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Millennial total sea-level commitments projected with the Earth system model of intermediate complexity LOVECLIM

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Abstract
Sea-level is expected to rise for a long time to come, even after stabilization of human-induced climatic warming. Here we use simulations with the Earth system model of intermediate complexity LOVECLIM to project sea-level changes over the third millennium forced with atmospheric greenhouse gas concentrations that stabilize by either 2000 or 2100 AD. The model includes 3D thermomechanical models of the Greenland and Antarctic ice sheets coupled to an atmosphere and an ocean model, a global glacier melt algorithm to account for the response of mountain glaciers and ice caps, and a procedure for assessing oceanic thermal expansion from oceanic heat uptake. Four climate change scenarios are considered to determine sea-level commitments. These assume a 21st century increase in greenhouse gases according to SRES scenarios B1, A1B and A2 with a stabilization of the atmospheric composition after the year 2100. One additional scenario assumes 1000 years of constant atmospheric composition from the year 2000 onwards. For our preferred model version, we find an already committed total sea-level rise of 1.1 m by 3000 AD. In experiments with greenhouse gas concentration stabilization at 2100 AD, the total sea-level rise ranges between 2.1 m (B1), 4.1 m (A1B) and 6.8 m (A2). In all scenarios, more than half of this amount arises from the Greenland ice sheet, thermal expansion is the second largest contributor, and the contribution of glaciers and ice caps is small as it is limited by the available ice volume of maximally 25 cm of sea-level equivalent. Additionally, we analysed the sensitivity of the sea-level contributions from an ensemble of nine different model versions that cover a large range of climate sensitivity realized by model parameter variations of the atmosphere–ocean model. Selected temperature indices are found to be good predictors for sea-level contributions from the different components of land ice and oceanic thermal expansion after 1000 years.

Keywords: sea-level change, climate change, ice sheets, glaciers, Earth system models

1. Introduction

It is widely recognized that greenhouse gases have a very long residence time in the atmosphere on centennial to multi-millennial timescales (Archer *et al* 2009). In combination with the inertia of the coupled climate system this gives rise to the notion that greenhouse gases emitted until one point in time will necessarily lead to already committed climate changes many years later (Solomon *et al* 2009). When using coupled climate models to estimate such climate changes, two different types of commitment are distinguished. The additional warming (and associated climate change) that occurs after the composition of the atmosphere has been fixed, is termed 'constant composition commitment' (Meehl *et al* 2005, Wigley 2005). It can be estimated using a wide range of coupled atmosphere–ocean models. If a carbon cycle component is available to simulate changes of the atmospheric composition due to natural processes, an estimate of the 'constant emission commitment' is possible (Meinshausen *et al* 2011). The latter includes the case of zero emissions for which projections show near constant temperature in the thousand years after emissions are ceased (Plattner *et al* 2008, Solomon *et al* 2009, Matthews and Weaver 2010, Gillett *et al* 2011). This effectively means that any realized warming is very likely irreversible on human timescales.

An important consequence of climatic warming is sea-level change. The Antarctic and Greenland ice sheets are by far the largest potential contributors to long-term sea-level rise (e.g. Huybrechts *et al* 2011). Their response times to climate changes cover a wide range up to millennial timescales, longer than those associated with either thermal expansion of the world's ocean or the melting of mountain glaciers and ice caps, the latter of which is self-limited by the amount of available ice. Land ice is not routinely included in Earth system models for long-term climate change projections. For the large polar ice sheets, that is mainly due to computational constraints, whereas for glaciers and ice caps this is mainly because of the difficulties in generalizing the response of the more than 200 000 individual glaciers found in very different climatic and dynamic settings (Radic and Hock 2010). Only a few global climate models currently exist that include dynamically coupled models for the Greenland (Ridley *et al* 2005, 2010, Driesschaert *et al* 2007, Charbit *et al* 2008) and/or Antarctic ice sheets (Mikolajewicz *et al* 2007, Swingedouw *et al* 2008, Vizcaíno *et al* 2010).

Solomon *et al* (2009) and Gillett *et al* (2011) discussed global and regional changes in temperature and precipitation, as well as thermosteric sea-level rise, after total cessation of carbon dioxide emission. These Earth system model studies emphasize the potential for irreversible damage to the climate system and find sea-level rises after 1000 years from thermal expansion alone of at least 20–40 cm for their most optimistic emission cessation scenarios.

Here the scope is extended to include all land ice (glaciers, ice caps and ice sheets) in a consistent framework. We use the Earth system model of intermediate complexity LOVECLIM (Goosse *et al* 2010), which includes bidirectionally coupled models of the Greenland and Antarctic ice sheets (Huybrechts *et al* 2011), a global glacier melt algorithm to account for mountain glaciers and ice caps (Raper and Braithwaite 2006), and a procedure to assess the thermosteric expansion of the ocean. These experiments assume constant atmospheric composition from the years 2000 or 2100 onwards until the end of the third millennium.

