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Response of ocean acidification to a gradual increase and decrease of atmospheric CO$_2$

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Abstract

We perform coupled climate–carbon cycle model simulations to examine changes in ocean acidity in response to idealized change of atmospheric CO$_2$. Atmospheric CO$_2$ increases at a rate of 1% per year to four times its pre-industrial level of 280 ppm and then decreases at the same rate to the pre-industrial level. Our simulations show that changes in surface ocean chemistry largely follow changes in atmospheric CO$_2$. However, changes in deep ocean chemistry in general lag behind the change in atmospheric CO$_2$ because of the long time scale associated with the penetration of excess CO$_2$ into the deep ocean. In our simulations with the effect of climate change, when atmospheric CO$_2$ reaches four times its pre-industrial level, global mean aragonite saturation horizon (ASH) shoals from the pre-industrial value of 1288 to 143 m. When atmospheric CO$_2$ returns from the peak value of 1120 ppm to pre-industrial level, ASH is 630 m, which is approximately the value of ASH when atmospheric CO$_2$ first increases to 719 ppm. At pre-industrial CO$_2$ 9% deep-sea cold-water corals are surrounded by seawater that is undersaturated with aragonite. When atmospheric CO$_2$ reaches 1120 ppm, 73% cold-water coral locations are surrounded by seawater with aragonite undersaturation, and when atmospheric CO$_2$ returns to the pre-industrial level, 18% cold-water coral locations are surrounded by seawater with aragonite undersaturation. Our analysis indicates the difficulty for some marine ecosystems to recover to their natural chemical habitats even if atmospheric CO$_2$ content can be lowered in the future.

Keywords: ocean acidification, carbon dioxide removal, carbon cycle modeling

Online supplementary data available from stacks.iop.org/ERL/9/024012/mmedia

1. Introduction

Since pre-industrial time human activities including fossil fuel combustion, cement production and land-use-change has released about 500 PgC (1 PgC = 10$^{15}$ g carbon) of CO$_2$ to the atmosphere (Canadell et al 2007). A recent study (Le Quéré et al 2013) estimates that between year 1959 and 2011 total anthropogenic CO$_2$ emission is 436 PgC, of which 27% is absorbed by the ocean and 29% is absorbed by the terrestrial biosphere with the remaining 44% staying in the atmosphere. Currently, the global ocean is a net sink of anthropogenic CO$_2$ with an absorption rate of about 2.5 PgC yr$^{-1}$ (Le Quéré et al 2013). From an atmosphere perspective, the ocean’s absorption of anthropogenic CO$_2$ moderates the amount of global warming by reducing atmospheric CO$_2$ concentrations.
From an ocean perspective, however, the ocean’s absorption of anthropogenic CO$_2$ perturbs ocean chemistry by increasing ocean’s acidity (Caldeira and Wickett 2003).

From pre-industrial time to the 1990s, global mean surface ocean pH has decreased from approximately 8.2 to 8.1, corresponding to a 26% increase in ocean’s acidity as measured by the concentration of hydrogen ion (Gattuso and Hansson 2011). The reduction in pH of the ocean caused by the uptake of CO$_2$ from the atmosphere, referred to as ocean acidification (Gattuso and Hansson 2011), has increasingly become a matter of concern. The concern of ocean acidification stems mainly from its potentially adverse effect on many marine ecosystems (e.g., Fabry et al. 2008; Doney et al. 2009), including the calcification rate of reef-building corals (e.g., Langdon and Atkinson 2005, Silverman et al. 2009, Crook et al. 2013) and other marine calcifying organisms (e.g., Orr et al. 2005, Hofmann et al. 2010), fish behavior (Devine et al. 2012, Ferrari et al. 2012), and marine biodiversity (Dupont et al. 2010, Hendriks et al. 2010).

Considering the effect of increasing atmospheric CO$_2$ on global climate and ocean chemistry, carbon dioxide removal (CDR), which aims to intentionally remove excess CO$_2$ from the atmosphere, has been proposed as one broad category of climate geoengineering (Royal Society 2005). The evaluation of climate and environmental effect of CDR geoengineering requires an understanding of the Earth system response to a forced decrease of atmospheric CO$_2$. Wu et al. (2010) and Cao et al. (2011) examined the response of temperature and precipitation change to a gradual decrease of atmospheric CO$_2$ from four times its pre-industrial value. Boucher et al. (2012) examined a wider range of change in climate variables such as precipitation, cloud, sea-ice, and carbon stocks in response to a gradual increase and decrease of atmospheric CO$_2$. These studies, in general, found that there is a time lag in the change of many climate variables relative to the change in atmospheric CO$_2$. As an extension of previous studies, here we perform model simulations to investigate the change of the ocean carbon cycle and carbonate chemistry at both the surface and sub-surface ocean in response to a gradual increase and then decrease of atmospheric CO$_2$. This study will provide new insight into the response of the ocean carbon cycle and ocean chemistry to changing atmospheric CO$_2$, and help us to better understand the potential effect of CDR geoengineering on ocean acidification.

