Avoiding Dangerous Climate Change

Editor in Chief

Hans Joachim Schellnhuber

Co-editors

Wolfgang Cramer, Nebojsa Nakicenovic, Tom Wigley, Gary Yohe

CAMBRIDGE
UNIVERSITY PRESS

CAMBRIDGE UNIVERSITY PRESS
Cambridge, New York, Melbourne, Madrid, Cape Town, Singapore, São Paulo

CAMBRIDGE UNIVERSITY PRESS
The Edinburgh Building, Cambridge CB2 2RU, UK

Published in the United States of America by Cambridge University Press, New York

www.cambridge.org
Information on this title: www.cambridge.org/9780521864718

First published 2006

Editorial project management by Frameworks
www.frameworks-city.ltd.uk

Typeset by Charon Tec Ltd, Chennai, India
www.charontec.com

Printed in the United Kingdom at the University Press, Cambridge

A catalogue record for this book is available from the British Library

ISBN: 13 978-0-521-86471-8 hardback
ISBN: 10 0-521-86471-2 hardback

CONTENTS

CHAPTER 4

The Role of Sea-Level Rise and the Greenland Ice Sheet in Dangerous Climate Change: Implications for the Stabilisation of Climate

Jason A. Lowe[1], Jonathan M. Gregory[1,2], Jeff Ridley[1], Philippe Huybrechts[3,4],
Robert J. Nicholls[5] and Matthew Collins[1]

[1] *Hadley Centre for Climate Prediction and Research, Met Office, Exeter, UK*
[2] *Centre for Global atmospheric Modelling, Department of Meteorology, University of Reading, UK*
[3] *Alfred-Wegener-Institut für Polar- und Meeresforschung, Bremerhaven, Germany*
[4] *Department Geografie, Vrije Unversiteit Brussel, Belgium*
[5] *School of Civil Engineering and the Environment, University of Southampton, UK*

ABSTRACT: Sea level rise is an important aspect of future climate change because, without upgraded coastal defences, it is likely to lead to significant impacts. Here we report on two aspects of sea-level rise that have implications for the avoidance of dangerous climate change and stabilisation of climate.

If the Greenland ice sheet were to melt it would raise global sea levels by around 7 m. We discuss the likelihood of such an event occurring in the coming centuries. The results suggest that complete or partial deglaciation of Greenland may be triggered for even quite modest stabilisation targets. We also examine the time scales associated with sea-level rise and demonstrate that long after atmospheric greenhouse gas concentrations or global temperature have been stabilised coastal impacts may still be increasing.

4.1 Introduction

Sea level is reported to have risen during the 20th century by between 1 and 2 mm per year and model predictions suggest the rise in global-mean sea level during the 21st century is likely to be in the range of 9–88 cm (Church et al., 2001). It is also well known that there has been considerable growth in coastal populations and the value of assets within the coastal zone during the 20th century, and this may continue in the future. Consequently, there is a concern that future increases in sea level will lead to sizeable coastal impacts (Watson et al., 2001). The issue of sea-level rise in dangerous climate change has also recently been discussed by Oppenheimer and Alley (2004) and Hansen (2005).

The main causes of increased global average sea level during the 21st century are likely to be thermal expansion of the ocean, melting of small glaciers, and the melting of the Greenland and Antarctic ice sheets (Church et al., 2001). Thermal expansion and the melting of small glaciers are expected to dominate, with Greenland contributing a small but positive sea-level rise, which may be partly offset by a small and negative contribution from Antarctica. This negative contribution results from an increase in precipitation over Antarctica, which is assumed to more than offset small increases in melting during the 21st century. With further warming the Antarctic ice sheet is likely to provide a positive sea-level rise contribution, especially if the West Antarctic Ice Sheet (WAIS) becomes unstable. Beyond the 21st century the changes in the ice sheets and thermal expansion are expected to be make the largest contributions to increased sea level.

In this work we concentrate on two issues associated with sea-level rise. First, how likely is it that the Greenland ice sheet will undergo complete or significant partial deglaciation during the coming centuries, thus providing a large additional sea-level rise? Second, what are the time scales of sea-level rise, especially those associated with thermal expansion and Greenland deglaciation, and what are the consequences of the time scales for mankind?

4.2 Models and Climate Change Scenarios

Results are presented from a range of physical models, including: simple climate models; complex climate models with detailed representation of the atmosphere, ocean and land surface; and a high-resolution model of the Greenland ice sheet.

A small number of long simulations have been performed with the coupled ocean-atmosphere general circulation climate model, HadCM3. This is a non flux-adjusted coupled model with an atmospheric resolution of $2.5° \times 3.75°$ and 19 levels in the atmosphere. The ocean is a 20 level rigid-lid model with a horizontal resolution of $1.25° \times 1.25°$ and 20 levels. More details of the model and its parameterisations are given by Pope et al. (2000) and Gordon et al. (2000).

