The Cryosphere Discuss., 5, 2349–2363, 2011
www.the-cryosphere-discuss.net/5/2349/2011/
doi:10.5194/tcd-5-2349-2011

The Cryosphere Discussions

How reversible is sea ice loss?

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Received: 29 July 2011 – Accepted: 22 August 2011 – Published: 8 September 2011

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Published by Copernicus Publications on behalf of the European Geosciences Union.

Abstract

It is well accepted that increasing atmospheric CO_2 results in global warming, leading to a decline in polar sea ice area. Here, the specific question of whether there is a tipping point in the sea ice cover is investigated. The global climate model HadCM3, is used to map the trajectory of sea ice area under idealised scenarios. The atmospheric CO_2 is first ramped up to four times pre-industrial levels ($4 \times CO_2$) then ramped down back to pre-industrial levels. We also examine the impact of stabilising climate at $4 \times CO_2$ prior to ramping CO_2 down to pre-industrial levels. Against global mean temperature Arctic sea ice area has little hysteresis while the Antarctic sea ice shows significant hysteresis – its rate of change slower, with falling temperatures, than its rate of change with rising temperatures. However, we show that the driver of the hysteresis is the hemispherical differences in temperature change between transient and stabilisation periods. We find no irreversible behaviour in the sea ice cover.

1 Introduction

The decline of Arctic sea ice in response to increasing concentrations of atmospheric CO_2 has been much studied in climate models (Gregory et al., 2002; Winton, 2006; Holland et al., 2006; Ridley et al., 2007; Wang and Overland, 2009). Typically these show an approximately linear relationship between the decrease in extent of the Arctic sea ice in September and global average near surface warming, although the constant of proportionality remains uncertain. Attempts to constrain the models using observations, provides one means of narrowing this uncertainty (e.g. Boe et al., 2009). Most climate models do not show a clear abrupt acceleration of the rate of summer sea ice loss with global temperature rise. However, in HadCM3, whilst the demise of the Arctic annual mean sea ice varies linearly with temperature over a wide range of values, the winter ice does show a more rapid decline as an apparent temperature threshold is reached. This only occurs after a large fraction of the winter ice has already been lost above a global warming of around 7 °C (Ridley et al., 2008).

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For the Antarctic the sea ice extent has seen a slight increase since the 1970s (Turner et al., 2009), and although climate models can simulate this, the model uncertainty is greater than in the Arctic (Arzel et al., 2006). The pattern of sea ice change is related to the southern annular mode (Lefebvre et al., 2004), and the history of changes in the Southern Ocean (Goosse and Renssen, 2005)

These studies of Arctic and Antarctic sea ice loss raise the question, if the global average warming is reversed will the sea ice quickly return? The answer has implications for how sea ice is viewed in discussions of dangerous climate change and policies to avoid them.

Schroder and Connolley (2007) performed experiments for both the southern and Northern Hemisphere sea ice using the HadCM3 climate model in which the ocean mixed layer was adjusted to represent that in ice free regions. They found that the sea ice recovered fully within 15 years. More recently, Tietsche et al. (2011) showed that the instantaneous removal of Arctic sea ice in a climate model, from various periods during its decline, results in its rapid regrowth to the state prior to its removal. Taken together the experiments suggest that there no threshold, either in sea ice extent or global average warming, that leads to sea ice loss becoming irreversible in climate models.

In this study we examine the issue of irreversibility using a different experimental approach. If the sea ice loss is found not to be irreversible then we will extend our analysis to estimate the magnitude of any lag in the recovery of sea ice behind changes in the climate forcing. Our approach involves ramping up the atmospheric CO_2 concentration from pre-industrial levels (280 ppm) then bringing it back down again either straight away or after a period of stabilisation. Although the rates of CO_2 change are idealised in our study, the rates of CO_2 increase are not inconceivable when compared to the range of available policy relevant emission scenarios (Nakienovi et al., 2000). The rates of CO_2 decline are more speculative but an increasing number of studies (e.g. Arora et al., 2011) envisage negative carbon emissions late in the 21st century from burning of biofuels with carbon capture and storage (Matthews, 2008).

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2 Method

The coupled Ocean-Atmosphere general circulation model HadCM3 is used in this study. The sea ice model in HadCM3 consists of a thermodynamic component with simple dynamics in which winds drive the ocean currents then advect the sea ice (Gordon et al., 2000). The model produces a reasonable simulation of Arctic sea ice change in response to climate forcing, although there is less skill for the Antarctic (Gregory et al., 2002).