2. Methods

We use an Earth system model of intermediate complexity to conduct idealized simulations in which atmospheric CO$_2$ concentration gradually increases with time and then decreases at the same rate. We perform idealized simulations as an illustration to investigate the likely response and reversibility of ocean acidification to changes in atmospheric CO$_2$ concentrations.

2.1. Model description

The model used in this study is University of Victoria (UVic) Earth System Climate Model (ESCM). The model has a horizontal resolution of 1.8° latitude and 3.6° longitude with 19 layers in the ocean. The physical component of the model includes a 3D ocean general circulation model that is coupled to a vertically integrated energy-moisture balance atmosphere model (Weaver et al. 2001). The ocean carbon cycle is simulated by a representation of inorganic carbon cycle component following the protocol of the Ocean Carbon-Cycle Model Intercomparison Project (Orr et al. 1999) and a representation of marine ecosystem component that simulates the interaction between nutrient, phytoplankton, zooplankton, and detritus (Schmittner et al. 2008). The representation of terrestrial carbon cycle and vegetation is based on the Hadley Centre Met Office TRIFFID dynamic vegetation model (Meissner et al. 2003). The UVic model reproduces well many aspects of the observed characteristics of the global climate (Weaver et al. 2001), the ocean carbon cycle (Schmittner et al. 2008), and historical uptake of carbon and its isotopes (Cao et al. 2009). The model has been widely used in many studies regarding global climate and the carbon cycle, such as future change of ocean biogeochemical cycles (e.g., Schmittner et al. 2008), interaction between climate change and the carbon cycle (e.g., Friedlingstein et al. 2006), and projection of ocean acidification (e.g., Hoegh-Guldberg et al. 2007; Cao and Caldeira 2008; Matthews et al. 2009; Turley et al. 2010).

2.2. Simulation experiments

The UVic model is first spun up for 10,000 years with a constant pre-industrial atmospheric CO$_2$ concentration of 280 ppm to reach a quasi-equilibrium state of climate and the carbon cycle. Starting from the end of the spin-up simulation a set of 1000-year transient simulations with time-varying atmospheric CO$_2$ is conducted: in the first simulation the effect of climate change and its feedback on the carbon cycle is included, and in the second simulation the effect of climate change is excluded by assuming that the change of atmospheric CO$_2$ has no direct radiative effect on global climate. Vegetation response to atmospheric CO$_2$ is not eliminated, which results in slight surface temperature change. For both simulations between year 1 and 140 atmospheric CO$_2$ is assumed to increase at a rate of 1% per year, reaching four times its pre-industrial concentration (4 × CO$_2$) at year 140. Then, atmospheric CO$_2$ decreases at the same rate of 1% per year, returning to its pre-industrial level at year 280. Thereafter, atmospheric CO$_2$ is kept constant at its pre-industrial level until year 1000.

2.3. Analysis of carbonate ocean chemistry

Ocean chemistry fields, including pH, carbonate ion concentration, and saturation state of calcium carbonate, are calculated based on the chemistry routine from the OCMIP-3 project (http://ocmip5.ipsl.jussieu.fr/OCMIP/phase3/) using the UVic model-simulated temperature, salinity, concentrations of dissolved inorganic carbon (DIC), alkalinity (ALK), and phosphate. Silicate concentration is taken from the World Ocean Atlas climatology (Garcia et al. 2010).

