The reversibility of sea level rise

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ABSTRACT

During the last century, global climate has been warming, and projections indicate that such a warming is likely to continue over coming decades. Most of the extra heat is stored in the ocean, resulting in thermal expansion of seawater and global mean sea level rise. Previous studies have shown that after CO2 emissions cease or CO2 concentration is stabilized, global mean surface air temperature stabilizes or decreases slowly, but sea level continues to rise. Using idealized CO2 scenario simulations with a hierarchy of models including an AOGCM and a step-response model, the authors show how the evolution of thermal expansion can be interpreted in terms of the climate energy balance and the vertical profile of ocean warming. Whereas surface temperature depends on cumulative CO2 emissions, sea level rise due to thermal expansion depends on the time profile of emissions. Sea level rise is smaller for later emissions, implying that targets to limit sea level rise would need to refer to the rate of emissions, not only to the time integral. Thermal expansion is in principle reversible, but to halt or reverse it quickly requires the radiative forcing to be reduced substantially, which is possible on centennial time scales only by geoengineering. If it could be done, the results indicate that heat would leave the ocean more readily than it entered, but even if thermal expansion were returned to zero, the geographical pattern of sea level would be altered. Therefore, despite any aggressive CO2 mitigation, regional sea level change is inevitable.

1. Introduction

Model-based studies have explored the evolution of climate in scenarios where atmospheric CO2 concentrations rise at various rates, following which CO2 is stabilized or the emissions are stopped (Meehl et al. 2005; Plattner et al. 2008; Solomon et al. 2009; Frölicher and Joos 2010; Gillett et al. 2011). In the latter case, global mean surface air temperature subsequently stays approximately constant or declines slowly as CO2 is removed from the atmosphere by natural processes. In some scenarios, including the later part of the policy-relevant Representative Concentration Pathway 2.6 (RCP2.6; Moss et al. 2010), which will be assessed in the next report of the Intergovernmental Panel on Climate Change (IPCC), CO2 is assumed to be artificially removed from the atmosphere at greater rates than natural uptake, producing a faster reduction in forcing and a faster cooling. However, sea level from thermal expansion continues to rise for many centuries following a stabilization of CO2 concentration in the atmosphere or a cessation of CO2 emissions (Meehl et al. 2005; Lowe et al. 2006; Plattner et al. 2008; Solomon et al. 2009; Frölicher and Joos 2010; Gillett et al. 2011). This raises the question of what it would take to reverse sea level rise.

In this work, we use a 3D atmosphere–ocean general circulation model (AOGCM) to explore how the emissions pathway affects the peak sea level and the subsequent decline. For simplicity, we consider CO2-only scenarios, since CO2 is the dominant forcing. We consider only the global mean thermosteric sea level, and
not the additional component from glaciers and the ice sheets of Greenland and Antarctica (Charbit et al. 2008; Ridley et al. 2009; Vizcaíno et al. 2010; Huybrechts et al. 2011). The thermosteric term is influenced by ocean interior transport processes, including large-scale oceanic circulations such as the Atlantic meridional overturning circulation (AMOC). In the deep ocean, temperature changes caused by these processes occur on a hundred- to thousand-year time scale. Although this is slow, the thermosteric sea level change may be reversed more quickly than ice sheet mass loss, which would require a longer time scale of thousands of years (Charbit et al. 2008; Ridley et al. 2009). Likewise, reversing the sea level change induced by glaciers and ice caps would be a slower process. Thermosteric sea level change is subject to less (though still substantial) scientific uncertainty than ice sheets and glaciers, and is sufficient alone to cause significant sea level rise impacts for many centuries into the future.

2. Method

We run simulations with the Fast Met Office/UK Universities Simulator (FAMOUS) AOGCM (Smith et al. 2008). FAMOUS is a low-resolution version of the third climate configuration of the Met Office Unified Model (HadCM3) AOGCM (Gordon et al. 2000); its atmosphere component runs on a 5° latitude by 7.5° longitude grid with 11 levels and its ocean component on a 2.5° latitude by 3.75° longitude grid with 20 levels. It is structurally almost identical to HadCM3 and produces climate and climate-change simulations that are reasonably similar to HadCM3, but runs about 20 times faster and is hence particularly useful for investigations involving many long integrations. Despite the coarse resolution, the present-day climatology of FAMOUS is adequate without flux correction.

