PHILOSOPHICAL TRANSACTIONS OF THE ROYAL SOCIETY A MATHEMATICAL, PHYSICAL & ENGINEERING SCIENCES

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Phil. Trans. R. Soc. A 2006 **364**, 2117-2133
doi: 10.1098/rsta.2006.1816

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Phil. Trans. R. Soc. A (2006) **364**, 2117–2133
doi:10.1098/rsta.2006.1816
Published online 23 June 2006

Extreme events due to human-induced climate change

By John F. B. Mitchell*, Jason Lowe, Richard A. Wood and Michael Vellinga

Meteorological Office, FitzRoy Road, Exeter EX1 3PB, UK

A recent assessment by the intergovernmental panel on climate change concluded that the Earth's climate would be 2–6 °C warmer than in the pre-industrial era by the end of the twenty-first century, due to human-induced increases in greenhouse gases. In the absence of other changes, this would lead to the warmest period on Earth for at least the last 1000 years, and probably the last 100 000 years. The large-scale warming is expected to be accompanied by increased frequency and/or intensity of extreme events, such as heatwaves, heavy rainfall, storms and coastal flooding. There are also several possibilities that this large change could initiate nonlinear climate responses which lead to even more extreme and rapid (on the time-scale of decades) climate change, including the collapse of the ocean 'conveyor belt' circulation, the collapse of major ice sheets or the release of large amounts of methane in high latitudes leading to further global warming. Although these catastrophic events are much more speculative than the direct warming due to increased greenhouse gases, their potential impacts are great and therefore should be included in any risk assessment of the impacts of anthropogenic climate change.

Keywords: climate change; extreme events; sea-level rise; storms; flooding; greenhouse gases

1. Introduction

Over the last three decades, there has been growing concern that increases in atmospheric greenhouse gases will lead to substantial changes in the Earth's climate. In addition to a general increase in temperature, it has been predicted that there will be changes in the geographical distribution, intensity and frequency of extreme events (IPCC 2001).

Most of the events considered in this meeting are of a rapid and catastrophic nature—earthquakes, volcanic eruptions, tsunamis and strikes by near-Earth objects. In contrast, the effects of human-induced climate change are likely to be insidious, with a gradual increase over a number of decades in the number and intensity of many climate extremes, including heatwaves, heavy precipitation and coastal flooding. This is discussed in the first part of this paper.

There is also the possibility, however small, that the gradual change in climate leads to parts of the climate system exceeding a threshold beyond which they may not be reversed by restoring greenhouse gases to pre-industrial

* Author for correspondence (john.f.mitchell@metoffice.com).

One contribution of 20 to a Discussion Meeting Issue 'Extreme natural hazards'.

Table 1. Equilibrium sensitivity of global mean surface temperature change (climate sensitivity) for a doubling of atmospheric carbon dioxide with estimates of typical feedbacks. (Based on Mitchell (2004).)

processes included	warming with cumulated feedbacks (°C)
direct radiative heating	1.0
in addition, water vapour feedback	2.0
further adding sea ice, snow, cloud feedbacks	1.5–5.5

