked, Arctic-like ice cover is able to reduce CO₂ concentrations down to glacial levels.

[4] Arguments have been put forward that vertical mixing increased during the LGM as a result of stronger surface cooling, ensuing sea-ice formation/salt rejection, and higher wind speeds [Moore et al., 2000]. This would further weaken the sea ice-CO₂ link. Vertical mixing is a key factor in controlling CO₂ outgassing, since appreciable CO₂ release can occur only after the 50–300-m-deep surface layer has been brought into contact with the CO₂-rich underlaying waters. Upper ocean destratification during the LGM would have enhanced CO₂ exchanges over the more constricted open water areas.

[5] At present, strong circumpolar winds pull the ice apart, and intense oceanic heat fluxes hinder ice growth, to the point that sea ice covers no more than 80–90% of the total winter ice-pack area [Gloersen et al., 1992]. The question naturally arises whether the harsher conditions that reigned during the LGM, notably lower air temperatures and faster winds, could have led to a nearly full sea-ice cover underneath which most of the Southern Ocean CO₂ would have been entrapped.

[6] Here, we examine simulations of a number of conceivable LGM scenarios carried out with a coupled sea ice-upper ocean model which suggest that the question above might need to be answered in the negative.

2. Model, Forcing, and Experiments

[7] The coupled sea ice-upper ocean model is that of [Fichefet and Morales Maqueda, 1997] (hereafter FMM). The ice model is a thermodynamic-dynamic one, and takes into account the heat capacity of snow and ice, the thermodynamic effects of the sub-grid-scale snow and ice thickness distributions, the storage of latent heat in brine pockets, and the formation of snow ice. It also allows for the dynamic creation and maintenance of leads within the ice cover (leads over a freezing ocean owe their existence exclusively to the motion of sea ice). The ice-dynamics formulation includes ice-ice plastic interactions. The upper ocean model consists of unconnected, 300-m-deep water columns incorporating parameterisations of the mixed layer and the diffusive pycnocline. The coupled model has a horizontal resolution of ~2.8° × 2.8°, and the ocean vertical levels vary in depth from 5 m at the surface to 50 m at the bottom.

[8] The following changes have been made from the initial FMM model. In FMM, an ad hoc redistribution of the heat flux through open water was applied to ensure that thermodynamic closure of leads did not occur. This artefact has been removed. The model uses now a physically-based formulation of the opening of leads by shear deformation, a mechanism to be reckoned with for concentrations above 85–90% [Stern et al., 1995]. In addition, we diagnose the collection thickness of new ice in leads as

\[ d_i = d_0 + \left( 2 \gamma \alpha \left( 1 - \rho_i / \rho_s \right) \right) \left( u_i - u_i^0 \right) \]

where \( d_0 = 0.1 \) m is the thickness of frazil ice arriving at the edge of a lead, \( g = 9.8 \) m s⁻², \( \alpha = 0.3 \) is the volumetric fraction of frazil ice in water, \( \rho_i = 950 \) kg m⁻³ is ice density, and \( u_i \) is the...
Experiment PD. Model runs span 15 integration years. 

thermodynamics processes (Åd and Åt, respectively). To support or 0.75 with 0.75/C2. A of leads (B), which was lower than today by 0.75/C1.

Boundary conditions. Of these, the most certain is insolation, and rates of change of H20 and H20/ month, both of them averaged over the pack-ice region encircled by the APF: mixed layer depth (Hm), and rates of change of Åv and Ås, respectively. Table 1 displays the values attained by these variables in each simulation. We define wintertime as the 5-month period with highest ice-area fractions, and assume the APF to be delineated by the 3 °C annual mean sea-surface isotherm of [Levitus, 1982], which encloses an oceanic area of ~32 x 10⁶ km².

[12] Experiment PD reproduces well the modern sea-ice cover. As shown in Figure 1, the location of the simulated ice front (15% ice concentration) is close to the satellite-derived one [Gloersen et al., 1992]. However, ice concentrations are overestimated all around Antarctica, and the mean ice compactness within the winter ice pack exceeds the observed value by ~6% (84% vs. 78%). This discrepancy arises from the absence in our model of short term variability in the ice dynamics, such as that caused by inertial and tidal motions, whose contribution to wintertime lead formation is significant [Eisen and Kottmeier, 2000]. The simulated mixed layer is on average only ~7 m shallower than inferred from [Levitus, 1982] based on a Δσ1 = 0.125 kg m⁻³ criterion.