Apart from the inherent uncertainties in emission pathways and climate scenario, large uncertainties also arise from the climate sensitivity of a particular climate model. The latter are a consequence of incomplete physics and uncertainties in physical parameters, and of uncertainties in boundary and initial conditions. These uncertainties can be accounted for by employing ensembles of different model versions with variable boundary conditions or parameters and analyse the results for the wide range of model responses (e.g. Murphy *et al* 2004). To better investigate the range of our projections we finally consider a range of model realizations that are all able to reproduce the current climate and sea-level change within the observational error, and conclude with an assessment of the projected sea-level change for each component as a function of the temperature change.

2. Model description

The Earth system model of intermediate complexity LOVECLIM 1.1 (Goelzer *et al* 2011, Loutre *et al* 2011) used here is a further development of version 1.0 (Driesschaert *et al* 2007, Swingedouw *et al* 2008, Huybrechts *et al* 2011). Differences to version 1.2 (Goosse *et al* 2010) are of a technical nature and have no impact on the physical response of the model described here. Therefore, only a brief summary is given here and the reader is referred to Goosse *et al* (2010) for a full model description.

In short, LOVECLIM includes components for the atmosphere (ECBilt), the ocean and sea-ice (CLIO), the terrestrial biosphere (VECODE), the ice sheets (AGISM) and the carbon cycle (LOCH). ECBilt (Opsteegh *et al* 1998) is a spectral atmospheric model with truncation T21, which corresponds approximately to a horizontal resolution of 5.625° in longitude and latitude, and incorporates three vertical levels. It includes simple parameterizations of the diabatic heating processes and an explicit representation of the hydrological cycle. Cloud cover is prescribed according to present-day climatology. CLIO—Coupled Large-scale Ice-Ocean model—is a global free-surface ocean general circulation model coupled to a thermodynamic sea-ice model (Fichefet and Maqueda 1997, Goosse and Fichefet 1999, Goosse *et al* 1999). The horizontal resolution of CLIO is $3° \times 3°$, and there are 20 unevenly spaced levels in the vertical. VECODE—VEgetation COntinuous DEscription model—is a reduced-form model of the vegetation dynamics and of the terrestrial carbon cycle (Brovkin *et al* 1997). LOCH—Liège Ocean Carbon Heteronomous model (Mouchet and François 1996)—is a comprehensive, three-dimensional oceanic carbon cycle model. Since the evolution of CO_2 concentration was prescribed in the present simulations, the carbon cycle component of the model was only used in diagnostic mode. AGISM—Antarctic and Greenland Ice Sheet Model (Huybrechts and de Wolde 1999)—consists of two three-dimensional thermomechanical ice-dynamic models for each of the polar ice sheets (Greenland Ice Sheet, GrIS; Antarctic ice sheet, AIS). The coupling of the ice sheet models (GISM, AISM) to the other components of LOVECLIM and their coupled response to greenhouse warming is described for two schematic CO_2 forcing scenarios in Huybrechts *et al* (2011). The interested reader is referred to the latter paper for a detailed description of the physical mechanisms of ice sheet changes, the retreat patterns and the mass balance components relevant for projecting sea-level contributions on a millennial timescale. Here we focus more on the magnitude of sea-level contributions from the ice sheets and the other relevant components of sea-level change.

Oceanic thermal expansion (OTE) is calculated offline from simulated changes in ocean temperature. The effect of haline contraction due to freshening of the ocean with melt water from the ice sheets and sea-ice is accounted for, but is negligible compared to the amplitude of the other OTE components. Furthermore, we account for the response of mountain glaciers and ice caps (GIC) by using a global glacier melt algorithm (Raper and Braithwaite 2006). The algorithm is run in offline mode and consists of a mass balance and a geometric glacier model. A separation is made between melt contributions from mountain glaciers and ice caps as these have distinctly different geometric characteristics. Their combined potential contribution to global sea-level rise is estimated at 24.6 cm, as glaciers and ice caps in the vicinity of the Greenland and Antarctica ice sheets are excluded from the budget (Raper and Braithwaite 2005). The latter ice masses are however implicitly included in the ice sheet models. The GIC algorithm also has an improved treatment of volume shrinkage to take into consideration simultaneous changes in glacier area. This allows glaciers to reach a new equilibrium under a climate warming, contrary to older models which use a time-constant sensitivity for mass balance, so that glaciers would melt away for any warming rather than approaching a new equilibrium (e.g. Van de Wal and Wild 2001). The algorithm is forced by applying annual temperature anomalies with respect to the 1961–1990 period. Precipitation changes are not considered in line with conclusions from several studies showing this to be of secondary importance (Braithwaite *et al* 2002).