The simulations employ idealized scenarios in which atmospheric CO₂ is first increased and then either held constant or abruptly decreased. The increase follows three pathways (Fig. 1a): an instantaneous increase of CO₂ to 4 times the preindustrial value (4×CO₂; the preindustrial value is 280 ppm), a 1% increase of CO₂ yr⁻¹ (1%CO₂) for 140 yr, and a 2% increase of CO₂ yr⁻¹ (2%CO₂) for 70 yr. Both the 1%CO₂ and 2%CO₂ ramps end at 4 times the preindustrial value. Thus, at t = 140 yr the CO₂ concentration is the same in all scenarios, but the three pathways that have led to that level differ. Beyond t = 140 yr the CO₂ concentration is either kept constant at this value or instantaneously reduced, to either the preindustrial value (1×CO₂) or half of it (0.5×CO₂), and then held constant at the reduced value; in section 4c we also consider a scenario in which CO₂ is ramped down. We consider these idealized scenarios because they give a clear demonstration of the qualitative behavior of sea level. In particular, this is the reason for choosing the 0.5×CO₂ concentration (140 ppm), which is of course unrealistic for natural carbon sinks, being lower than the concentration at glacial maxima such as during the Last Glacial Maximum (atmospheric CO₂ concentration of around 190 ppm approximately 21 000 yr ago).

As the CO₂ concentrations scenarios are idealized, they are not evaluated in terms of feasibility or corresponding emissions. For the latter, a coupled carbon–climate model would be needed, which would include the fact that if
emissions are cut to zero, the biosphere and ocean would progressively take up less carbon as they saturate. This would suggest that to achieve the assumed rates of CO₂ reduction, actively removing CO₂ from the atmosphere by artificial means may be needed, since the CO₂ would otherwise stay in the atmosphere for a long time (several thousand years; Eby et al. 2009). The reduction of forcing does not necessarily require CO₂ reduction and could possibly be obtained by other suggested geo-engineering mechanisms (Shepherd et al. 2009), but the feasibility of this is not further discussed here. We calculate sea level rise due to thermal expansion from ocean temperature change diagnosed from the model. As found previously (Russell et al. 2000), global mean thermosteric sea level rise \( \eta \) and global mean ocean temperature rise \( T_o \) (equivalent to the increase in ocean heat content) are in a nearly linear relationship (Fig. 2) and so we speak of them interchangeably. In the following sections, we interpret the qualitative features of results of the AOGCM for surface air temperature and sea level change using a range of simpler models (summarized in Table 1).

3. Surface air temperature change

During the first part of the simulations (the “high-CO₂ phase”), while the atmospheric CO₂ concentration is increasing, the change of global mean surface air temperature \( T_s \) tracks the CO₂ concentration (Figs. 1a,b). In the case where the CO₂ concentration is instantaneously increased, \( T_s \) takes a couple of decades to respond. After 140 yr, \( T_s \) is the same in the three simulations. It has similarly been shown that, under scenarios in which emissions peak and decline, the peak warming attained depends only on the cumulative CO₂ emission and is largely insensitive to the emissions pathway up to the time of maximum temperature (Allen et al. 2009a; Matthews et al. 2009; Zickfeld et al. 2009; Bowerman et al. 2011).

Table 1. Summary of the main results from the models.