Recently, we used this model to simulate around 1000 years for an experiment in which atmospheric carbon dioxide concentration was increased from a pre-industrial level of approximately 285 ppm at 2% compound per annum, then stabilised after 70 years at four times the pre-industrial value for the remainder of the simulation. An increase in atmospheric carbon dioxide to four times pre-industrial atmospheric carbon dioxide corresponds to a radiative forcing of around 7.5 Wm^{-2}, which is comparable to the 6.7 Wm^{-2} increase in forcing between years 2000 and 2100 for the SRES A2 scenario and 7.8 Wm^{-2} for SRES A1FI (IPCC 2001, Appendix 2). In a second simulation, HadCM3 was coupled to a 20 km resolution dynamic ice sheet model (Ridley et al., 2005; Huybrechts et al., 1991) and used to simulate more than 3000 years of ice sheet evolution. Importantly, the coupling method allowed changes in climate to influence the evolution of the ice sheet and changes in the ice sheet to feedback on the climate, affecting its subsequent evolution.

We have also made a number of additional simulations using a large number of slightly different but plausible versions of HadCM3. These models used a simplified slab ocean, which responds to radiative forcing changes much faster than the ocean in the fully coupled model, allowing estimates of equilibrium response to be made relatively quickly. For this work we used an ensemble of 129 simulations in which atmospheric carbon dioxide levels were first prescribed at pre-industrial levels ($1\times CO_2$) and then doubled ($2\times CO_2$). In both the $1\times CO_2$ and $2\times CO_2$ phases the simulations were run until they first reached an equilibrium and then for a further 20 years.

Like other models, the Hadley Centre climate model contains a number of parameters that may be modified within a sensible range. In this work, there is one ensemble member in which model parameters and parameterisation schemes take their standard values (Pope et al., 2000), with the exception of the use of a prognostic sulphur cycle model component. In the remaining 128 ensemble members, perturbations were made simultaneously to these standard values for a range of important model parameters. The choice of parameters perturbed and the effects of perturbations on global mean equilibrium climate sensitivity are described in Murphy et al. (2004) and Stainforth et al. (2005).

The precise algorithm for generating the perturbations is complex but, briefly, the ensemble was designed on the basis of linear statistical modelling to produce a range of different magnitude climate sensitivities while maximising the chance of high-fidelity model base climates and exploring as much of the model parameter space as possible. More details are given in Webb et al. (2005), together with an assessment of cloud feedbacks in the ensemble. A method for producing probability density functions of future climate change predictions is to first run the ensemble of simulations to generate a frequency distribution and second to give a relative weight to each ensemble member based on some assessment of its 'skill' in simulating the forecast variable of interest. The details of the correct way of doing this are still subject to considerable debate and require much further work, particularly when addressing the question of regional climate change as we do here. We therefore limit ourselves to the production of frequency distributions. The consequences of this for the use of these results in a formal risk assessment are discussed in Section 3. A further limitation is that our model ensemble is based on a single climate model and we have not attempted to account for results from other climate models. However, we do note that the range of climate sensitivities produced by the 129 member ensemble are not inconsistent with those published in other studies (e.g. Frame et al., 2005) which tend to use simple models and a range of different observational constraints.

Finally, we have used simple model formulations in which both temperature change and sea-level rise are represented using Green's functions. The Green's functions are taken as the sum of two exponential modes derived from the 1000 year HadCM3 stabilisation experiment without an ice sheet. Predictions were made with the simple model by convolving either the temperature Green's function or sea-level rise Green's function with an estimate of the radiative forcing. These simple models have only been used here to extend more complex Hadley Centre model results further into the future or to scale to alternative emissions scenarios.

4.3 Likelihood of a Deglaciation of Greenland

If the Greenland ice sheet were to melt completely, it would raise global average sea level by around 7 m (Church et al., 2001). Without upgraded sea defences this would inundate many cities around the world. There are also concerns that the fresh water from Greenland could help trigger a slowdown or collapse of the ocean thermohaline circulation[1] (Fichefet et al., 2003). This could lead to a significant cooling over much of the northern hemisphere (Vellinga and Wood, 2002).