3. Model initialization for the present-day

The experiments take into account the 2000 AD background evolution of land ice and thermal expansion from an initialization as far back as the last glacial maximum (19 kyr BP) concerning the ice sheets. This is followed by a proper spin-up procedure in several steps from 500 AD onwards in fully coupled LOVECLIM mode forced with historic records of solar irradiance, volcanic activity, greenhouse gas concentrations and sulfate aerosol load between 1500 and 2000 AD. An iterative procedure is applied as the ice sheet models are forced with monthly temperature and precipitation anomalies against a 1970–2000 reference state that is initially unknown, and therefore needs to be defined by a precursor experiment. The outcome of the procedure are GrIS and AIS that are not in steady state, but rather carry the long-term memory of their history with them (Goelzer *et al* 2011, Huybrechts *et al* 2011). Three climate change projections are then carried out for the third millennium with greenhouse gas and sulfate aerosol forcing following the IPCC

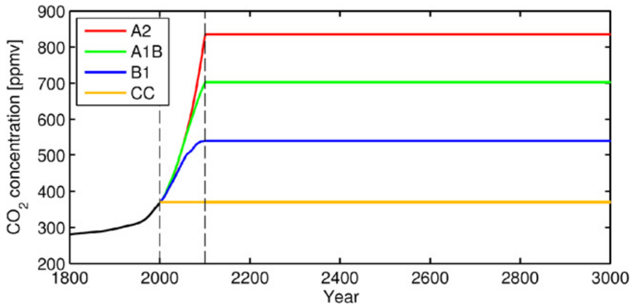

Figure 1. Imposed atmospheric CO_2 concentrations for stabilization scenarios A2, A1B, B1 and CC.

SRES scenarios A2, A1B and B1 until 2100 AD, and held constant thereafter. For an additional experiment (CC), the concentrations of all constituents are held constant at year 2000. Being the dominant forcing agent in these experiments, the time evolution of CO_2 is used to illustrate the different scenarios (figure 1).

Our preferred model version of LOVECLIM is based on a ‘low’ climatic parameter set and ‘medium’ ice sheet parameter sets for AISM and GISM as discussed in Goelzer *et al* (2011). In their model ensemble, it corresponds to the version with the lowest effective climate sensitivity after 1000 years of about 1.6 °C, i.e. at the lower end of the range estimated in IPCC AR4 (Randall *et al* 2007). However, the relatively low climate sensitivity of LOVECLIM combines with a relatively high polar amplification to yield polar temperature changes that are in line with more comprehensive AOGCMs for a given radiative forcing (Ridley *et al* 2005, Gregory and Huybrechts 2006). This indicates that the limited complexity of the climate component of LOVECLIM should not be an obstacle to yield acceptable long-term ice sheet projections. Furthermore, OTE for our preferred model is in the mid-range of results obtained with higher-complexity models for a similar forcing scenario (Meehl *et al* 2007). This is despite the relatively strong oceanic heat uptake in LOVECLIM, especially in model versions with high climate sensitivity (Loutre *et al* 2011). The reference version is also very similar in its temperature response to LOVECLIM 1.0 used in previous studies on climate-ice sheet interactions (Driesschaert *et al* 2007, Swingedouw *et al* 2008, Huybrechts *et al* 2011).

For our preferred set of model parameters, we find a 20th century background sea-level trend of -0.04 mm yr^{-1} for GrIS and $+0.2$ mm yr^{-1} for AIS, in line with earlier results with the same models (Huybrechts *et al* 2004). For the period 1993–2003, these numbers increase to $+0.2$ and $+0.3$ mm yr^{-1} for GrIS and AIS respectively, well within the range given in IPCC AR4 (Lemke *et al* 2007). For the same time period, OTE is 1.4 mm yr^{-1}, and the contribution from GIC is 0.21 mm yr^{-1}. The latter contribution is strongly affected by the prescribed rate of change of 0.19 mm yr^{-1} for the year 1990 in the glacier melt algorithm (Raper and Braithwaite 2006). The total sea-level change 1993–2003 calculated by LOVECLIM of about 2.1 mm yr^{-1} is biased low compared to the observed range of 3.1 ± 0.7 mm yr^{-1} obtained