The saturation state of calcium carbonate (Ω), either aragonite or calcite, is calculated as (Feely et al. 2004):
Figure 1. Model-simulated time series of annual and global mean variables for the simulation with climate change (solid lines) and without it (dashed lines) as a function of year for (a) change in sea surface temperature (SST), (b) intensity of North Atlantic deep water formation (NADW), (c) depth of convection, (d) oceanic CO$_2$ uptake, (e) cumulative oceanic CO$_2$ uptake, (f) concentration of surface dissolved inorganic carbon (DIC), (g) surface pH, (h) surface [CO$_3^{2-}$], (i) surface $\Omega_{\text{aragonite}}$, (j) ocean mean pH, (k) ocean mean [CO$_3^{2-}$], (l) ocean mean $\Omega_{\text{aragonite}}$, (m) saturation horizon of aragonite, (n) the fraction of ocean volume with $\Omega_{\text{aragonite}} > 1$, (o) number of cold-water coral locations surrounded by seawater with aragonite undersaturation ($\Omega_{\text{aragonite}} < 1$). Results presented here are up to year 280 with results between year 1 and 140 shown in black and results between year 141 and 280 shown in purple. A version with 1000-year simulation results are given in figure S3.

where $K^*$ is the stoichiometric solubility product for aragonite or calcite.

Based on the calculated saturation state of carbonate minerals at model’s each ocean depth, we determine saturation horizon (the depth above which $\Omega > 1$ and below which $\Omega < 1$) by linear interpolation. There is evidence suggesting that global distribution of deep-sea cold-water corals is partly limited by the depth of the aragonite saturation horizon, and deep ocean acidification has the potential to alter the distribution of cold-water coral ecosystems (Guinotte et al 2006). To link simulated ocean chemistry to cold-water coral locations, we estimate aragonite saturation state for cold-water coral locations by assuming that each cold-water coral location possesses the ocean chemistry of the model grid cell that surrounds the coral. The information of longitude, latitude, and depth of cold-water coral location is taken from Freiwald et al (2005). Here, we provide a first-order estimation of the chemistry environment of cold-water corals with the acknowledgment that sea water chemistry within the coral reef system could be substantially different from that of the surrounding seawater (e.g., Andersson et al 2014).

3. Results

For the pre-industrial time spin-up the model captures the observed large scale pattern of key variables in the ocean carbon system such as DIC and ALK (figures S1, S2 available at stacks.iop.org/ERL/9/024012/mmedia). For both tracers the surface-to-deep ocean gradient is reasonably captured by the model, and the pattern difference in different ocean basins is reproduced well by the model. We also compare aragonite saturation horizon (ASH) calculated from model simulations with that calculated from GLODAP observations (Key et al 2004). Averaged globally (excluding grid points with no data in GLODAP), pre-industrial ASH calculated from model simulations is 1091 m, compared with 971 m that is calculated from GLODAP data. The model-simulated historical uptake of anthropogenic CO$_2$ during the mid-1990s is 2 PgC yr$^{-1}$, which compares well with the observational-based estimates (Cao et al 2009).

Next we discuss model-simulated change in the ocean carbon cycle and ocean chemistry. To highlight changes on the time scale of centuries, in the following we present results focusing on the period between year 1 and 280 when atmospheric CO$_2$ increases to 4×CO$_2$ and then decreases to pre-industrial level. The simulated response on longer time scales up to year 1000 is presented in the supplemental material (available at stacks.iop.org/ERL/9/024012/mmedia). All results presented in the following are from the simulation with climate change unless otherwise stated. We compare the results between the simulation with climate change and that without it near the end of this section.

Changes in global mean climate and carbon cycle fields as function of year are presented in figures 1 and S3, and their changes as a function of atmospheric CO$_2$ are given in figures 2 and S4. In our simulations at the time of 4×CO$_2$, sea surface temperature increases by 3.1 K (figure 1(a)); the intensity of North Atlantic deep water formation (NADW),
which is represented by the maximum meridional overturning streamfunction in the North Atlantic, decreases from 21.5 to 15.3 Sv (1 Sv = 10^6 m^3 s^-1) (figure 1(b)). Meanwhile, convective depth of the ocean, averaged globally and yearly, decreases from 16.3 to 13.0 m (figure 1(c)). When atmospheric CO₂ starts to ramp down and stabilizes at pre-industrial level, both NADW and convective depth start to rebound and experience a period of overshoot before finally returning to the pre-industrial level (figures 1(b), (c), S3(b), (c)).