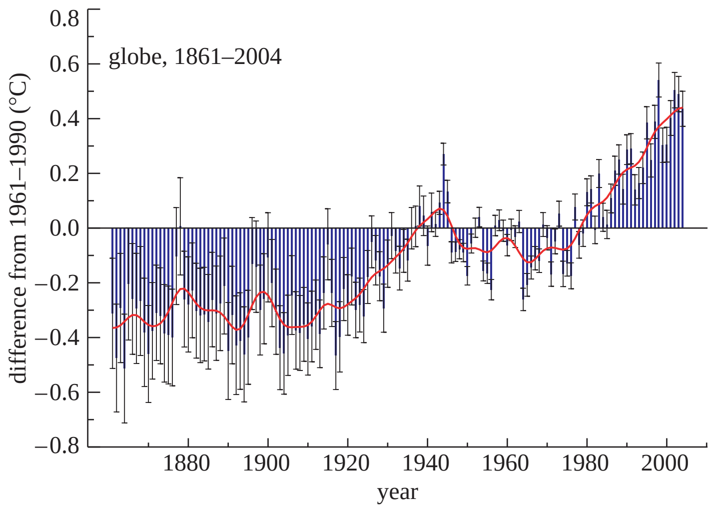

Figure 1. Global mean surface temperature anomalies (°C) for the period 1861–2004. The bars show an estimate of uncertainty for individual years. The red line is obtained using a 21-point filter which gives a near decadal average (from Parker *et al.* 2005 with permission from the Met Office. © Crown Copyright).

concentrations. Examples that have been proposed include the collapse of the North Atlantic thermohaline circulation and the disappearance of the Greenland ice sheet. Again, these events would not be sudden but would occur over decades or, in the case of sea level, over centuries, but their effect would undoubtedly be extreme. This is discussed in the second part of this paper.

2. The greenhouse effect and climate change

The Earth and atmosphere are heated by solar energy, which is balanced by the emission of thermal (long wave) radiation back to space. Atmospheric greenhouse gases absorb thermal radiation from the surface and re-emit (upwards and downwards). Water vapour is the main greenhouse gas in the current atmosphere.

Seen from space, adding greenhouse gases leads to the average level of emission to space occurring from higher, and therefore cooler, levels of the atmosphere. Since the thermal emission is reduced as temperature decreases, this leads to a net radiative warming of the surface and atmosphere. The surface and atmosphere therefore warms until the emission of thermal radiation again

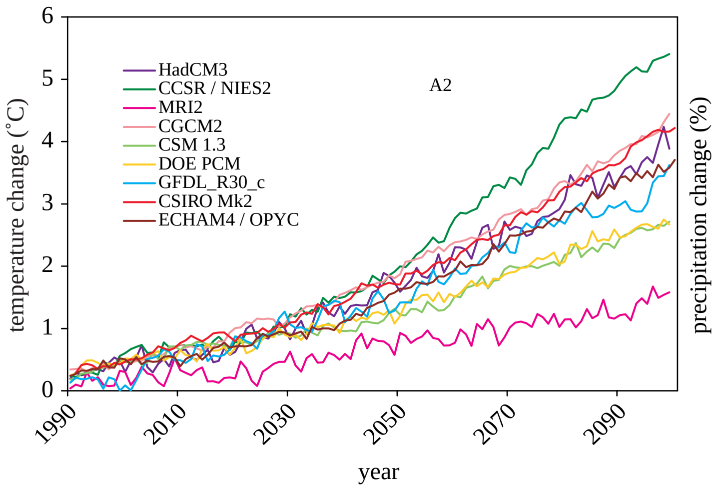

Figure 2. The evolution of globally averaged temperature (°C) changes relative to the years 1961–1990 for the SRES A2 scenario from nine general circulation models (from Cubasch *et al.* 2001 with permission from IPCC).

balances the incoming solar radiation. The main man-made contributor to increase in radiative heating is carbon dioxide, the concentration of which has increased by 33% from 1958 to 2000 (IPCC 2001). This is due principally to human activity through burning fossil fuels—coal, gas and oil. Future emissions and atmospheric concentrations of CO_2 are the subject of considerable uncertainty (e.g. Nakicenovic *et al.* 2000), but most scenarios suggest an effective doubling of the pre-industrial concentration of atmospheric carbon dioxide by about the middle of this century.

A doubling of atmospheric carbon dioxide is estimated to increase downward radiation at the Earth's surface by 3.8 W m^{-2} at the tropopause (e.g. Myhre *et al.* 1998). This compares to a net solar heating of the current Earth by about 240 W m^{-2}. (The tropopause is the boundary between the lower atmosphere, or troposphere, and the stratosphere.) If the Earth's surface and the troposphere behaved simply as a black body, this would lead to an equilibrium warming of about 1 °C. However, the warming is likely to lead to changes in other climatic elements that may enhance or reduce the basic warming (table 1). For example, warming is likely to increase the water content of the atmosphere and, as water vapour is a greenhouse gas, lead to further warming. Modelling and empirical evidence suggests that this would almost double the response. The main source of uncertainty in the physical feedbacks is due to cloud and associated processes (table 1, see also for example Senior & Mitchell 1993). More recently, other feedbacks with climate involving the biosphere (Cox *et al.* 2000; Friedlingstein *et al.* in press) and atmospheric chemistry (e.g. Johnson *et al.* 2001) have been investigated.