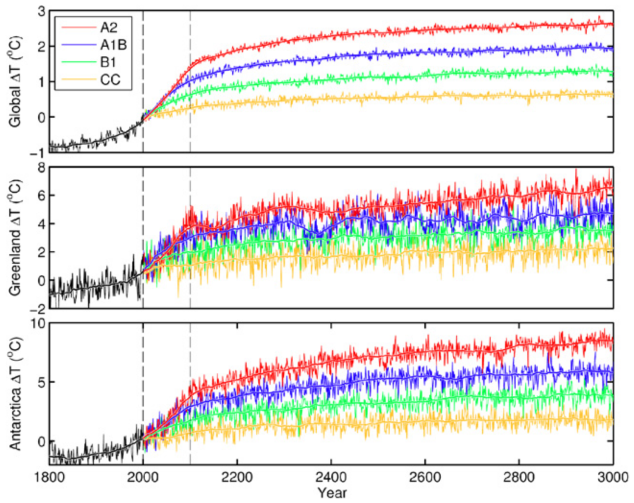

Figure 2. Time evolution of global mean temperature, and annual mean temperature over the Greenland and Antarctic ice sheets for the reference model under constant composition scenarios CC, B1, A1B, and A2. In all cases the full inter-annual variability is shown around a 30 years running mean.

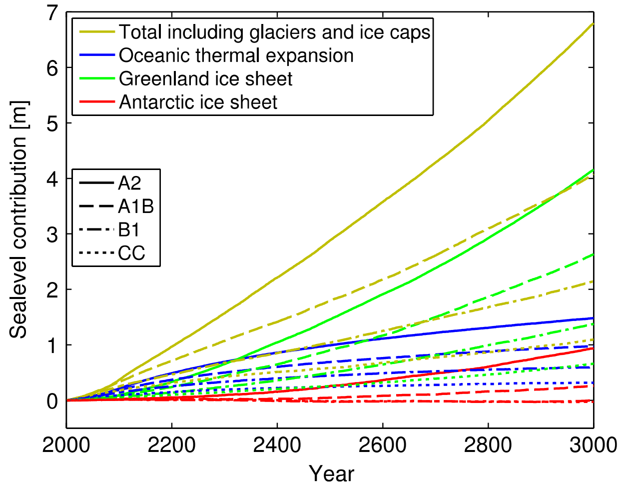

Figure 3. Committed evolution of global sea-level and its components for the constant composition scenarios CC, B1, A1B and A2.

in IPCC AR4 (Bindoff *et al* 2007). This is mainly because of the low GIC contribution, which IPCC AR4 estimates at 0.77 ± 0.22 mm yr^{-1} (Lemke *et al* 2007). The relevance of the somewhat low LOVECLIM estimate however drops rapidly for the millennial projections considered here, which are governed by the long-term climatic response of the model, and in which the glaciers and small ice caps play a minor role.

4. Committed sea-level change after 1000 years

Sea-level contributions are estimated separately for the ice sheets, oceanic thermal expansion, and glaciers and ice caps. Their contributions at 3000 AD are interpreted as 'constant composition sea-level commitments' (Wigley 2005). Results from scenario CC are interpreted as already committed sea-level change, while scenarios B1, A1B and A2 are used to estimate sea-level changes that may be committed by the year 2100.

Polar temperature changes are the main driver for ice sheet changes. These are illustrated for all scenarios in figure 2. Global mean annual temperatures increase by between 0.6 °C (CC) and 2.5 °C (A2) by 3000 AD compared to 2000 AD and are found to be significantly amplified over the GrIS and AIS by a factor of about 2.1 and 3.1, respectively. Such a polar amplification is a robust characteristic of the climate system but is stronger in LOVECLIM than in most other models. It compensates for the low climate sensitivity of LOVECLIM compared to the IPCC AR4 best estimate of 3 °C (Randall *et al* 2007). Gregory and Huybrechts (2006) typically found a polar amplification of around 1.5 for a representative suite of IPCC AOGCMs.

Figure 3 shows the time evolution of ice volume and thermal expansion expressed in sea-level change. For the ice-sheet contribution, results are best interpreted in comparison with the total warming obtained in the $2 \times CO_2$ and $4 \times CO_2$ experiments discussed in Huybrechts *et al* (2011) with a similar version of LOVECLIM. For all scenarios considered here, the Greenland ice sheet is always the largest contributor to sea-level rise after 1000 years. The GrIS decline is almost entirely driven by meltwater runoff (ablation), which dominates the surface mass balance and exceeds snow accumulation for a mean annual warming of about 2.5 °C. This level of warming delimits the ultimate survival of the Greenland ice sheet. For a larger warming, the ice sheet can no longer be sustained, even when the calving flux were to be reduced to zero (Gregory *et al* 2004). For scenario A2, the volume loss by 3000 AD corresponds to 4.2 m s.l.e., or a little more than half of the 7.9 m contained in the entire ice sheet in the model. The associated average warming for scenario A2 in excess of 6 °C implies that the threshold of viability is largely passed and the Greenland ice sheet will completely melt away if the warming were to be sustained for another millennium or so. The committed sea-level rise in scenario A1B is 2.6 m though the corresponding atmospheric warming of 4–5 °C equally implies a committed total loss for a sustained warming over a longer period. In this experiment, the iceberg calving rate drops by 50% already in the first 200 years as the ice sheet retreats from the coast. One of the implications is that marginal erosion of calving fronts from a warmer ocean will quickly become less effective as a mechanism to remove mass. The sea-level commitment from experiment CC is 0.7 m and this figure increases to 1.4 m with additional 21st century greenhouse gas forcing in scenario B1. In these cases, the Greenland ice sheet is expected to shrink for many millennia more but is likely to still reach a new equilibrium.