<table>
<thead>
<tr>
<th>Model</th>
<th>Main features of the results</th>
<th>Successfully explains</th>
<th>Fails to explain</th>
</tr>
</thead>
<tbody>
<tr>
<td>FAMOUS AOGCM (Smith et al. 2008)</td>
<td>Sea level evolution depends on the emission pathway</td>
<td>Sea level rise can be reversed without negative forcing (the model assumes heat loss into an infinite heat sink that does not warm up and therefore does not give back heat spontaneously)</td>
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<tr>
<td></td>
<td>Sea level continues to rise after stabilization</td>
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<td></td>
<td>Sea level rise is reversed if forcing becomes zero or negative</td>
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<tr>
<td></td>
<td>Sea level rise and fall are not symmetric</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Zero-layer model</td>
<td>Sea level rise depends on the emission pathway</td>
<td></td>
<td>Asymmetry of the sea level fall and rise (the single lower layer has too much heat capacity and does not warm up very much, so heat flux depends almost entirely on the upper temperature)</td>
</tr>
<tr>
<td>(Gregory and Mitchell 1997; Raper et al. 2002; Gregory and Forster 2008; Held et al. 2010)</td>
<td>Sea level continues to rise after stabilization</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Surface climate energy balance</td>
<td>Features explained successfully by the zero-layer model</td>
<td></td>
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<tr>
<td></td>
<td>Sea level rise is reversible: it falls during low-CO₂ phase</td>
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<tr>
<td></td>
<td>(the model is able to distinguish surface temperature from heat content and can lose heat spontaneously)</td>
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<tr>
<td>Two-layer model</td>
<td>Features explained successfully by the zero-layer model</td>
<td></td>
<td></td>
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<tr>
<td>(Gregory 2000; Held et al. 2010)</td>
<td>Sea level rise is reversible: it falls during low-CO₂ phase</td>
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<td>Asymmetry of the sea level fall and rise (the single lower layer has too much heat capacity and does not warm up very much, so heat flux depends almost entirely on the upper temperature)</td>
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</tr>
<tr>
<td>Step model</td>
<td>Features explained successfully by the two-layer model</td>
<td></td>
<td>Remaining part of the asymmetry (due to the different penetration of heat during warming and cooling, because the step model does not have any functional dependence on vertical stability)</td>
</tr>
<tr>
<td>(Good et al. 2011)</td>
<td>Part of the asymmetry between the sea level fall and rise</td>
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<td></td>
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<tr>
<td></td>
<td>(because of multiple layers in the ocean)</td>
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The second part of the simulations (the “low-CO2 phase”) begins at year 141, when the CO2 concentration is either stabilized or lowered. After CO2 stabilizes, the rate of surface air temperature increase is greatly reduced (Fig. 1b). If the radiative forcing abruptly returns to zero, $T_s$ falls most of the way to zero within around two decades. If CO2 is set to a value smaller than the initial one, the radiative forcing becomes negative, and $T_s$ falls below its initial value. It is apparent that $T_s$ depends roughly linearly on the forcing (Figs. 1a and 1b are similar in form), as can be understood from simple “zero-layer” heat balance equations for time-dependent climate change (Gregory and Mitchell 1997; Raper et al. 2002; Gregory and Forster 2008; Held et al. 2010; see Table 1):

$$F = (\alpha + \kappa)T_s \Rightarrow T_s = \frac{F}{\alpha + \kappa},$$

where $\alpha$ is the climate feedback parameter, $\kappa$ is the ocean heat uptake efficiency, and the radiative forcing $F$ depends roughly linearly (actually logarithmically) on the atmospheric CO2 concentration $C$. In this zero-layer model, there is no feedback of ocean temperature on the rate of heat uptake and the ocean is effectively infinitely deep (i.e., it can absorb heat indefinitely).

$C$ is equivalent to the increase in the CO2 burden of the atmosphere (1 GtC = 0.47 ppm). Hence $C$ is determined uniquely at each time by the cumulative CO2 emissions $E$ through the relationship $C = AE$, provided the airborne fraction $A$ is constant. Observations indicate that $A$ has been approximately constant during recent decades (Knorr 2009), but it could change in the future (Friedlingstein et al. 2006). How much it might change appears to depend on both the emissions scenario (Gloor et al. 2010) and the behavior of the carbon cycle (Friedlingstein et al. 2006), which is uncertain. For the shapes of scenario considered here, and assuming that a linear model provides a reasonable first-order representation of the carbon cycle (Joos et al. 1996), then we would expect a constant value of $A$ to also be an acceptable approximation. If so, $T_s$ will track $E$ (Matthews et al. 2009; Gregory et al. 2009).

4. Sea level change

a. Dependence of sea level rise on the pathway of emissions

Sea level behaves qualitatively differently from $T_s$. The evolution of sea level rise due to thermal expansion depends on the pathway of emissions (Stouffer and Manabe 1999; Zickfeld et al. 2012) and reaches different values at 140 years despite the CO2 concentration and $T_s$ being the same at that time (Fig. 1c). Comparing Figs. 1c and 1d, we see that sea level rise depends almost linearly on the integral of the radiative forcing while the forcing is positive (high-CO2 phase). This is consistent with the zero-layer model, in which the rate of ocean heat uptake $N$ is proportional to $T_s$,

$$N = \kappa T_s = F \times \frac{k}{\alpha + \kappa},$$

and hence the rate of sea level rise depends on $F$. However, the actual sea level rise is given by

$$\eta \approx \int N \, dt = \int F \times \frac{k}{\alpha + \kappa} \, dt = \frac{k}{\alpha + \kappa} \int F \, dt,$$

provided $\kappa$ and $\alpha$ are constant (Gregory and Forster 2008).