Observations of surface air temperature, averaged over the globe, indicate a warming of 0.7 °C from 1860 to 2000 (figure 1, from Parker *et al.* 2005). The temporal pattern of global mean warming is similar whether using independent data from land, sea surface or the air above the ocean. Attempts to simulate the temperature record using comprehensive three-dimensional climate models can only reproduce the rapid warming in the last three decades when the effect of anthropogenic greenhouse gases is included (e.g. Stott *et al.* 2000, in press) and the warming is unlikely to be explained by natural internal variability of the climate system.

Although there is a broad consistency between observations and model simulations during the 140 years of the global instrumental surface temperature record, there is a wide range in the sensitivity of models to a given increase in greenhouse gases. The apparent good agreement with observations is partly due to different estimates of the historical forcing due to factors other than well-mixed greenhouse gases, especially sulphate aerosols (which tend to cool climate and so partly offset the effects of increasing greenhouse gases; Penner *et al.* 2001), and also because models with a high climate sensitivity reach equilibrium more slowly than those with a low sensitivity (Hansen *et al.* 1985). Hence, it is important to attempt to quantify the uncertainty in climate predictions.

The range of uncertainty has been estimated by looking at the range of sensitivity from the available global three-dimensional climate models (e.g. Le Treut & McAvaney 2000). More recently, attempts have been made to estimate the uncertainty by using a particular climate model and calculating the climate sensitivity using combination of parameter values, spanning what are thought to be reasonable limits for the range of those parameters (Murphy *et al.* 2004; Stainforth *et al.* 2005). Stainforth *et al.* (2005) used results from many thousands of experiments by making the experimental set-up available to personal computers through the World Wide Web. They found few models with sensitivity below 2 °C, with most models having a value of 3–4 °C and a number of models with a sensitivity of 8 °C or larger. The skewness of this frequency distribution can be explained using a simple energy balance model approach (e.g. Dickinson 1986). Using this approach with other models would widen the range of sensitivity. However, although there is considerable uncertainty in the equilibrium climate response to a doubling of CO_2, it is likely to be of the order of 2–3 °C. In order to make effective use of climate model predictions, a second type of uncertainty must also be treated, the uncertainty in future emissions of greenhouse gases and aerosol precursors. This is traditionally dealt with by performing several simulations for different estimates of plausible future emissions (e.g. Nakicenovic *et al.* 2000). However, Schneider & Lane (2006) highlight the difficulty in using current emissions estimates in a risk assessment framework. A further uncertainty results from the conversion of emissions into concentrations, as discussed in §4*a* for the carbon cycle.

3. Mean climate change and extreme weather events

(*a*) *Atmospheric and surface changes*

In §2, we considered the equilibrium response to increases in greenhouse gases. We now consider the time-dependent response of climate to increases in greenhouse gases, taking into account the inertia and dynamics of the oceans and sea ice.

Predictions which take into account uncertainties in both greenhouse gas emissions and climate modelling suggest a warming of 1.4–5.8 °C by the end of the twenty-first century (figure 2, from Cubasch *et al.* 2001). A similar range is obtained when predictions are constrained by observations over the previous 50–100 years

(Stott & Kettleborough 2002). The CO_2 emissions scenarios ranged from 5 to about 30 GtC yr^{-1} in 2100 (current anthropogenic emissions are about 76 GtC yr^{-1}). Despite the uncertainties and range of emissions, increases in greenhouse gases are likely to have a substantial warming effect over the next century. Note also that the specification of emissions scenario has little impact on the predicted temperature changes over the next two to three decades—some measure of climate change in the next few decades is probably inevitable. To put the changes in context, the predictions are also plotted along with reconstructions of the Northern Hemisphere climate record over the last 1000 years (figure 3). Even taking low emission scenarios, low sensitivity climate models and allowing for uncertainty in the reconstructions, the predicted changes are unprecedented compared to the last millennium, and possibly since the last interglacial.