Surface melting initially plays a negligible role in the mass balance of the Antarctic ice sheet, even for moderate warming conditions. Under all four scenarios considered, surface runoff never exceeds snow accumulation, the latter of which increases with atmospheric warming by values between 12% (B1) and 20% (A2). Basal melting below the ice shelves, on the other hand, steadily increases on an adjustment timescale set by the thermal inertia of the ocean, which is

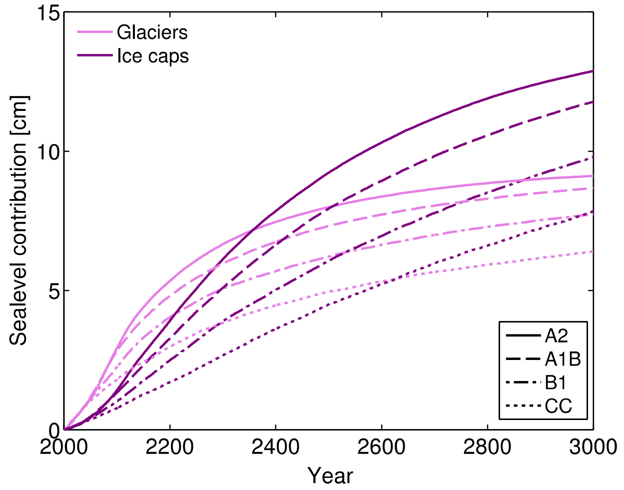

Figure 4. Global sea-level contributions from glaciers and ice caps for the scenarios CC, B1, A1B and A2 for the preferred model version.

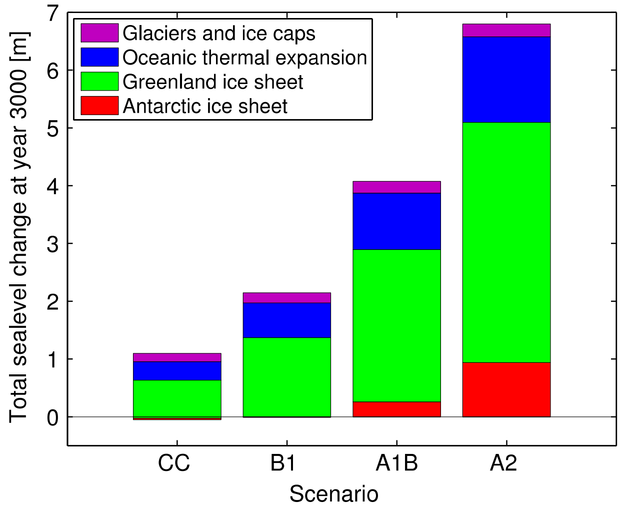

Figure 5. Committed total sea-level contributions from the different components considered in this study by the year 3000 relative to 2000 AD for prolonged SRES scenarios B1, A1B and A2 and constant year 2000 composition CC. The total bar is shifted downwards to represent the negative contribution from the Antarctic ice sheet (B1, CC) and display the total sea-level rise on the left axis correctly.

typically of the order of several hundreds of years. In scenario A1B, increased oceanic heat input from the surrounding ocean leads to almost tripling of the mean melt rate below the ice shelves by the year 3000. This eventually leads to grounding-line retreat and an AIS contribution to sea-level rise of 26 cm. The committed mean annual warming is larger for scenario A2 at around +8 °C and this leads to a committed sea-level rise of 94 cm after 1000 years. For scenarios CC and B1, committed sea-level contributions are slightly negative due to increased snow accumulation and a lack of substantial melting at the ice sheet margin throughout most of the simulation as well as very limited grounding-line retreat from increased melting below the ice shelves (figure 3).