Two simulations were conducted to map the possible trajectories of the ice area in response to CO_2 induced global temperature change (Fig. 1). In both simulations the atmospheric CO_2 concentration is initially increased at a rate of 2 % a year from preindustrial concentrations for 70 years until a CO_2 concentration four times preindustrial is reached (point "a" on Fig. 1). The first experiment then immediately ramps down CO_2, at 2 % a year, until pre-industrial concentrations are reached (point "b"), followed a period of stabilisation with CO_2 pre-industrial concentrations for 160 years. We will call this the "fast" scenario. The second, "slow" scenario, stabilises concentrations at $4 \times CO_2$ for 1000 years (point "c"), and then ramps down CO_2 at 2 % a year for 70 years (point "d") followed by 160 years with CO_2 at pre-industrial levels.

3 Results and discussion

3.1 Arctic ice extent

As expected from Ridley et al. (2007) changes in the annual mean Arctic ice area are approximately linear with global temperature change (Fig. 2a). During the atmospheric CO_2 ramp up the Arctic becomes ice free in September following a global temperature rise of approximately 4 °C, and is ice free year round following a global temperature rise of 7.5 °C – a global temperature only achieved during the long stabilisation period at $4 \times CO_2$. There is a slight change in the gradient once the summer ice is lost, and winter ice loss accelerates towards the end of its demise. With the immediate ramp down from point "a" the ice area is found to be entirely reversible with global mean temperature, and there is no evidence of the sea ice recovery lagging behind the global temperature decline. After a sustained period of stabilisation (point "c"), the CO_2 ramp-down also sees the winter ice recover as global temperatures fall. Indeed there is some evidence of the ice area recovering slightly more rapidly than in the "fast" scenario.

The sea ice area change is plotted against the regional temperature change (Fig. 2b), defined as the average 1.5m air temperature north of 70° N. The scatter is reduced and the simulation of ice area is entirely reversible with this measure of local temperature change. The "slow" scenario lag in Fig. 2a is absent in Fig. 2b, indicating that the effect is due to the regional ocean cooling faster than the global ocean, rather than being an innate feature of the sea ice properties. It also suggests that any changes that occur in the ocean vertical structure, as a consequence of heat uptake, do not influence the surface heat fluxes and the associated sea ice response.

3.2 Antarctic sea ice extent

In the Antarctic (Fig. 3) the return trajectory, points a–b, the ice cover is reversible if the atmospheric CO_2 concentration and temperatures are reduced quickly after the CO_2 peak. (Fig. 3a). However, on stabilisation at $4 \times CO_2$ the temperature sensitivity of the ice area, although still linear, is noticeably increased. The ice sensitivity to temperature changes is different again during the CO_2 ramp-down, points c–d, being lower than during either the ramp-up or stabilisation period.

When the ice area change is depicted as a function of regional temperature (Fig. 3b), defined as the average 1.5 m air temperature between 75° S and 55° S, the ice response is more similar to the Arctic case (Fig. 2b), being almost reversible in the slow scenario. Only a slight deviation in ice area against temperature is seen between the ramp-up and ramp-down pathways. Such a deviation, not investigated here, may be due to differences in the modelled southern annular mode and its impact on the sea ice spatial coverage (Lefebvre et al., 2004).

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3.3 Asymmetry in the Hemispheric temperature response

The radiative forcing in HadCM3 first results in a temperature anomaly in the ocean mixed layer system which gradually spreads to the deep ocean, a process noted by Hansen et al. (1984). High latitudes are important for deep ocean heat uptake, as the characteristic deep vertical mixing of these regions provides access to store heat in the deep ocean (Russell et al., 2006). The deeper mixed layers at high southern latitudes are associated with enhanced deep ocean heat uptake in the transient response to CO_2 forcing, and consequently lead to regionally lesser surface warming (Boe et al., 2009; Winton et al., 2010)

In the HadCM3 transient climate simulations, atmospheric CO_2 ramp-up and ramp-down, the Northern Hemisphere 1.5m air temperature changes 1.6 times faster than that of the Southern Hemisphere. This is a consequence of a combination of enhanced Southern Ocean heat uptake and land-sea temperature contrast (Joshi et al., 2008; Dong et al., 2009), in that land warms/cools faster than ocean, and the bias in the global distribution of land to the Northern Hemisphere. During the period of stabilisation at $4 \times CO_2$, as the climate system approaches equilibrium, the balance is reversed and the Southern Hemisphere warms at a rate of 1.2 times that of the Northern Hemisphere. The transition between the faster warming northern and Southern Hemispheres occurs over a period of ~30 years (Fig. 4).