The Greenland ice sheet can only persist if the loss of ice by ablation and iceberg discharge is balanced by accumulation. Under present day conditions the two loss terms are each roughly half the accumulation. If the accumulation were greater than the sum of the loss terms then the ice sheet would grow. However, in a warmer climate it is expected that the increase in ablation will outweigh the increase of accumulation. Under these circumstances, the ice sheet will shrink. For a small warming, the ice sheet could still evolve towards a new equilibrium by reducing its rate of iceberg calving and/or obtaining a different geometry that reduces ablation sufficiently to counterbalance the initial increase of the surface melting. However, as reported in the IPCC's third assessment report (Church et al., 2001), based on Huybrechts et al. (1991; see also Oerlemans, 1991; Van de

[1] The ocean thermohaline circulation plays a role in the transport of large amounts of heat from the tropics to high latitudes.

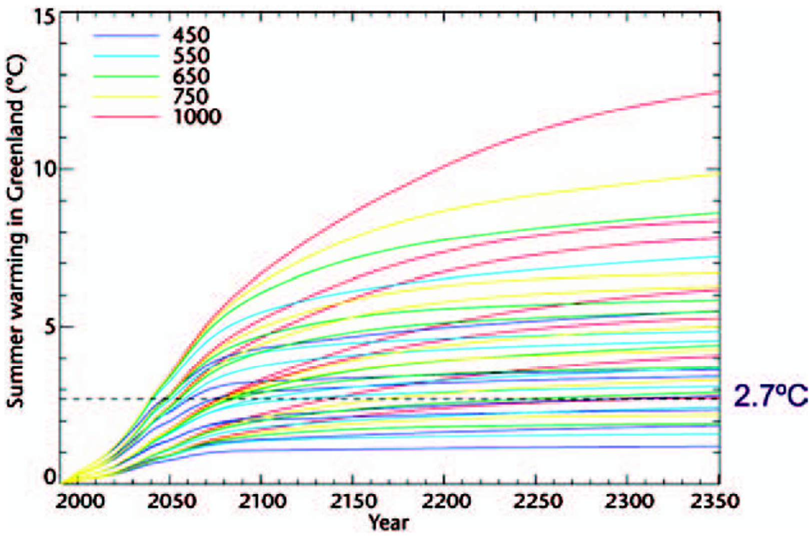


Figure 4.1 Predicted warming for various CO_2 stabilisation levels (purple, 450 ppm; light blue, 550 ppm; green, 650 ppm; yellow, 750 ppm; red, 1,000 ppm). Scenarios involving higher carbon dioxide concentrations stabilize later. The threshold for deglaciation is shown as a dotted line.

Wal and Oerlemans, 1994), for a mean temperature rise of 2.7°C the ablation is predicted to increase beyond the accumulation. Since the ice sheet can not have a negative discharge, this represents the temperature above which the ice sheet can no longer be sustained and will retreat in-land, even if the calving rate were to be reduced to zero.

Alternative thresholds could have been defined, such as the temperature rise leading to a particular loss of Greenland ice by a particular time. Huybrechts and De Wolde (1999) showed that for a local Greenland temperature rise of 3°C the ice sheet would lose mass equivalent to around 1 m of global mean sea-level rise over 1000 years and that the rate of sea-level rise at the end of the 1000-year simulation remained sizeable. In their 5.5°C warming scenario the sea-level rise contribution from Greenland over 1000 years was around 3 m. Thus, we believe that above the chosen temperature threshold a significant Greenland ice loss will occur, although we acknowledge that for warming that is close to the threshold the warming may either not lead to complete deglaciation or that a complete deglaciation may take much longer than a millennium. In Ridley et al., (2005) and Section 4 of this article the ice loss for a high forcing scenario is reported.

Gregory et al. (2004) used the simple MAGICC climate model (Wigley and Raper, 2001), with a range of climate sensitivity and heat uptake parameters to look at the warming over Greenland for a range of greenhouse gas emission scenarios that lead to stabilisation of atmospheric carbon dioxide at levels between 450 ppm and 1000 ppm. The emissions of other greenhouse gas species followed the SRES A1B scenario up to 2100 and were then stabilised. The climate model parameters and the relationship between global mean warming and local warming over Greenland were estimated from the more complex models used in the IPCC third assessment (Church et al., 2001). When annual mean warming was considered, all but one of the model simulations led to a warming above the 2.7°C threshold by approximately 2200. When uncertainty in the threshold and only summer seasonal warming were considered, 69% of the model versions led to the threshold being exceeded before 2350 (Figure 4.1). This use of summer only warming is more appropriate because little melting occurs during the cold winter months.

We have recently attempted to re-examine this issue using the 'perturbed parameter ensemble' of Hadley Centre complex climate models (described in Section 2). For each ensemble member the carbon dioxide stabilisation level that would lead to a Greenland temperature rise equal to the threshold for deglaciation is estimated, assuming a logarithmic relationship between stabilisation carbon dioxide concentration and equilibrium temperature increases. We also make the assumption that the ratio of the summer warming over Greenland to global mean warming and the climate sensitivity will remain constant for a given model over a range of climate forcing and temperature rise.