With increasing atmospheric CO₂ concentrations the ocean continuously absorbs CO₂ from the atmosphere (figures 1(d), (e)). For the simulation with climate change at the time of 4×CO₂ global ocean has absorbed 632 PgC. As atmospheric CO₂ starts to decrease, the ocean’s uptake of atmospheric CO₂ begins to decrease as well, and the ocean gradually turns from a sink to a source of atmospheric CO₂ (figures 1(d), (e)). When atmospheric CO₂ returns to the pre-industrial level the ocean’s cumulative CO₂ uptake is 409 PgC. Nevertheless, due to the slow surface-to-deep-ocean transport, after the decrease in atmospheric CO₂ excess carbon continues to accumulate in the deep ocean (figure 3). For the same level of atmospheric CO₂, oceanic CO₂ uptake is distinctly different between the period with atmospheric CO₂ increase and the period with atmospheric CO₂ decrease (figures 2(d), (e)). For comparison, surface DIC concentration is largely independent of the pathway of CO₂ change: at the same atmospheric CO₂ level the difference in DIC concentration is slight between the period with atmospheric CO₂ increase and the period with atmospheric CO₂ decrease (figure 2(f)).

With continuous uptake of atmospheric CO₂ the ocean becomes more acidic as inspected from simulated changes in carbonate chemistry fields (figure 1). In the simulation with climate change at the time of 4×CO₂ global mean surface pH decreases by 0.5 units from its pre-industrial value of 8.2, corresponding to a 62% reduction in surface [CO₃²⁻]. Changes in surface ocean chemistry largely follow the change in atmospheric CO₂ (figures 2(g)–(i), figure 4). When atmospheric CO₂ returns from its peak value of 1120 ppm to the pre-industrial level of 280 ppm, surface ocean chemistry fields such as pH, [CO₃²⁻], and Ω_dissolve largely return to their pre-industrial values (figure 2(g)–(i), figure 4). In some regions, particularly the Arctic Ocean, relatively large difference in surface chemistry between year 1 and 280 is observed (figure 4). This is consistent with the fact that in the Arctic after the reduction of atmospheric CO₂, the ocean continues to absorb CO₂ from the atmosphere (figure S5). As a result, surface DIC concentration in the Arctic (figure S5), compared to that of the globe ocean (figure 2(f)), is more dependent on the pathway of atmospheric CO₂.

Penetration of excess CO₂ into the ocean makes the ocean interior more acidic. For the simulation with climate change, at the time of 4×CO₂, the ocean mean pH decreases by 0.11 units, corresponding to a 22% decrease of [CO₃²⁻] (figures 1(j), (k)). Unlike the change in surface ocean chemistry that largely follows changes in atmospheric CO₂, the change of deep ocean chemistry exhibits substantial time lag relative to the change of atmospheric CO₂ (figures 2(j)–(l)) because of the long time scale associated with the dynamic processes that transport CO₂ from the surface to the deep ocean. For example, in our analysis at pre-industrial time, 32% of the ocean is supersaturated with respect to aragonite. At the time of 4×CO₂ only 11% of the ocean is supersaturated, and when atmospheric CO₂ returns to its pre-industrial level, 17% of the ocean is supersaturated with respect to aragonite (figure 1(n)). Also, at the time when atmospheric CO₂ returns to its pre-industrial level, the ocean below about 1000 m is more acidic relative to that when atmospheric CO₂ reaches 4×CO₂ (figure 5). It is interesting to note that the pathway for the change of ocean carbonate fields differs from that for ocean temperature (figure S6). This can be explained by the different geographical distribution.

Figure 2. The same as figure 1, but for model-simulated time series of annual and global mean variables against changes in atmospheric CO₂. Corresponding 1000-year simulation results are given in figure S4.
of heat and carbon fluxes, which result in different transport of temperature and carbon from surface to the deep ocean (Frölicher et al 2013).

One important metric of deep ocean acidification is the depth of saturation horizon (the limit between saturation and undersaturation of calcium carbonate minerals) (Orr et al 2005). In the simulation with climate change global mean aragonite saturation horizon shoals from its pre-industrial value of 1288 to 143 m at the time of 4 × CO₂ (figure 1(m)). When atmospheric CO₂ starts to decrease, ASH begins to deepen, but the linear rate of deepening (3.6 m per year) during the period with atmospheric CO₂ decrease is much smaller than the linear rate of shoaling (9.5 m per year) during the period with atmospheric CO₂ increase. After atmospheric CO₂ returns to the pre-industrial level, simulated ASH is shallower than its pre-industrial depth in most parts of the ocean (figures 3, 4 and 6). Global mean ASH is 630 m when atmospheric CO₂ returns to 280 ppm, which is approximately the value of ASH at year 94 when atmospheric CO₂ reaches 719 ppm. At year 1000 global mean ASH is still 95 m shallower than its pre-industrial value (figures S3(m), S4(m)), indicating the long time scale associated with the recovery of deep ocean chemistry.