Throughout the $4 \times CO_2$ stabilisation period (points “a” to “c”) the intermediate waters of the southern ocean take up heat in summer. However, the modelled mixed layer depth deepens in winter such that a portion of the heat from below is entrained into the mixed layer resulting in warmer winter air temperatures. Figure 5 shows the evolution of the Southern Ocean temperature structure. Over the fast scenario (Fig. 5a) an imbalance of surface fluxes leads to ocean heat uptake, resulting in a warming at ~400 m depth. The heat diffuses both into the deep ocean and upwards where it is entrained by the mixed layer. The entrainment is evident after the peak in the ramp-up of CO_2 at (a), as the sea ice minimum extent lags peak CO_2 by ~12 years while the heat

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anomaly is eroded. The full recovery of the sea ice cover, on return to $1 \times CO_2$ (point "b") is similarly delayed. During the slow scenario (Fig. 5b) and stabilisation at $4 \times CO_2$ the ocean heat uptake is considerable and the temperature anomaly at 400 m depth grows to 4 °C. As a consequence of the release of the heat the ocean surface remains ~2 °C warmer even after a 160 year period of stabilisation at $1 \times CO_2$. Warming of the deep ocean continues to the end of the simulation. Consequently, the Antarctic sea ice recovery lags behind that in the Arctic.

4 Conclusions

We have conducted idealised simulations with the climate model HadCM3, in which the atmospheric concentration of CO_2 is first ramped up from pre-industrial to four times pre-industrial levels. In one experiment the CO_2 is immediately ramped back down to, and then stabilised at, pre-industrial levels. In a second experiment the CO_2 was stabilised at four times pre-industrial levels for 1000 years before ramping down to pre-industrial levels.

We have investigated the polar climate response to climatic changes arising from the two CO_2 scenarios. It is clear that the sea ice in the fast scenario responds linearly to global mean temperature almost equally for the ramp-up and ramp-down of CO_2. However, in the slow scenario, the deep waters of the Southern Ocean become warmer and more saline over centennial timescales. The dominant source of the oceanic heat is increased surface heat uptake as sea ice declines. The heat is stored temporarily in intermediate waters at ~400 m depth before eventually being transferred both back to the surface and to the deep ocean. The intermediate waters become a store of heat that can be accessed through mixed layer entrainment as the climate cools, slowing the rate of recovery of the sea ice in some regions of Antarctica. In the Arctic, a surface freshening reduces the mixed layer depth and prevents a surface release of the stored heat before it is lost to the deeper ocean.

An analysis of the climate sensitivity of the sea ice cover shows that against global mean temperature rise there is apparently a lag in the recovery of lost sea ice, which is especially noticeable in the Antarctic. When the same analysis is conducted against local temperatures no lag is evident. The apparent sea ice lag arises because the Southern Hemisphere temperatures lag behind those in the north during the ramp-up and immediate ramp-down experiments. However, during the stabilisation at $4 \times CO_2$ the southern and northern temperatures are close to equalising. The ramp-down in the slow scenario, then does not start from the same state as the ramp-down in the fast scenario, and a hemispheric temperature hysteresis is created. This feeds through to an effect on hemispheric climate processes such as those related to sea ice.

Thus, we find no clear evidence of a threshold of irreversibility in the sea ice, but Antarctic sea ice does lag behind global temperature during a return to lower temperatures after a period of extended CO_2 stabilisation. We recommend extending this study to other models with a more sophisticated treatment of sea ice physics and different ocean mixing characteristics.

Acknowledgements. This work was supported by the Joint DECC/Defra Met Office Hadley Centre Climate Programme (GA01101).