The orange curve in Figure 4.2 shows a smoothed frequency distribution of the stabilisation carbon dioxide levels that lead to a local Greenland warming of 2.7°C and, thus, a complete or partial Greenland deglaciation being triggered. The red and green curves are the carbon dioxide stabilisation levels that would lead to warmings of 2.2°C and 3.2°C respectively, which represents uncertainty in the value of the deglaciation threshold. The vertical bars show the raw data to which the orange curve was fitted. The results suggest that even if carbon dioxide levels are stabilised below 442 ppm to 465 ppm then 5%

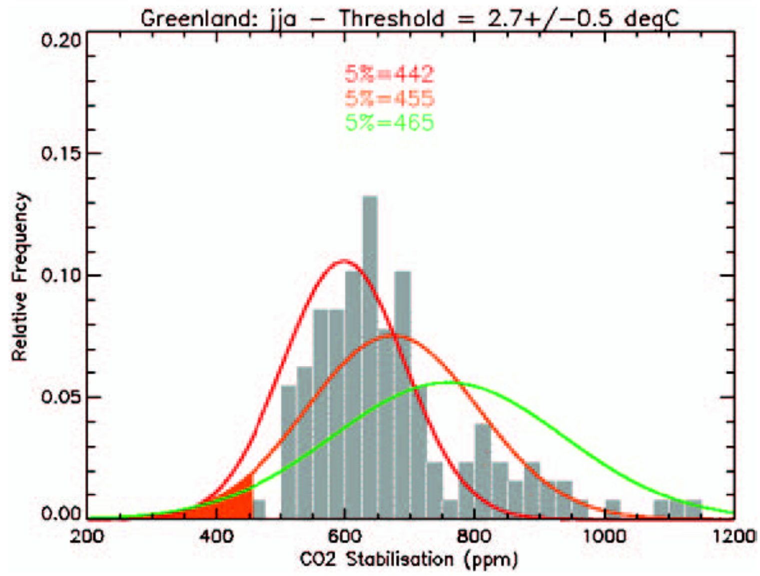


Figure 4.2 Predicted CO_2 stabilisation levels that lead to the local Greenland warming exceeding the threshold for deglaciation, and ± 0.5°C of this amount. The raw results are shown as bars for the central threshold case and the curves are a fit to the raw results.

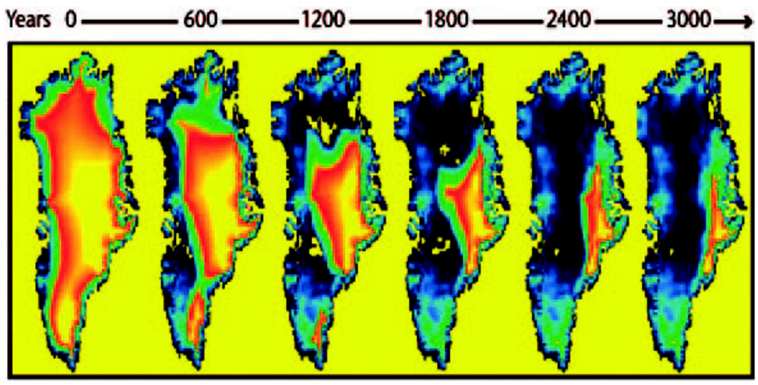


Figure 4.3 Predicted change in the ice sheet volume following a quadrupling of atmospheric CO_2. Red and yellow indicate thick ice while green and blue indicate thin (or no) ice.

of our plausible model simulations will still lead to a complete or partial deglaciation. A stabilisation level of 675 ppm would lead to 50% of our model versions exceeding 2.7°C. At this level, however, the uncertainty in the value of the threshold becomes more important and, when this is taken into account, the carbon dioxide concentration level that leads to 50% of the model version reaching the deglaciation threshold varies between 600 ppm to 750 ppm.

It is important to emphasize that because the ‘perturbed parameter ensemble’ technique is still in its infancy and we have not attempted to apply a weighting to the frequency distribution of carbon dioxide stabilisation levels, so this result can not be taken as a formal probability density function or definitive estimate of the risk of collapse. Rather, we have used the ensemble to illustrate the method whereby such a risk may be estimated. To that end, our results are likely to be a credible first attempt at linking the collapse of the Greenland ice sheet to a particular stabilisation level using a perturbed parameter approach with complex climate models.