Thus, because the thermosteric sea level is proportional to the time integral of the radiative forcing, the longer the forcing lasts, the bigger the change in thermosteric sea level rise (and oceanic temperature). The sea level change is the largest in the simulation where the largest fractional increase in the atmospheric burden of CO2 occurs the earliest, which implies where CO2 has been emitted the earliest. Because of a focus on impacts related to peak warming, it has been proposed that policy targets for avoiding dangerous climate change might be set in terms of cumulative CO2 emission (Allen et al. 2009b). However, if we are concerned with mitigating sea level impacts, targets must be set on the rate of emission as well.

b. Long-term commitment to sea level rise

If CO2 is stabilized, ocean temperature and sea level continue to rise (Fig. 1c), in contrast to $T_s$ (Fig. 1b). After the CO2 concentration is stabilized, because $T_s$ is approximately the same in the three stabilization simulations (Fig. 1b; $\sim$5.5°C), the rate of thermosteric sea level rise is nearly the same in all three. That is, the three solid lines in Fig. 1c are nearly parallel; the rate is $\sim$4 mm yr$^{-1}$, much higher than the recent rate of 0.9 mm yr$^{-1}$ for 1993–2008 (Church et al. 2011). Thus the zero-layer model still explains the evolution of both temperature and sea level in the AOGCM for at least a century after stabilization of the CO2 concentration.

c. Reversibility of sea level rise

The experiments with stabilized CO2 may give an impression that sea level rise is irreversible, but if CO2 is returned instantaneously to its initial value, sea level immediately falls, albeit at a slow and decreasing rate so that after 160 years it has fallen only half the way to its initial level. Any practical measures to remove CO2...
from the atmosphere would produce a gradual rather than an instantaneous decrease; if CO\textsubscript{2} is ramped down, the oceanic temperature initially continues to rise, then gradually stabilizes, and then begins to fall. This is illustrated by a simulation in which CO\textsubscript{2} increases at 1% yr\textsuperscript{-1} for 140 yr and then decreases at 1% yr\textsuperscript{-1} for the next 140 yr (cyan lines in Fig. 1). In this experiment the time-integral forcing is the same as in the scenario of 4\times CO\textsubscript{2} followed by 1\times CO\textsubscript{2} (dotted red lines); the later reduction in CO\textsubscript{2} means that sea level is higher following the ramp-down. If the radiative forcing becomes negative (dashed lines), sea level falls back more rapidly to its initial value. Thus, sea level rise is in principle reversible, provided that the radiative forcing is reduced sufficiently. A long-term multi-century commitment to rising thermal expansion is not physically inevitable, although it may be practically so.

However, the zero-layer model fails to explain the sea level reversibility in the AOGCM. Since the sea level rise is proportional to the integral of the forcing \( F \), it predicts that sea level will remain constant if the forcing is returned to zero, and will fall only if a negative forcing is applied. The zero-layer model assumes heat loss into an infinite heat sink that does not warm up and therefore does not give back heat spontaneously. It cannot account for sea level falling under zero forcing because it neglects changes happening below the ocean surface layer. In reality, the ocean is finite and more than one time scale is relevant. The rapid time scale of one or two decades relates to the ocean surface layer and accounts for most of the response of surface air temperature to the radiative forcing following abrupt change in CO\textsubscript{2}. The time scale of the “re-calcitrant” response (Held et al. 2010) is another time scale that applies to the evolution of sea level.