[13] The experiment labelled T10W50 is our best guess LGM scenario, as it is the one with surface-temperature and wind-speed perturbations closest to those derived from the paleorecords. Figure 1c shows that the simulated ice front extends amply past the APF, and indeed also beyond the location of the LGM ice front estimated by CLIMAP [Moore et al., 2000], suggesting that a glacial anomaly of ΔTm = −10 °C might, in fact, be too large, at least for latitudes above ~55° S. In spite of the vastness of this ice cover, the average ice concentration southward of the APF reaches only ~92%, which in the model of SK corresponds to a lowering of CO₂ by just 35 ppm. We also note that, as evidenced by the deepening of the mixed layer, vertical mixing is now stronger than in the PD case. An analysis of the complete set of LGM experiments will help understanding these results.

[14] The model’s response in the sequence of LGM simulations exhibits a tidy pattern. For a given wind-speed perturbation, both Åv and Hm increase nonlinearly with decreasing air temperatures. The two quantities experience relatively large changes when the atmosphere cools down from ΔTm = −5 °C to ΔTm = −10 °C. For further temperature drops, however, fairly smaller increases occur (for Hm, this is most apparent at high wind speeds, since mixing is then mainly controlled by mechanical stirring). This nonlinearity is due to the fact that, barring heat transport through ice, net oceanic cooling is proportional to the area of leads, which decreases with decreasing Tm. Thus, if the ice concentration is 90%, say, a doubling of the surface heat fluxes will cause a meagre ~20% increase in ice formation, salt rejection, and vertical mixing per unit oceanic area. It is worthwhile noting that, while lead-closure rates southward of the APF (Åt) for a lead opening on the LGM ice simulation, they were very close to be in balance in the LGM experiments. Of the two processes that add up to Åt, namely shearing deformation and ice

Table 1. Simulated wintertime mixed layer and sea-ice cover characteristics

<table>
<thead>
<tr>
<th>Exp.</th>
<th>ΔTa (°C)</th>
<th>Δv9 (%)</th>
<th>Hm (m)</th>
<th>Åv (%)</th>
<th>Ås (%)</th>
<th>ΔHm (m)</th>
<th>ΔHs (m)</th>
<th>ΔF (W/m²)</th>
<th>ΔV (m/s)</th>
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<tr>
<td>PD</td>
<td>0</td>
<td>0</td>
<td>107</td>
<td>46.8 (282)</td>
<td>17</td>
<td>17</td>
<td>17</td>
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<td>0</td>
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<tr>
<td>T05W00</td>
<td>−5</td>
<td>0</td>
<td>83</td>
<td>77.1 (270)</td>
<td>17</td>
<td>17</td>
<td>17</td>
<td>0</td>
<td>0</td>
</tr>
<tr>
<td>T10W00</td>
<td>−10</td>
<td>0</td>
<td>117</td>
<td>93.7 (243)</td>
<td>17</td>
<td>17</td>
<td>17</td>
<td>0</td>
<td>0</td>
</tr>
<tr>
<td>T15W00</td>
<td>−15</td>
<td>0</td>
<td>145</td>
<td>95.4 (238)</td>
<td>17</td>
<td>17</td>
<td>17</td>
<td>0</td>
<td>0</td>
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<tr>
<td>T20W00</td>
<td>−20</td>
<td>0</td>
<td>154</td>
<td>96.6 (236)</td>
<td>17</td>
<td>17</td>
<td>17</td>
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<td>0</td>
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<tr>
<td>T05W50</td>
<td>−5</td>
<td>+50</td>
<td>148</td>
<td>75.9 (271)</td>
<td>17</td>
<td>17</td>
<td>17</td>
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</tr>
<tr>
<td>T10W50</td>
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<td>+50</td>
<td>174</td>
<td>92.3 (247)</td>
<td>17</td>
<td>17</td>
<td>17</td>
<td>0</td>
<td>0</td>
</tr>
<tr>
<td>T15W50</td>
<td>−15</td>
<td>+50</td>
<td>178</td>
<td>94.3 (242)</td>
<td>17</td>
<td>17</td>
<td>17</td>
<td>0</td>
<td>0</td>
</tr>
<tr>
<td>T20W50</td>
<td>−20</td>
<td>+50</td>
<td>180</td>
<td>95.6 (238)</td>
<td>17</td>
<td>17</td>
<td>17</td>
<td>0</td>
<td>0</td>
</tr>
<tr>
<td>T05W75</td>
<td>−5</td>
<td>+75</td>
<td>175</td>
<td>75.6 (267)</td>
<td>17</td>
<td>17</td>
<td>17</td>
<td>0</td>
<td>0</td>
</tr>
<tr>
<td>T10W75</td>
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<td>192</td>
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<td>17</td>
<td>17</td>
<td>17</td>
<td>0</td>
<td>0</td>
</tr>
<tr>
<td>T15W75</td>
<td>−15</td>
<td>+75</td>
<td>192</td>
<td>93.5 (244)</td>
<td>17</td>
<td>17</td>
<td>17</td>
<td>0</td>
<td>0</td>
</tr>
<tr>
<td>T20W75</td>
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<td>+75</td>
<td>195</td>
<td>94.9 (240)</td>
<td>17</td>
<td>17</td>
<td>17</td>
<td>0</td>
<td>0</td>
</tr>
</tbody>
</table>