The increases in temperature are, not surprisingly, expected to lead to an increase in the frequency of extreme warm periods (e.g. Kharin & Zwiers 2005). Increases in high extremes may also be exacerbated by changes in other aspects of climate, including circulation (e.g. Meehl & Tibaldi 2004), or a drying out of the ground that reduces evaporative cooling. It is not possible to attribute an individual extreme event to increases in greenhouse gases. However, one can estimate the effect of increases in greenhouse gases on the likelihood of an extreme event, in much the same way as one can estimate the increased risk of premature death due to smoking. A recent study (Stott *et al.* 2004) estimated that the increase in anthropogenic greenhouse gases to date has probably doubled the likelihood of a heatwave in Europe of the severity of that experienced in summer 2003 (Schar *et al.* 2004). The heatwave was estimated to have led to an increase of a few tens of thousands premature deaths. Furthermore, simulations of future climate (based on the IPCC Special Report on Emissions Scenarios (SRES) A2 emissions scenario; Nakicenovic *et al.* 2000) suggest that such summers would occur about every second year by about 2050, and would be regarded as a cool summer by the end of this century (figure 4, based on Stott *et al.* 2004).

The increase in mean temperature increases the saturation vapour pressure of water, and hence potentially the water content of the atmosphere. All other things being equal, this would, in turn, lead to increased intensity in precipitation events. If relative humidity (the fractional saturation of the atmosphere) is preserved, which appears to be true to first order in model simulations, then the water content of the atmosphere would increase by about 6% for each degree Celsius of warming, following the Clausius Clapeyron equation. J. Gregory (quoted in Allen & Ingram 2002) found that the most extreme precipitation rates did increase by this level in a simulation with increased atmospheric carbon dioxide, but the intensity of moderate and light events decreased. The increase in atmospheric water increases the radiative warming of the surface, and combined with the radiative cooling of the free atmosphere (Mitchell *et al.* 1987) tends to reduce the static stability of the atmosphere, supporting more intense precipitation.

Global-averaged precipitation rates increase in simulations with increased greenhouse gas concentrations, but at a rate of only 1–3% per degree warming (e.g. Le Treut & McAvaney 2000). However, this global figure masks strong regional variations, with intensification in the transport of water by the atmosphere from the subtropics to high latitudes (e.g. Wu *et al.* 2005).