4.4 Timescales of sea level response

Having established that even for quite modest carbon dioxide stabilisation levels the Greenland ice sheet might become deglaciated, we now discuss the time scales over which this might occur. For a pessimistic, but plausible, scenario in which atmospheric carbon dioxide concentrations were stabilised at four times pre-industrial levels (Section 2) a coupled climate model and ice sheet model simulation predicts that the ice sheet would almost totally disappear over a period of 3,000 years, with more than half of the ice volume being lost during the first millennium (Figure 4.3). The peak rate of simulated sea-level rise was around 5 mm/year and occurred early in the simulation. These results are discussed more fully by Ridley

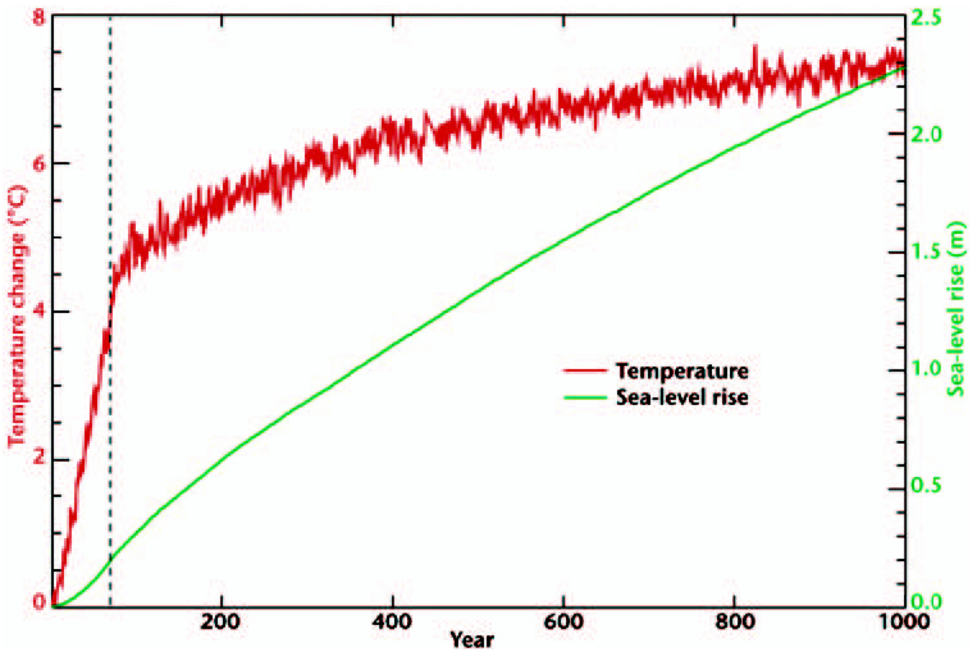


Figure 4.4 Simulated temperature rise and thermal expansion for a 4×CO_2 experiment.

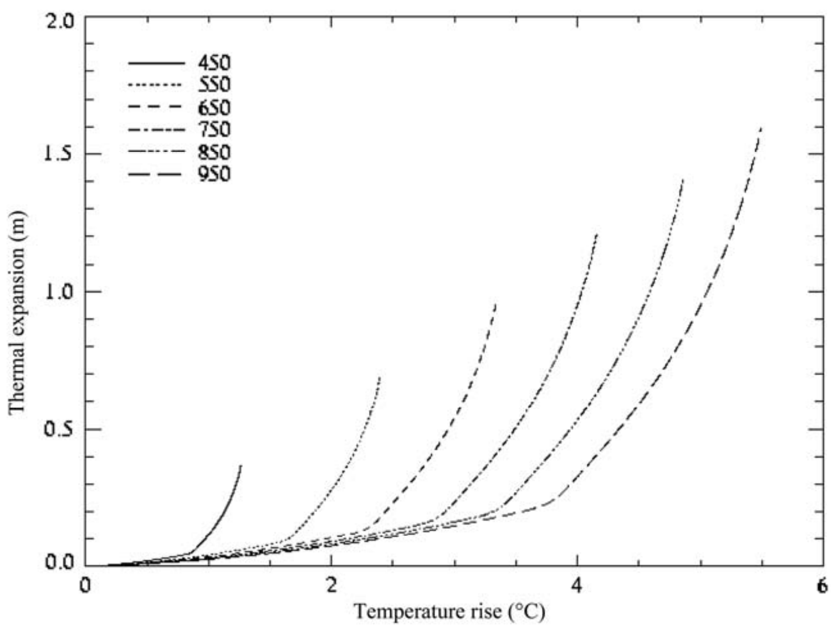
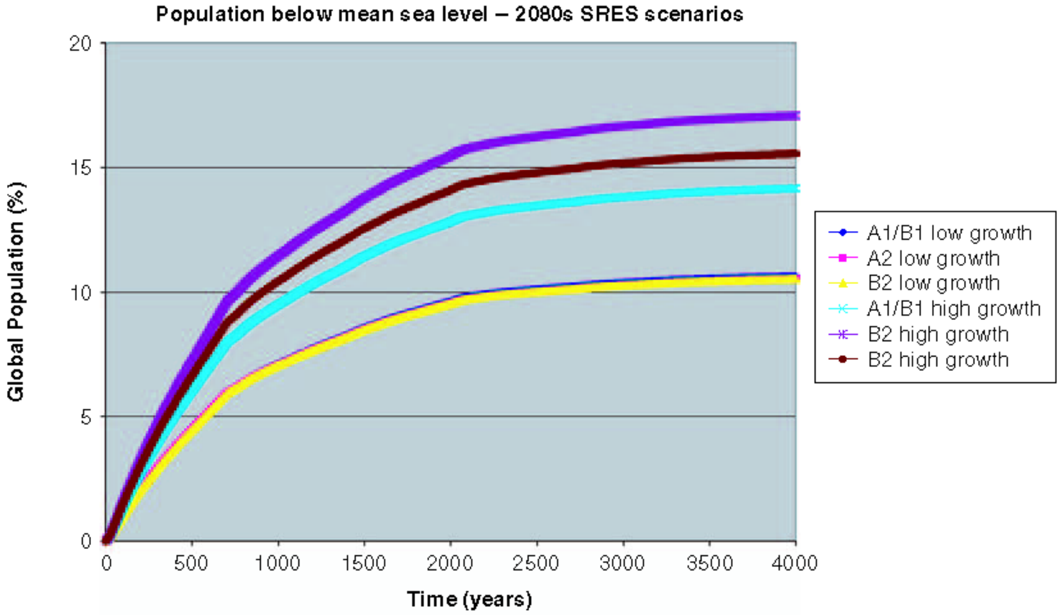