The figure in brackets is the atmospheric CO₂ concentration (in ppm) according to [Stephens and Keeling, 2000].

The first (second) term on the left-hand side is the contribution to Åt from shearing deformation (ice divergence).
dissipation, changes in the latter are those that chiefly assist in establishing a closure/opening balance as air temperatures decrease. As is shown in Table 1, shearing deformation generates opening rates that, for a given wind speed, increase only weakly with ice concentration. In contrast, net ice-area divergence southward of the APF (i.e., total ice-area transport across the APF) rises significantly as air temperatures drop, thus neutralising the larger lead-closure rates. The increase in ice export across the APF results from the Coriolis intensification of the meridional component of the ice velocity as the ice, which by and large moves in free drift at those latitudes, grows thicker (cf. Table 1, last two columns).

To ascertain the role of wind perturbations in controlling ice-area fractions, we first note that shearing deformation, ice divergence, and lead energy budget (column $B_l$ in Table 1) are all three approximately proportional to wind speed. One would therefore expect changes in the wind to have only a lesser impact on $A$, were it not for the fact that the collection thickness of new ice tends to grow with $v_g$ (roughly quadratically). As Table 1 reveals, the increase in ice collection thickness with enhanced winds creates a further mechanical hindrance to lead closure, which, for a given $\Delta T_a$, results in a decrease in $A$ as $\Delta v_g$ augments. Note that lower winds during the LGM, as suggested by [Sigman and Boyle, 2000], would have led in contrast to an increase in fractional ice coverage.

4. Conclusions

This investigation has been motivated by the conjecture of [Stephens and Keeling, 2000] that the glacial atmospheric CO$_2$ decay was caused by a reduction in outgassing rates over the Southern Ocean owing to a wintertime expansion and compaction (up to 99.5%) of the sea-ice cover. To assess whether the existence of such a compact ice cover is feasible in a LGM environment, we have carried out several integrations of the sea ice-upper ocean annual cycle in the Southern Hemisphere under boundary conditions representative of today’s and LGM’s climates. Despite the model’s propensity to overestimate ice concentrations by several percents, the maximum value of the simulated ice-area fraction southward of the Antarctic Polar Front is $\sim95-97\%$, and our best guess LGM scenario yields a figure of $\sim92\%$. The persistence of relatively large amounts of open water within the ice pack stems from the alliance of three dynamical effects that tend to counter lead closure: shearing deformation and ice divergence, which promote lead opening, and wind-induced variations in the collection thickness of new ice, which decelerate lead closure. Our simulations show as well that the Southern Ocean mixed layer might have been up to $\sim90$ m deeper during the LGM. Presumably, the stronger oceanic vertical mixing would have supplied additional deep-ocean CO$_2$ to the surface, thus partly compensating for the increase in ice-area fractions. Inserting our simulated ice fractions into the ice-area-versus-CO$_2$ concentration curve of [Stephens and Keeling, 2000], we estimate that entrainment of CO$_2$ by sea ice could account for not more than 15–50% of the glacial CO$_2$ decrease. Although it is implausible that Antarctic sea-ice expansion alone can fully explain the perplexing atmospheric CO$_2$ depletion during the LGM, our results do not discard sea ice as a potentially important control, one that might need to be considered, in conjunction with other mechanisms, by theories striving to solve the puzzle.

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References


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