The effect of climate change on ocean acidification can be inspected by comparing modeled ocean carbonate fields in the simulation with climate change and without it (table 1). Globally, climate change decreases the amount of oceanic uptake of atmospheric CO₂ mainly as a result of surface ocean warming that decreases the solubility of CO₂ and reduced ocean ventilation and deep water formation that decreases the transport of CO₂ from the surface to deep ocean. At the ocean surface climate change has negligible effect on pH, but modestly increases [CO₃²⁻]. The decoupled response of pH and [CO₃²⁻] to climate change is mainly a result of

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**Figure 3.** Model-simulated latitude-depth distribution of the change (relative to year 1) in dissolved organic carbon (DIC) concentration at year 70, 140, and 280 for the Atlantic–Arctic and Pacific-Indian basins. Results are shown for the simulations with climate change (a, c) and without it (b, d), respectively. Overplotted with the distribution of DIC anomaly is the aragonite saturation horizon in the corresponding year (solid lines) and pre-industrial time (dashed lines).

**Table 1.** Model-simulated ocean chemistry change (relative to year 1) at year 140 (first numbers) and 280 (second numbers) for the simulations with and without climate change.

<table>
<thead>
<tr>
<th></th>
<th>With climate change</th>
<th>Without climate change</th>
</tr>
</thead>
<tbody>
<tr>
<td>Cumulative CO₂ uptake (PgC)</td>
<td>632/409</td>
<td>690/513</td>
</tr>
<tr>
<td>Δ surface mean pH</td>
<td>−0.5/−0.03</td>
<td>−0.5/−0.02</td>
</tr>
<tr>
<td>Δ ocean mean pH</td>
<td>−0.11/−0.07</td>
<td>−0.11/−0.08</td>
</tr>
<tr>
<td>Δ surface mean [CO₃²⁻] (µmol kg⁻¹)</td>
<td>−125/−6</td>
<td>−134/−8</td>
</tr>
<tr>
<td>Δ ocean mean [CO₃²⁻] (µmol kg⁻¹)</td>
<td>−19/−11</td>
<td>−21/−15</td>
</tr>
<tr>
<td>Δ aragonite saturation horizon (m)</td>
<td>−1145/−658</td>
<td>−1178/−715</td>
</tr>
</tbody>
</table>

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of heat and carbon fluxes, which result in different transport of temperature and carbon from surface to the deep ocean (Frölicher et al 2013).
Different thermodynamic dependence of these two variables on temperature (McNeil and Matear 2007). In the deep ocean climate change acts to reduce the extent of ocean acidification (figures 1(j)–(o), figure 5) mainly due to decreased ocean ventilation and North Atlantic deep water formation that slows the transport of excess CO$_2$ into the deep ocean. For example, at the time of 4 × CO$_2$ global mean ASH is 143 m in the simulation with climate change, compared to the value of 110 m in the simulation without it. When atmospheric CO$_2$ returns to pre-industrial level, global mean ASH is 630 and 573 m for the simulation with and without climate change, respectively (figures 1(m), 2(m)).

4. Discussion and conclusions

Here we use an idealized CO$_2$ ramp up and ramp down scenario to investigate the response of ocean carbon cycle and ocean chemistry to a gradual increase and then decrease of atmospheric CO$_2$. Atmospheric CO$_2$ concentration pathway used here can be considered as an idealized scenario of carbon dioxide removal geoengineering. The decrease of atmospheric CO$_2$ by 1% per year to its pre-industrial level can in principle be achieved by some proposed means of CDR geoengineering such as direct capture of CO$_2$ from the ambient air (Keith 2009). In our simulations with the effect of climate change, considering air–sea exchange of CO$_2$ alone (i.e. assuming a neutral terrestrial biosphere), to decrease atmospheric CO$_2$ from 4 × CO$_2$ to pre-industrial level at a rate of 1% per year implies a negative CO$_2$ emission of 1990 PgC. By including CO$_2$ exchange with the terrestrial biosphere, to decrease atmospheric CO$_2$ from 4 × CO$_2$ to pre-industrial level at a rate of 1% per year implies a negative CO$_2$ emission of 2211 PgC. If we consider atmospheric response alone, to reduce atmospheric CO$_2$ from 4 × CO$_2$ of 1120 to 280 ppm indicates a removal effort of 1764 PgC (assume 1 ppm = 2.1 PgC). Therefore, our results suggest that as a result of CO$_2$ release from both the ocean and land in response to the reduced burden of atmospheric CO$_2$, the amount of CO$_2$ needs to be removed from the atmosphere is larger than what is expected from the consideration of atmospheric response alone.