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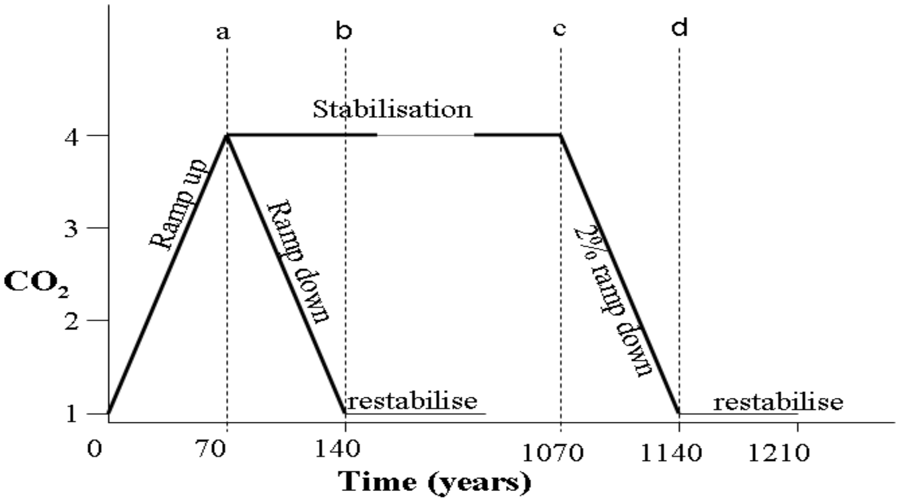

Fig. 1. The prescribed CO_2 trajectories used in the HadCM3 experiments. Trajectory 0-a-b represents the "fast" scenario. Trajectory 0-a-c-d represents the "slow" scenario.

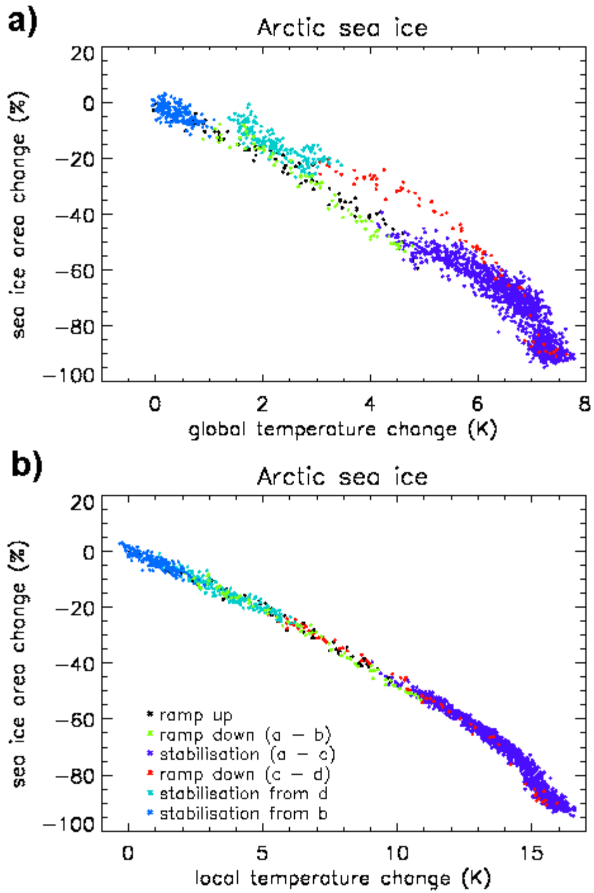

Fig. 2. The sensitivity of Arctic sea ice area change to **(a)** global temperature change and **(b)** local temperature change. Colours refer to the CO_2 trajectory markers specified in Fig. 1.