Thermal expansion of the world oceans, initially the largest contributor to sea-level rise, falls behind Greenland's contribution around year 2500 (2300) for scenario B1 (A2). The rate of change is constantly decreasing after 2100 in all scenarios as the climate system relaxes to a steady state under a constant radiative forcing. The sea-level commitment (CC) for OTE is found as 0.32 m at 3000 AD. This figure is only slightly more than the 23 cm reported for emission cessation in 2010 by Gillett *et al* (2011) in their model study that however included uptake of CO_2 by the terrestrial biosphere and the oceans. OTE sea-level commitments are larger for the experiments with concentration stabilization by 2100 and range in LOVECLIM between 0.4 m (B1) and 1.5 m (A2) after 1000 years. These values are very similar to the 0.4–1.9 m range of OTE for emission cessation between 2100 and 2200 in Solomon *et al* (2009), taking into account that the latter results considered a higher peak CO_2 concentration up to 1200 ppmv (compared to 850 ppmv in scenario A2) but included a fall-off of the atmospheric CO_2 concentration from carbon cycle feedbacks.

Mountain glaciers melt away rapidly followed by ice caps at a slower pace (figure 4). In all three SRES scenarios, at least half of the total mountain glacier volume is lost within the first 500 years. For scenario A1B, almost 80% of the combined potential GIC melt volume is lost during the 3rd millennium and half of that within 400 years. Interestingly, the already committed sea-level contribution from GIC combined (scenario CC) is more than half (14 cm) of the total available melt volume of 26 cm. This clearly suggests that a large fraction of the global land ice cover outside of the polar ice sheets may well be irreversibly lost even without a further increase of the atmospheric CO_2 concentration. For the highest scenario considered here (A2) almost all of the glaciers and ice caps have in fact disappeared after 1000 years. The potential GIC sea-level rise is 10.6 cm for glaciers and 14 cm for ice caps (Raper and Braithwaite 2006).

Figure 5 combines the total sea-level commitment by 3000 from all components for all scenarios. The associated total sea-level rises are respectively 2.14 m (B1), 4.08 m (A1B), and 6.80 m (A2). The already committed total sea-level rise is 1.09 m (CC). As demonstrated in figure 3, the rate of sea-level rise is steadily increasing during the third millennium. For scenario A1B the average rate is about 3 mm yr^{-1} over the 21st century but reaches a maximum rate of 5.6 mm yr^{-1} at the end of the simulation, and is still increasing by then. It is also clear that for the millennial timescale under consideration here, the level of greenhouse gas stabilization is far more important than the time of stabilization (year 2000 for CC versus year 2100 for B1, A1B and A2).

5. Sampling uncertainty of the climate model

The range of committed sea-level change by 3000 discussed above arises from uncertainties in anthropogenic forcing but this may not sample the full range of plausible outcomes. Additionally, uncertainty arises from the range of climate sensitivity of the atmospheric and oceanic components of LOVECLIM. In Loutre *et al* (2011) and Goelzer *et al* (2011)

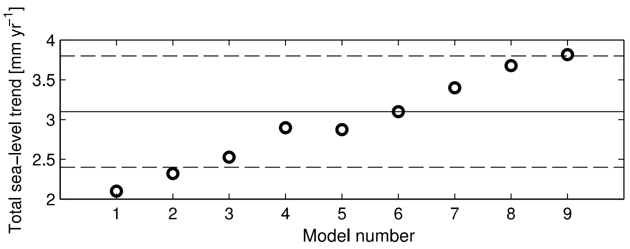

Figure 6. Simulated trends of total sea-level rise for the period 1993–2003 for 9 different model versions (circles) compared to IPCC AR4 sea-level estimates of 3.1 ± 0.8 mm yr^{-1} (horizontal full and dashed lines). The model numbers are ordered according to the climate sensitivity of the individual ensemble member. The preferred model version corresponds to model number 1.

nine different model versions were assessed that differed in their parameter sets used for issues like the radiation scheme, cloud representation, albedo of ocean and sea-ice, and more. The selected models differed mainly in their climate sensitivity (in the range 1.6–4.1 °C after 1000 years) and in their sensitivity of the MOC of the North Atlantic Ocean to water hosing but were shown to reproduce many aspects of 20th century climate change within the observational uncertainty.

Here we find that all nine LOVECLIM versions produce total sea-level trends that are in good agreement with the IPCC AR4 observed range of 3.1 ± 0.8 mm yr^{-1} for the period 1993–2003 (Bindoff *et al* 2007; figure 6). The modelled sea-level trends scale almost linearly with climate sensitivity despite a large climate variability over the considered period. Model number 1 corresponds to our preferred model. While the two models of highest climate sensitivity would be slightly above the IPCC AR4 error margin for total sea-level trend when corrected for our systematic underestimation of the GIC contribution of almost 0.6 mm yr^{-1}, all other models fall within the range of observations when corrected for this bias. This gives some confidence in the ability of LOVECLIM to reproduce current climate and sea-level trends, but is not considered sufficient to qualify a model for realistic long-term projections. Instead we selected our preferred model version above based on its expected (polar) temperature response compared to AOCGMs.