Our simulation shows that the recovery of surface ocean chemistry largely follows the decrease of atmospheric CO$_2$, but as a result of slow penetration of excess CO$_2$ to the deep ocean the recovery of deep ocean chemistry lags substantially behind the reduction of atmospheric CO$_2$. Ocean acidification not only poses a threat to marine ecosystems near the surface ocean, but also threatens marine organisms and ecosystems.
Aragonite saturation horizon and deep-sea cold-water coral. (left) model-simulated aragonite saturation horizon at year 1, 70, 140, and 280. White regions are where calculated $\Omega_{\text{aragonite}}$ is greater than one at the bottom of the ocean (i.e., no aragonite saturation horizon is found when the bottom of the ocean has been reached). Areas in pink represent regions where aragonite saturation horizon has reached surface. Overplotted on the maps are cold-water coral locations with different colors representing corals at different depth range. The longitude, latitude, and depth information for cold-water coral locations is from Freiwald et al (2005); (right) histograms showing the number of cold-water corals in each aragonite saturation bin of seawater surrounding the coral locations. What is shown here is from the simulation with climate change. A version from the simulation without climate change is shown in figure S7.

dwelling in the ocean depth (Fabry et al 2008). For example, global distribution of deep-sea cold-water scleractinian corals is at least partly limited by the depth of the aragonite saturation horizon, and a shoaling of the aragonite saturation horizon and an overall decrease in saturation state of aragonite could alter the distribution of deep-sea corals and marine ecosystems depending on them (Guinotte et al 2006). In our analysis, during the pre-industrial period, 9% (or 419) of deep-sea coral locations are surrounded by seawater that is undersaturated with aragonite ($\Omega < 1$). With increasing atmospheric CO$_2$, seawater surrounding deep-sea corals becomes more acidic (figure 1(o), figure 6). At the time of 4 $\times$ CO$_2$, 73% (or 3413) of deep-sea coral locations are surrounded by seawater that is undersaturated with aragonite for the simulation with climate change, compared to 88% (or 4083) in the simulation without climate change. When atmospheric CO$_2$ returns to the pre-industrial level, 18% (or 844) of deep-sea coral locations are supersaturated with aragonite for the simulation with climate change, compared to 23% (or 1061) in the simulation without climate change (figure 1(o)). The above analysis demonstrates the difficulty for deep-sea corals to recover their natural chemical habitat even if atmospheric CO$_2$ can be lowered in the future.

The embryos of krill, one critical component of the marine ecosystem in the Southern Ocean, are reported to sink down to a depth of about 700–1000 m before hatching (Quetin and Ross 1984), and hatch rate is found to be sensitive to ocean acidification (Kawaguchi et al 2013). In our simulations the average ASH in the Southern Ocean (here we define Southern Ocean to be the ocean South of 55$^\circ$S, which is the approximate northern limit of krill habitat as defined in Kawaguchi et al (2013)) is 1039 m during pre-industrial time. At the time of 4 $\times$ CO$_2$ ASH in the Southern Ocean shoals to surface (figure 6). When atmospheric CO$_2$ returns to the pre-industrial level, averaged ASH in the Southern Ocean is 253 and 278 m in the simulation with climate and without it, respectively. The
shoaling of ASH and substantial delay for the recovery of ASH has great implication for the resilience of krill and the Southern Ocean ecosystems in response to the change in atmospheric CO₂.

Because of the long time scale associated with ocean mixing and circulation that transport CO₂ from the surface to the deep ocean, the threats of increasing atmospheric CO₂ to deep ocean biota may not be felt immediately. On the other hand, the time lag in the recovery of deep ocean chemistry relative to the decrease of atmospheric CO₂ indicates that if future atmospheric CO₂ can be lowered to a level that would be ‘safe’ for global climate and surface ocean organisms, it could continue to pose a threat to deep-ocean biota and ecosystems for many years. This study does not include the interactive feedback from deep-sea carbonate sediment, which could buffer part of the deep ocean chemistry change on a time scale longer than a millennium (Archer et al 1998). Nevertheless, the long time scale for deep ocean to recover to the unperturbed ocean chemical states emphasizes the need to avoid continuing accumulation of atmospheric CO₂ through a deep reduction of anthropogenic CO₂ emission.

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