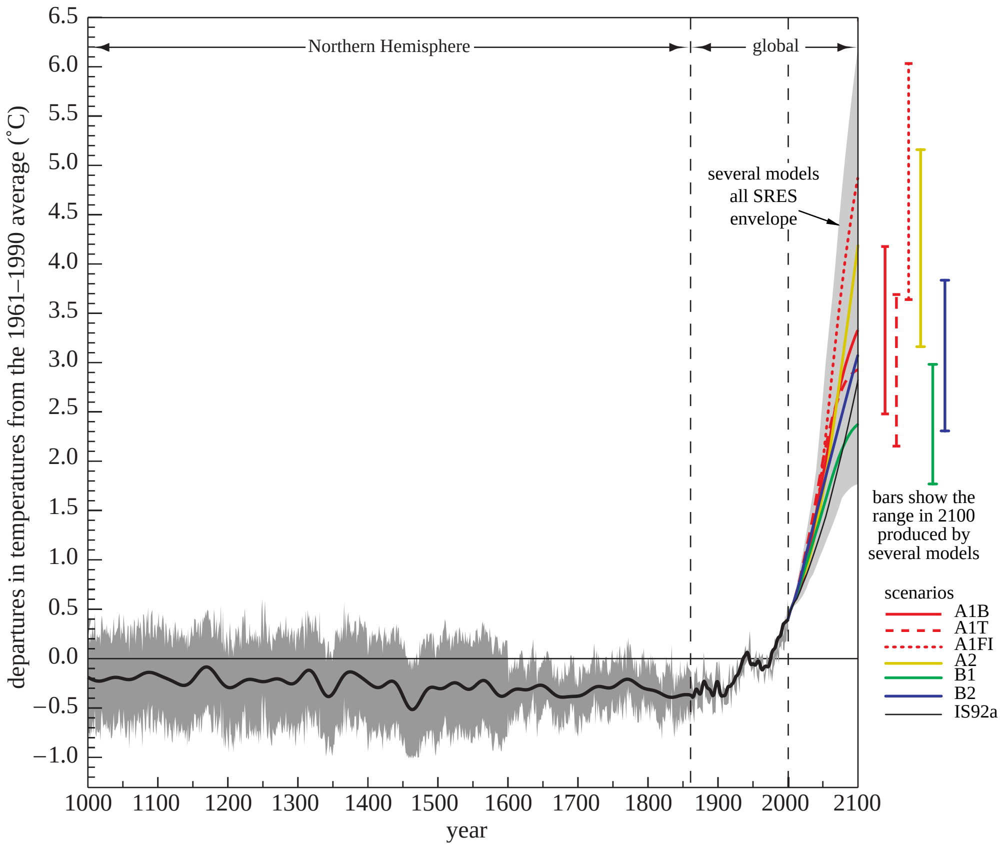

Figure 3. Comparison of global mean projections of global average surface temperature (°C) change from 2000 to 2100 with reconstructions of Northern Hemisphere average temperature changes (IPCC 2001). Shading before 2000 indicates uncertainty in the observations, shading after 2000 indicates the range of uncertainty due to the range of emissions uncertainties and modelling uncertainties. The individual curves after 2000 represent individual emission scenarios. Figure reprinted with permission from IPCC.

The extra latent heat available through increased sea surface temperatures and higher precipitation rates has the potential to increase the intensity of atmospheric disturbances. Both theoretical (Emanuel 1987) and recent observational evidence (Emanuel 2005) suggest an increase in the destructive power of the most extreme hurricanes. For example, Emanuel (2005) considered the contribution from the wind speed cubed (assumed to be a measure of destructive power) and the duration of hurricanes observed over the last 30 years. Some, but not all, of this increase can be attributed to increases in sea surface temperatures, but that warming may be a result of natural climate variability rather than increasing greenhouse gases. Webster *et al.* (2005) also found an increase in the frequency of Category 4 and 5 hurricanes in the period since 1970. Most climate models do not resolve tropical storms adequately, and hence rely on the correlation between large-scale circulation indices and hurricanes to estimate the frequency and intensity of hurricanes. Simulations using regional models which do resolve tropical storms suggest a modest increase