Since the committed sea-level response after 1000 years is governed mainly by the long-term response of the climate model, the argument can however be reversed. Consequently, the temperature response resulting from any of the nine LOVECLIM model versions can be tentatively used as a predictor for the committed sea-level change for each component separately. This is illustrated in figure 7. For all components, sea-level contributions at the year 3000 are related to the total mean annual temperature change (year 2000–3000). The latter is a temperature change index that gives a measure of the integrated temperature anomaly imposed over any of the sea-level components we considered here.

For the ice sheets, the different experiments that combine both the range of scenarios and climate sensitivities align on a relatively narrow curve, indicating that polar temperature changes are the dominant driver for sea-level contributions from either ice sheet. It also shows that either changing the climate sensitivity of the model or changing the forcing scenario results in a similar response. This outcome is largely independent of the exact choice of averaging period. Averaging temperature anomalies over the last 200 or 500 years of the experiment yields a slightly wider spread but overall similar results. The upper range of very high temperature anomalies and high sea-level contributions results from the combination of relatively high climate sensitivity models with the high polar amplification of LOVECLIM. We nevertheless display the full range of experiments and caution to interpret the results as ice sheet response for a given polar warming rather than for a given scenario.

The sea-level contribution from the GrIS at the end of the third millennium for the most extreme temperature changes is limited by the total available ice volume in the model, corresponding to 7.9 m s.l.e. For a local mean annual warming level of between 2 and 6 °C, the committed sea-level rise increases by about 1 m per degree of warming. The committed sea-level contribution from the AIS is below 20 cm or even negative for warmings of less than 5 °C. For higher warmings, its contribution rises non-linearly with temperature. This is a direct consequence of the functional relationship between surface mass balance and temperature change. For warmings below about 5 °C, the surface mass balance in the model is larger than today on account of the limited surface runoff, but the latter increases as a second order function of temperature as both the amount of melting as the duration of the melting season increases in a warmer climate (Huybrechts and Oerlemans 1990). In contrast to the GrIS, no limit of the AIS contribution is reached within the studied range of temperature anomaly and time period. As a consequence of increased ice shelf melting substantial ice loss occurs due to grounding-line retreat for mean annual warmings in excess of +8 °C, in particular for the West Antarctic ice sheet (Huybrechts *et al* 2011).

Committed sea-level contributions by 3000 from OTE scale almost linearly with the committed global warming averaged for the whole 3rd millennium (figure 7, panel (c)) at a slope of about 1 m per degree of global warming. The approximate linearity is to be expected on theoretical grounds. Any (small) deviation probably indicates that the adjustment timescale becomes larger for a larger forcing. For lower warming, the ocean is already closer to equilibrium with the given forcing and the slope is flattening. This is confirmed by a relatively lower ocean heat uptake in these experiments (not shown). The slope of about 1 m per degree of global warming is however higher than the ones found in Solomon *et al* (2009) and Gillett *et al* (2011). That is probably a consequence of both the lack of an interactive carbon cycle and the relatively strong oceanic heat uptake in our model, especially for versions with high climate sensitivity (Loutre *et al* 2011).

Sea-level contributions from glaciers (figure 7, panel (d), below) and ice caps (panel (d), top) are best displayed against average global warming over glaciated areas as the latter are found everywhere on the globe (Raper and Braithwaite