Figure 4.5 Simulated temperature rise and thermal expansion for a range of stabilisation levels. The stabilisation of atmospheric carbon dioxide takes place 70 years into the experiment following a linear increase.

et al. (2005) who also note that in the Hadley Centre climate model, the freshwater provided by the melting of Greenland ice had a small but noticeable effect on the model's ocean circulation, temporarily reducing the thermohaline circulation by a few per cent. However, this was not enough to lead to widespread northern hemisphere cooling.

A further issue associated with the loss of ice from Greenland is that of reversibility. If the climate forcing were returned to pre industrial levels once the ice sheet had become totally or partially ablated could the ice sheet eventually reform? If not, when would the point of no return be reached? The studies of Lunt et al. (2004) and Toniazzo et al. (2004) offer conflicting evidence on whether a fully-ablated ice sheet could reform, and this is an active area of current research.

In the parallel HadCM3 experiment without an ice sheet the thermal expansion was estimated and also found to make a considerable sea-level rise contribution over millennial time scales (Figure 4.4). The timescale associated



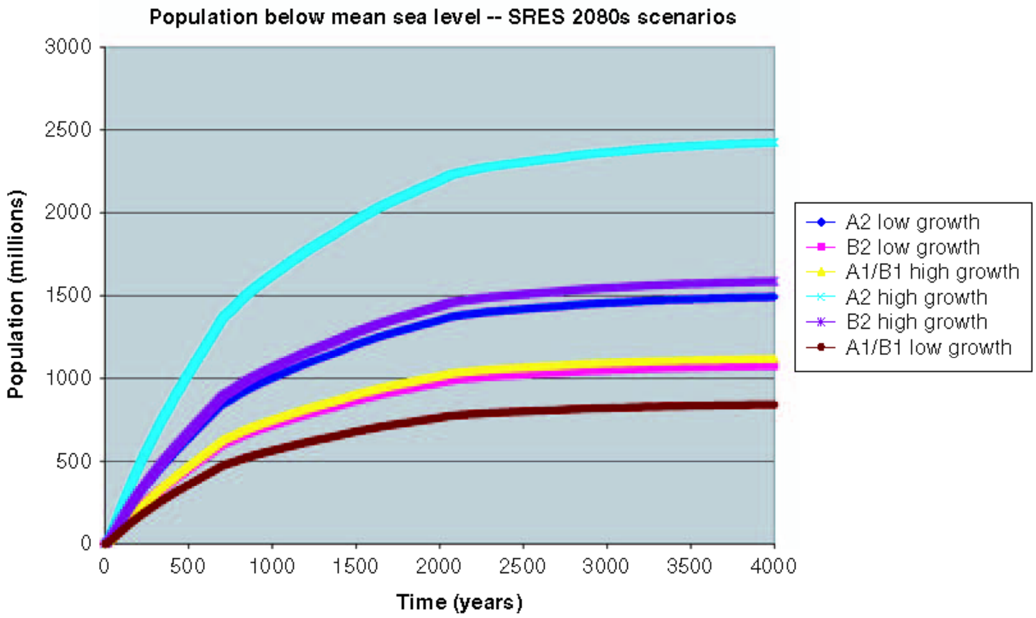


Figure 4.6 (a) Exposed population and (b) percentage of world population exposed to Greenland deglaciation and the thermal expansion from a stabilisation level of four time pre-industrial values.