This can be illustrated with a two-layer model (Gregory 2000; Held et al. 2010; Table 1). The two-layer model is composed of an ocean with an upper layer of thickness \( d_u \) (100 m) and temperature \( T_u \), and a deep layer of thickness \( d_l \) (2000 m) and temperature \( T_l \), where both temperature are perturbations from an equilibrium with zero surface heat flux. The upper layer has little heat capacity and plays the role of a regulator; it absorbs the forcing, determines the radiation of heat to space, and is thermally coupled to the lower layer, which has large heat capacity and provides inertia. The heat flux between the two is proportional to their temperature difference. The temperatures thus follow the two equations

\[
\begin{align*}
c d_u \frac{dT_u}{dt} &= F - ck \frac{(T_u - T_l)}{0.5 \times (d_l + d_u)} - \alpha T_u \quad \text{and} \\
c d_l \frac{dT_l}{dt} &= ck \frac{(T_u - T_l)}{0.5 \times (d_l + d_u)} ,
\end{align*}
\]

with \( c = 4.218 \times 10^6 \text{ J m}^{-3} \text{ K}^{-1} \) being the volumetric heat capacity, \( F \) (W m\textsuperscript{-2}) the surface heat flux, \( k = 1 \times 10^{-4} \text{ m}^2 \text{ s}^{-1} \) the thermal diffusivity, and \( \alpha = 0.81 \text{ W m}^{-2} \text{ K}^{-1} \) the climate feedback parameter.

When the radiative forcing \( F \) changes abruptly, \( T_u \) rapidly reaches a new near-equilibrium (Fig. 3a), in which \( F \) is nearly balanced by transfer of heat to or from the lower layer. If \( F \) reverts to zero, the upper layer rapidly becomes cooler than the lower layer, which therefore loses heat (Fig. 3b, dotted lines). If a negative forcing is applied, the upper layer gets still colder, the temperature difference between the two layers is greater, the transfer of heat from the lower to the upper ocean is faster, and the lower layer cools down more rapidly (Fig. 3b, dashed lines). Unlike the zero-layer model,
the two-layer model reproduces remarkably well these features of the evolution of surface temperature and thermosteric sea level in the much more complex AOGCM (Figs. 3a and 3b are qualitatively very similar to Figs. 1b and 1c). The two-layer model thus accounts for the reversibility of sea level rise when the radiative forcing diminishes because it has a finite ocean and the model is able to distinguish surface temperature from heat content so that it can lose heat spontaneously.

d. Asymmetry of sea level evolution

In the zero-layer model, the ocean temperature and sea level change should be back to zero when the integral of $F$ returns to zero (i.e., after a negative forcing of the same time integral as the one during the high-CO$_2$ phase). For example, for the 1×CO$_2$ scenario followed by 0.5×CO$_2$, this is at 280 yr, but actually the sea level in the AOGCM reaches zero sooner, at 224 yr.

To quantify better the link between the ocean temperature change and the radiative forcing, we have run additional AOGCM simulations (Fig. 4), beginning with a high-CO$_2$ phase of either constant 4×CO$_2$ or a 1% CO$_2$ ramp-up. In these simulations, CO$_2$ is reduced abruptly to 0.5×CO$_2$ after various times between 10 and 100 yr (instead of 140 yr as in the scenario of Fig. 1).

During the low-CO$_2$ phase, the time for the oceanic temperature to cool down to its initial value depends linearly on the oceanic warming achieved during the high-CO$_2$ phase (Fig. 5a), and hence on the integral of the forcing during that phase (Fig. 5b). However, it takes less negative integral radiative forcing $F$ (or less time with a symmetrical forcing evolution) for the oceanic
temperature to go back to its initial value (Fig. 5c) \(|\int F_{\text{down}} \, dt| < |\int F_{\text{up}} \, dt|\). Thus, the relationship between the oceanic temperature change and the integral of the forcing is different during the high-CO2 and low-CO2 phases (Fig. 5d). The amount of negative radiative forcing needed for the sea level to go back to its initial value is linearly related to the amount of positive radiative forcing:

\[ |\int F_{\text{down}} \, dt| \approx 0.58 |\int F_{\text{up}} \, dt| \]

The numerical factor depends on the scenario and model.

Although it exhibits reversibility, the two-layer model cannot explain why the ocean cools down more effectively than it warms up. In the two-layer model, the relationship between ocean temperature and \(\frac{dT}{dt}\) is the same in the high-CO2 and low-CO2 phases (Fig. 3c), unlike in the AOGCM (Fig. 5d). The lower layer is a very large well-mixed heat capacity that sets the long time scale for approach to a steady state. Consequently, the lower layer warms much less than the upper layer in the high phase (at 140 yr in the 1% scenario, \(T_u = 4.4^\circ C\) in Fig. 3a and \(T_l = 0.4^\circ C\) in Fig. 3b). Because \(T_u\) closely tracks \(F\), and \(T_l \ll T_u\), the heat flux \(\propto T_u - T_l\) between the layers has roughly the same time profile as \(F\), and \(T_l \propto \int F \, dt\). Because the single lower layer has too much heat capacity and does not warm up very much, the heat flux depends almost entirely on the upper temperature and hence on the integral of the radiative forcing. More layers would be needed to explain the faster loss of heat during the low-CO2 phase.