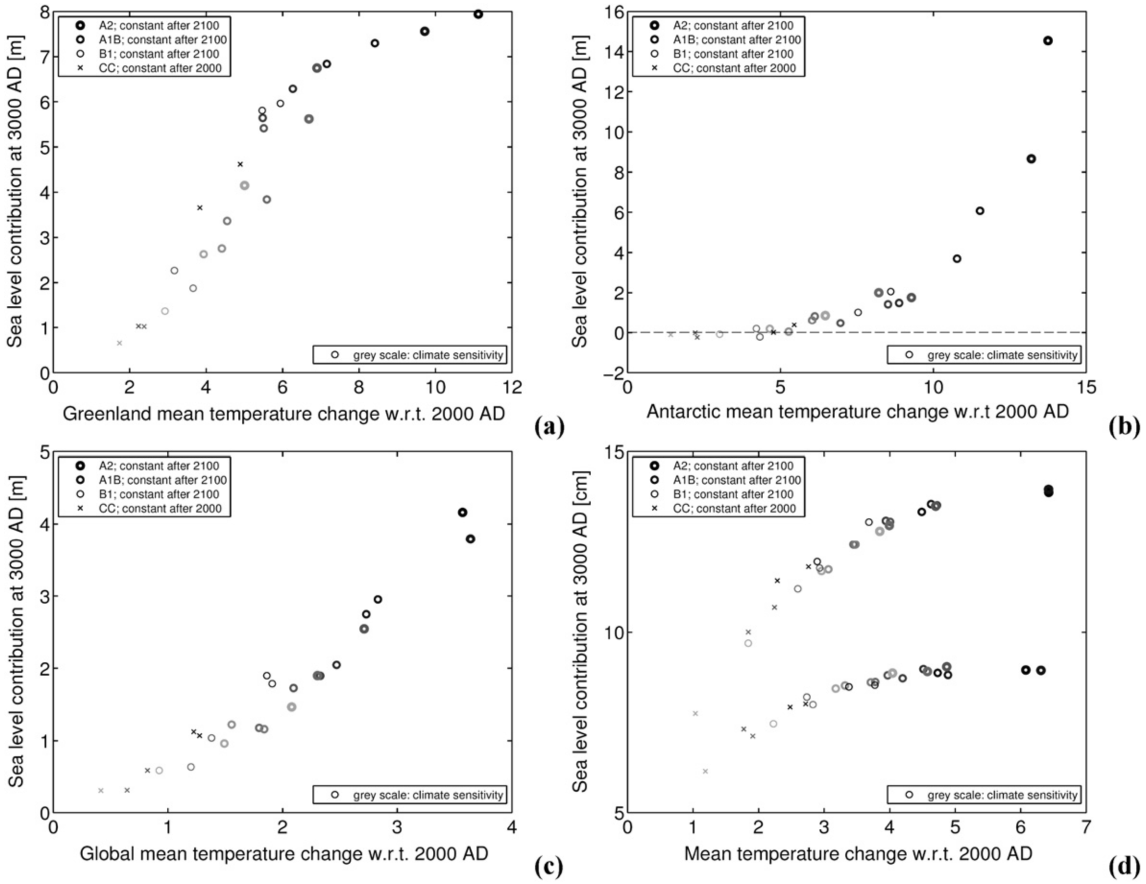

Figure 7. Committed sea-level contributions after 1000 years as a function of mean annual temperature changes over the third millennium (2000–3000 AD) for the Greenland (a) and Antarctic (b) ice sheets, oceanic thermal expansion (c) and glaciers and ice caps (d). Sea-level contributions are shown as a function of temperature change over the present-day grounded ice area ((a), (b)), over the entire globe (c) and over glaciated areas only (d). The figures combine sea-level contributions as a function of the scenario (symbols) and as a function of the climate sensitivity of an individual ensemble member (light to dark grey from low to high sensitivity). In (d) the upper symbols are for ice caps and the lower row of symbols are for mountain glaciers.

2005). Ice caps exhibit both a longer response timescale and a larger integral sea-level contribution than glaciers (Raper and Braithwaite 2006). By the end of the third millennium, however, both glaciers and ice caps have almost completely vanished for all but the lowest temperature forcing. Their committed sea-level contribution therefore converges as a function of increasing temperature forcing almost to the maximum available melt volume.

6. Conclusion

Anthropogenic carbon dioxide is causing global warming and this will lead to irreversible sea-level rise on human timescales even when emissions were to reduce drastically to reach greenhouse gas stabilization in the atmosphere. Our model results indicate that a sea-level rise of at least 1.1 m by the year 3000 is already committed by the compounded effect of greenhouse gas buildup by the year 2000. Several metres more are likely unavoidable if measures to curb emissions drastically are not taken during the next few decades and greenhouse gas concentrations continue to rise to the levels implied by the SRES scenarios used in this study. This outcome is directly linked to the longevity of anthropogenic CO_2 and the long response timescales inherent in the ocean and the ice sheets. On the millennial timescale, the results stress the dominance of melting of the Greenland ice sheet to future sea-level rise and the relative insensitivity of the Antarctic ice sheet as long as the mean annual warming over the ice sheet does not exceed 5–6 °C. Glaciers and ice caps are found to largely disappear within centuries irrespective of further increases of radiative forcing but the total implied sea-level rise is limited to less than 30 cm in the algorithm used here. Oceanic thermal expansion is only dominant for present-day conditions and the next few centuries but is never the main contributor to committed sea-level changes by the year 3000.

The results presented here are subject to uncertainties both in climate sensitivity and aspects of the ice models. Our preferred LOVECLIM model version had a low climate sensitivity and considered medium physical parameters for the ice sheets. However, strong relations were found between selected temperature indices and committed sea-level changes for all sea-level components across all model versions and all forcing scenarios. This implies that the established relations can be used to estimate first-order sea-level commitments given estimates of spatially resolved global temperature change. These could come from other Earth system models that do not include land ice components. Uncertainties of the ice sheet projections also arise from poorly constrained physics in prescribing ice-sheet mass balance, basal sliding conditions, and the effects of oceanic erosion of ice shelves and calving fronts. Such limitations are thought to be less crucial for the Greenland ice sheet than for the Antarctic ice sheet, but were not investigated further with the current model setup.

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