with the thermal expansion component of sea-level rise depends strongly on the rate at which heat can be transported from near the surface into the deep ocean. The thermal expansion response time in the Hadley Centre coupled climate model was found to be greater than 1000 years, which is much longer than the time needed to stabilise temperature (the global average surface temperature rise for the same experiment is also shown in Figure 4.4). Using the simple Green's function model formulations for thermal expansion and temperature rise, tuned to the HadCM3 results, we have constructed a set of curves showing the time dependent relationship between the two quantities for a range of different carbon dioxide stabilisation levels. These curves were generated for scenarios in which the carbon dioxide was increased linearly over 100 years then fixed at the stabilisation levels.

Figure 4.5 shows that during the period of rapidly-increasing carbon dioxide concentration, the sea-level rise and temperature both increase and there is an approximately linear relationship between them. However, once the carbon dioxide concentration has stabilised, the differing time scales affecting surface temperature and sea-level rise become important and the gradient of the curves increases significantly.

Taken together, the Greenland deglaciation and the thermal expansion results show that sea level is likely to continue rising long after stabilisation of atmospheric carbon dioxide, agreeing with earlier studies, such as Wigley (1995). Changes in the WAIS are also likely to provide an important contribution to future multi-century increases in sea level. However, we can not yet comment with any degree of confidence on the time scales of Antarctic ice sheet collapse. A review of expert opinions (Vaughan and Spouge, 2002) suggested this is not thought likely to occur in the next 100 years, although recent work (Rapley, this volume) suggests the Antarctic ice sheet may make a sizeable contribution to sea-level rise earlier than previously thought.

4.5 Consequences of these Results for Mankind

A detailed assessment of impacts is beyond the scope of this paper. However, it is instructive to add together the Greenland and thermal expansion sea-level rise estimates and consider the potential exposure of people to this rise. The thermal expansion estimate for the first 1000 years is from HadCM3 but this is extended using the simple Green's function climate model formulation. The exposure is based on population estimates for the 2080s, when they are expected to have increased significantly compared to the present situation. The base data comes from the CIESIN PLACE database (http://:www.ceisin.org), and this is transformed using the SRES scenarios, including different growth rates for coastal areas (see Nicholls, 2004).

Figure 4.6 shows the population that is exposed based on absolute numbers and as a proportion of the global population estimates in the 2080s. While this is translating changes over 4000 years, the potential scale of impacts is evident. Within 500 years, the exposed population could be in the range of 300–1000 million people, rising to 800 to 2400 million people at the end of the simulation. This is 10–17% of the world's population, and represents the number of people who would need to be protected or relocated. Nicholls and Lowe (this volume) have extended the calculation to include a contribution from the WAIS but acknowledge that this term is likely to be even more uncertain than the contribution from Greenland.

4.6 Conclusions

Simulations of the Greenland ice sheet and ocean thermal expansion have highlighted several issues that are relevant to the stabilisation of climate at a level that would avoid dangerous changes. In particular:

- Complete or partial deglaciation of Greenland may be triggered for even quite modest stabilisation targets.
- Sea level is likely to continue rising for more than 1000 years after greenhouse gas concentrations have been stabilised, so that with even a sizeable mitigation effort adaptation is also likely to be needed.

We are currently addressing the question of whether the Greenland deglaciation is irreversible or whether, if greenhouse gas concentrations were reduced, the ice sheet could be regrown. If it can recover, we also need to establish the greenhouse gas levels that would permit this to occur. Finally, we note that there is a large uncertainty on sea-level rise predictions, especially those made for times beyond the 21st century.

REFERENCES

Church, J.A., Gregory, J.M., Huybrechts, P., Kuhn, M., Lambeck, K., Nhuan, M.T., Qin, D. and Woodworth, P.L. Changes in Sea Level. In: Houghton, J.T., Ding, Y., Griggs, D.J., Noguer, M., van der Linden, P.J. and Xiaosu, D. (eds.) Climate Change 2001. The Scientific Basis. Cambridge University Press, Cambridge, 639–693, 2001.

Fichefet, T., Poncin, C., Goosse, H., Huybrechts, P., Janssens, I., Le Treut, H. Implications of changes in freshwater flux from the Greenland ice sheet for the climate of the 21st century. Geophysical Research Letters, 30(17), 1911, doi: 10.1029/2003GL017826, 2003.