To account for the asymmetry, we apply the step-response model of Good et al. (Good et al. 2011; Table 1). In this model, the evolution of any climate variable \(X(t)\) is estimated by regarding it as the linear superposition of the responses to a succession of small instantaneous step-forcing changes \(\delta F(t)\) in forcing, such that \(F(t) = \int_0^t \delta F(t') \, dt'\). We carry out a single AOGCM experiment in which CO2 is quadrupled at \(t = 0\), giving forcing \(F(4 \times \text{CO2})\), and then held constant. If \(X = X(t)\) in this experiment, we estimate that \(X(t) = \int_0^t X(t - t') \delta F(t')/F(4 \times \text{CO2}) \, dt'\)—that is, a convolution of the response to the \(4 \times \text{CO2}\) step with the forcing time series. Good et al. (2011) show that this method is accurate for surface temperature and has skill at reproducing ocean heat uptake. We use the step model to study the experiment with 1%CO2 followed by 0.5 \(\times\) CO2 as an example.

The step model successfully reproduces the evolution of the surface air temperature \(T_s\) and ocean temperature \(T_o\) during the high-CO2 phase (Figs. 6a,b). (Note that \(T_o\) from the step model lacks the variability of the AOGCM because the convolution tends to smooth it.)
FIG. 6. Results for the climate response of the AOGCM and the step model in response to 1%CO$_2$ for 140 yr, then 0.5×CO$_2$. (top) Time series of (a) global mean surface air temperature and (b) global mean oceanic temperature. (c)–(f) zonal mean ocean temperature at (left) $t = 140$ yr and (right) $t = 224$ yr (when the mean oceanic temperature change is back to 0°C), for (c),(d) the AOGCM and (e),(f) the step model.
The step model estimate of the ocean temperature change latitude–depth distribution is also very good (cf. Figs. 6c,e), showing the warming spreading downward from the surface, with deeper penetration at high latitudes. The step model gives a better approximation of the evolution during the low-CO2 phase than the zero- and two-layer models (Table 1). In particular, it correctly predicts that \( T_s \) returns to its initial value before year 280; it is qualitatively superior to the two-layer model in partly reproducing the asymmetry between the warming and cooling phases (Fig. 7). This is because the step model, like the AOGCM, has many layers and time scales. Unlike in the two-layer model, the ocean below the surface is not well mixed. The shallower layers have relatively small heat capacity and warm substantially during the high-CO2 phase, and as time passes the warming spreads to deeper layers. During the low-CO2 phase, the shallow layers likewise cool rapidly. In the AOGCM, this causes a large temperature contrast between the newly cooled surface and previously warmed subsurface layers, which forces heat out of the ocean more rapidly than it entered. In the step model, the large temperature contrast and rapid heat loss are predicted as a consequence of the large negative step in forcing from \( 4\times CO_2 \) to \( 0.5\times CO_2 \). In both models, as a consequence of the high-CO2 phase, additional heat is still being transported to deeper layers during the low-CO2 phase at the same time as the cooling penetrates into the surface. In the global mean, similar qualitative behavior would be obtained from a vertical diffusion model of ocean temperature (e.g., Marčelj 2010). The advantages of the step model are that it can be applied in 3D, and that it emulates the AOGCM without any tuning of parameters being required.

However, the step model shows some inaccuracies. In particular, \( T_s \) falls too quickly at the start (Fig. 6a), and after about 30 years the step model underestimates the oceanic cooling progressively more seriously (Fig. 6b). These phenomena can be linked: both are related to an insufficient heat flux from lower layers toward the surface. Correspondingly, the cooling anomaly does not penetrate as deeply at high latitudes in the step model as in the AOGCM simulation, while in middle and low latitudes the thermocline does not cool as quickly (cf. Figs. 6d and 6f). The asymmetry between high-CO2 and low-CO2 phases is thus not as pronounced as in the AOGCM (Fig. 7).