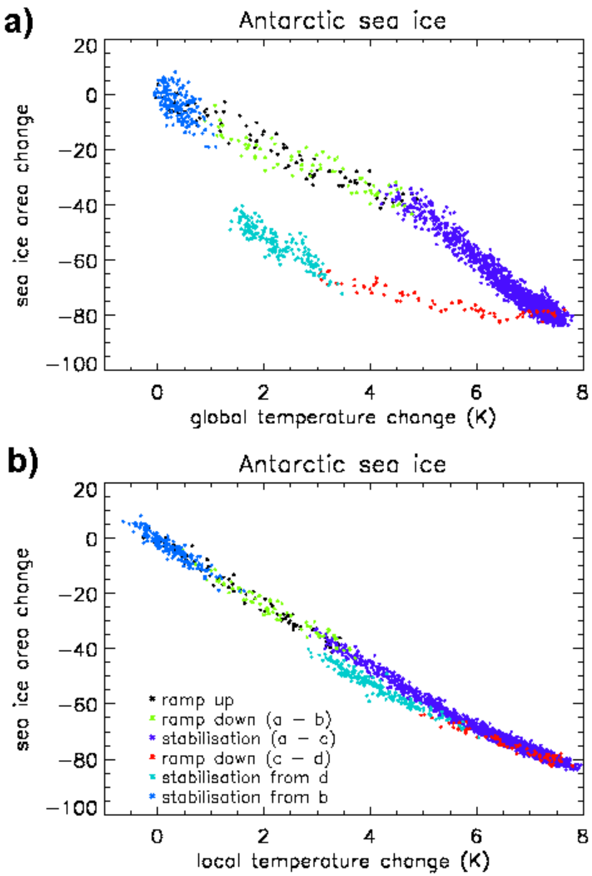

Fig. 3. The sensitivity of Antarctic sea ice area change to **(a)** global temperature change and **(b)** local temperature change. Colours refer to the CO_2 trajectory markers specified in Fig. 1.

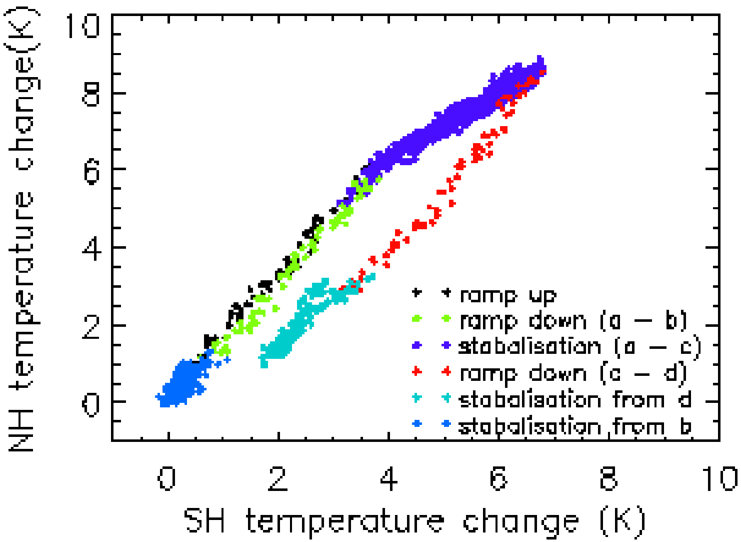

Fig. 4. The hysteresis in the hemispheric (NH = Northern Hemisphere, SH = Southern Hemisphere) near-surface air temperature. Stages "a" to "d" are as depicted in Fig. 1.

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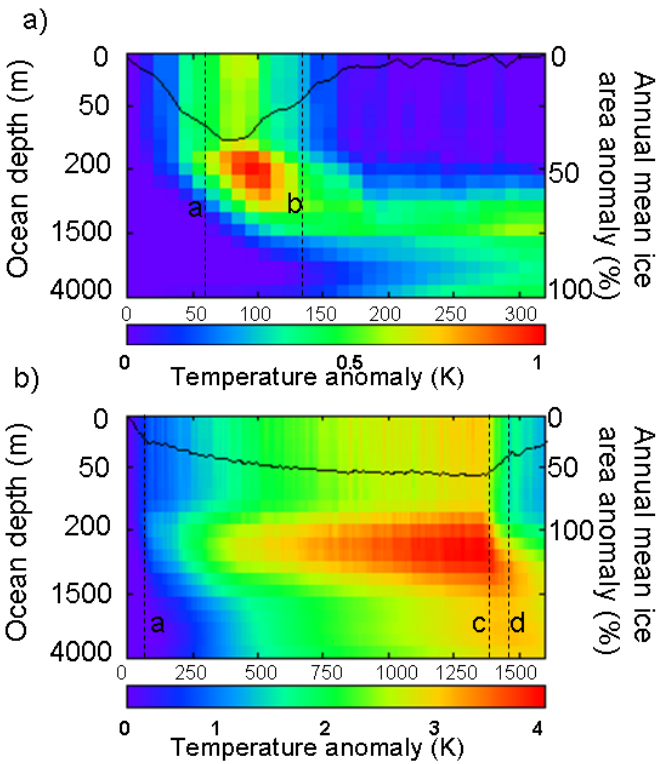

Fig. 5. The Southern Ocean potential temperature anomaly, as a function of ocean depth (averaged between 75° S and 55° S), in simulated time (years) for the fast scenario (top) and the slow scenario (bottom). At the top is the change in the associated anomaly in Antarctic sea ice area. Note that the ocean depth axis has is a non-linear scale. Labels are as depicted in Fig. 1.

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