Frame, D.J., Booth, B.B.B., Kettleborough, J.A., Stainforth, D.A., Gregory, J.M., Collins, M. and Allen, M.R. Constraining climate forecasts. Geophysical Research Letter, 32(9), L09702, doi: 10.1029/2004GL022241, 2005.

Gordon, C., Cooper, C., Senior, C.A., Banks, H., Gregory, J.M., Johns, T.C., Mitchell, J.F.B. and Wood, R.A. The simulation of SST, sea ice extents and ocean heat transports in a version of the Hadley Centre coupled model without flux adjustments. Climate Dynamics, 16, 147–168, 2000.

Gregory, J.M., Huybrechts, P. and Raper, S.C.B. Threatened loss of the Greenland ice-sheet. Nature 428, 616, 2004.

Hansen, J.E.A. slippery slope: How much global warming constitutes "dangerous anthropogenic interference"? Climatic Change, 68, 269–279, 2005.

Huybrechts, P., Letreguilly, A. and Reeh, N. The Greenland Ice Sheet and greenhouse warming. Paleogeography, Paleoclimatology, Paleoecology (Global and Planetary Change Section), 89, 399–412, 1991.

Huybrechts, P. and de Wolde, J. The dynamic response of the Greenland and Antarctic ice sheets to multiple-century climatic warming. Journal of Climate, 12, 2169–2188, 1999.

IPCC. 'Climate Change 2001: The Scientific Basis. Contribution of Working Group I to the Third Assessment Report of the Intergovernmental Panel on Climate Change', Houghton, J.T., Ding, Y., Griggs, D.J., van der Linden, P.J., Dai, X., Maskell, K. and Johnston, C.A. (eds.), Cambridge University Press, Cambridge, UK and New York, NY, USA, pp. 881, 2001.

Murphy, J.M., Sexton, D.M.H., Barnett, D.N., Jones, G.S., Webb, M.J., Collins, M. and Stainforth, D.A. Quantification of modelling uncertainties in a large ensemble of climate change simulations. Nature, 430, 768–772, 2004.

Nicholls, R.J. Coastal flooding and wetland loss in the 21st Century: Changes under the SRES climate and socio-economic scenarios. Global Environmental Change, 14, 69–86, 2004.

Nicholls, R.J. and Lowe, J.A. Climate Stabilisation and Impacts of Sea-Level Rise. This volume.

Oerlemans, J. The mass balance of the Greenland ice sheet: sensitivity to climate change as revealed by energy balance modelling. Holocene, 1, 40–49, 1991.

Oppenheimer, M. and Alley, R.B. Ice sheets, global warming, and Article 2 of the UNFCCC. Climatic Change, 68, 257–267, 2005.

Pope, V.D., Gallani, M.L., Rowntree, P.R. and Stratton, R.A. The impact of new physical parameterisations in the Hadley Centre climate model – HadAM3, Climate Dynamics, 16, 123–146, 2000.

Rapley, C. The Antarctic ice sheet and sea level rise. This volume.

Ridley, J.K., Huybrechts, P., Gregory, J.M. and Lowe, J.A. Elimination of the Greenland ice sheet in a high CO_2 climate. Journal of Climate, 18, 3409–3427, 2005.

Stainforth, D.A., Aina, T., Christensen, C., et al. Uncertainty in predictions of the climate response to rising levels of greenhouse gases. Nature, 433, 403–406, 2005.

Van de Wal, R.S.W. and Oerlemans, J. An energy balance model for the Greenland ice sheet. Glob. Planetary Change, 9, 115–131, 1994.

Vaughan D.G. and Spouge, J.R. Risk estimation of collapse of the West Antarctic Ice Sheet. Climatic Change 52, 65–91, 2002.

Vellinga M. and Wood R.A. Global climatic impacts of a collapse of the Atlantic thermohaline circulation. Climatic Change 54, 251–267, 2002.

Watson, R.T. and the Core Writing Team (eds.). Climate Change 2001: Synthesis Report, Cambridge University Press, Cambridge, pp. 396, 2001.

Webb M.J., Senior, C.A., Williams, K.D., et al. On the contribution of local feedback mechanisms to the range of climate sensitivity in two GCM ensembles. Submitted to Climate Dynamics.

Wigley, T.M.L. and Raper, S.C.B. Interpretation of high projections for global-mean warming. Science 293, 451–454, 2001.

Wigley, T.M.L. Global-mean temperature and sea level consequences of greenhouse gas concentration stabilization. Geophysical Research Letters, 22, 45–48, 1995.