We infer that these remaining discrepancies arise from nonlinear behavior of the AOGCM that the step model does not capture. In particular, the responses of the AOGCM to positive and negative step changes in forcing of equal magnitude are not equal and opposite (Fig. 8). This asymmetry cannot be captured by the step model, which is constructed by using the response to a positive step only, and presumably arises from the dependence of vertical heat transport on stability in the AOGCM (Stouffer and Manabe 1999). In particular, for the same absolute value of radiative forcing, the warming penetrates relatively deeper in the Arctic whereas

**FIG. 7.** Global mean ocean temperature change (°C) as a function of the integral of the radiative forcing during the phase with high CO2 (1%CO2, solid lines) and continuing under 0.5×CO2 (dashed lines).

**FIG. 8.** Zonal-mean ocean temperature anomaly (°C) as a function of depth after 70 yr of simulation, (a) under \( 4\times CO_2 \) (warming) and (b) under \( 0.25\times CO_2 \) (cooling). Both simulations start from the same initial state. The anomaly is with respect to the control simulation.
the cooling goes deeper in the Southern Ocean (Fig. 8). This different penetration of the heat anomaly during warming and cooling simulations has previously been described by Manabe et al. (1991), with similar results in the Southern Ocean but slightly different penetration of heat in the North Atlantic. After a sequence of warming (1%CO$_2$ experiment) followed by a cooling (0.5×CO$_2$ experiment), this different penetration of the heat anomaly results in a warmer Arctic and cooler Southern Ocean (except in the surface) in the AOGCM than in the step model at the time (224 years) when $T_o$ in the AOGCM is back to its initial value.

e. Regional sea level changes

Impacts of sea level change arise from its regional distribution. As shown in Fig. 6, even when the global mean ocean temperature is back to its initial value, heat in the ocean is distributed differently, with some areas warmer and others colder. This means that there is a still a regional pattern of thermosteric sea level change when $T_o$ returns to 0°C (Fig. 9). Although the global mean $\eta$ is 0 m, sea level is higher in the initial state in the Arctic and Atlantic south of 45°N, and lower in the North Atlantic and southern Indian Ocean. This distribution is likely to be model dependent as the regional sea level changes from model to model (Pardaens et al. 2011), but the same qualitative point would hold. Thus, even if it were practical to eliminate the commitment to global thermosteric sea level change by negative radiative forcing, the commitment to regional sea level change would be yet more recalcitrant.

5. Conclusions

We have studied the future evolution of global mean sea level rise due to thermal expansion (i.e., not including contributions from ice sheets and glaciers) with an AOGCM under idealized CO$_2$ scenarios. Unlike surface temperature change, sea level change depends not only on the cumulative emission of CO$_2$ but also on the emission pathway. A greater rise in sea level results from earlier emissions than from later, for the same cumulative emission. Hence, targets to limit sea level rise would need to refer to rates of emissions as well as the total.

Thermal expansion will continue for many centuries if CO$_2$ emissions cease or if CO$_2$ concentration is stabilized. However, it is in principle reversible. Reducing the radiative forcing sufficiently would halt or reverse it, and a negative forcing would reverse it more quickly. Of course, reducing the forcing would require removal of CO$_2$ from the atmosphere or other geoengineering, which is not yet technologically feasible and might have side effects, while a large negative forcing would produce a climate colder than the preindustrial. Even if sea level returned to its preindustrial global mean, the geographical pattern would be different from the initial one because the penetration of heat is different during warming and cooling.

We have interpreted the behavior of thermal expansion in the AOGCM by comparison with a range of simpler models. The important qualitative difference between surface temperature and sea level is that the former depends on the prevailing radiative forcing and hence the cumulative CO$_2$ emission, while the latter depends on the time integral of radiative forcing and hence the time profile of CO$_2$ emission. The reversibility of thermal expansion cannot be explained without this distinction, which arises because surface temperature relates to the temperature of the upper ocean only, and thermal expansion to the full depth of the ocean. The AOGCM shows that sea level rise and fall are not symmetrical with respect to forcing. When the radiative forcing is reduced, heat leaves the ocean more readily than it entered. This is partly due to the vertical profile of the ocean temperature change, which retains a memory of the time profile of radiative forcing, and partly due to the dependence of vertical heat transport processes on temperature gradients. The success of the step-response model in largely reproducing the AOGCM results indicates that this method could be a useful way to continue investigations into AOGCM ocean heat uptake in a scenario-independent way although consideration of the nonlinear behavior may be necessary.

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