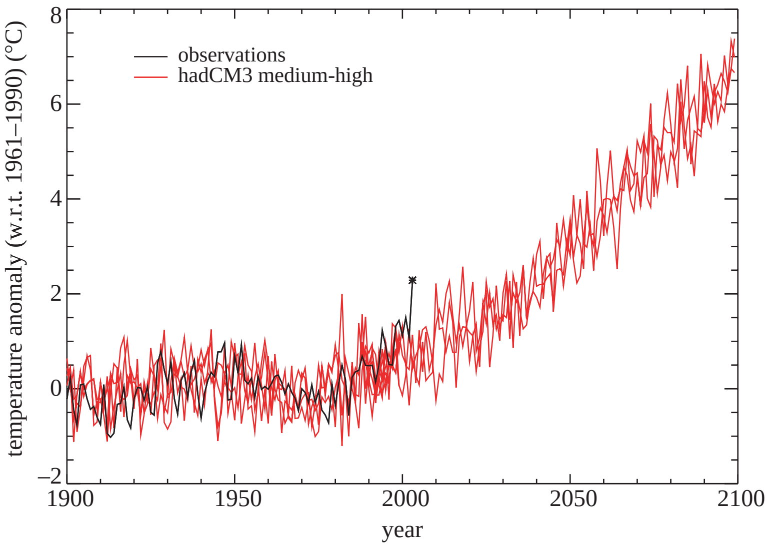

Figure 4. Anomalies of European area average summer surface temperature anomalies (°C). The black curve is based on observations from 1900, ending with the warm summer of 2003. The red curve is an ensemble of model simulations with the Hadley centre climate model HadCM3 forced with an estimate of past forcing to 2000 and a medium-high emission scenario to 2100 (based on Stott *et al.* 2004 with permission from the Met Office).

in future intensity (e.g. Knutson & Tuleya 2004). The evidence for extra-tropical storms is less clear, although a number of model studies suggest an increase in the intensity of most intense storms (e.g. Cubasch *et al.* 2001; Fyfe 2003; Leckesbusch & Ulbrich 2004).

(*b*) *Sea-level and coastal impacts*

Coastal regions are vulnerable to damage from a range of extreme natural hazards. Recent examples of natural events include the December 2004 tsunami, which affected the nations around the Indian Ocean, and the storm surge driven by Hurricane Katrina, which affected the United States. The events had very different causes (the tsunami had a geological cause, whereas the US flooding was driven by meteorological conditions), but each provides an example of the present-day vulnerability of the coastal zone to capital damage and casualties. A report commissioned by Defra (Halcrow Group Ltd 2001) estimated that £222 billion of assets are potentially at risk from flooding in the United Kingdom, with around 59% of the total being property at risk from sea and tidal flooding. At current defence standards, the annual average damages were estimated to be around £0.8 billion per year. The most severe coastal damage tends to result from short-lived extreme flooding events, whose characteristics are likely to alter in the future as a result of both increases in mean sea level and changes in storm-surge activity.

The large-scale planetary warming mechanisms discussed in §3*a* will lead to an increase in mean sea level due to thermal expansion of the ocean as it warms and the melting of additional ice over land as the atmosphere warms. Estimates of the global mean sea-level rise by the end of the twenty-first century (Church *et al.* 2001) vary from about 0.1 to 0.9 m (figure 5), with considerable variation across the globe and little agreement in the geographical pattern from model to model. The impact of these changes may be augmented or reduced on a local scale by the vertical movements of the land; for example, the south east of England is currently sinking, whereas Scandinavia is rising.

Storm surges are short-lived increases in water level driven by low atmospheric pressure and strong winds acting in shallow seas. They are made worse if the coastal geometry is such that it can funnel the water, as it does at the southern end of the North Sea and in the northern Bay of Bengal. We discuss changes in both of these regions below. The impacts of the surge are worse when they happen at or near a high tide.

In order to simulate the effect of climate change on storm-surge height and frequency, a series of climate models were used, starting with a global climate model, then adding local detail with a regional climate model. The atmospheric outputs from the regional model were then used to drive a depth-average storm-surge model developed by the Proudman Oceanographic Laboratory (Flather *et al.* 1998).

Around the British Isles, the global model was used to drive a regional model and a 35 km surge model (Lowe & Gregory 2005). Figure 6 shows the simulated return period curve just offshore from the Thames Estuary region for the present day and in the 2080s after allowing emissions to follow the SRES A2 scenario. The future prediction combines the effects of the mean sea-level rise, the effect of changing weather on surges and the vertical land movement. The extreme water level shown in figure 6 is measured relative to the present-day tide and the curve shows that the magnitude of event experienced on average once every 50 years in the present-day climate might be expected more often than every other year by the end of the twenty-first century.

In a study focusing on the Bay of Bengal, a similar technique was employed, but the older HadCM2 and HadRM2 climate models were used. The surge model also had a higher resolution of around 10 km. Figure 7 shows the change in height of a 50-year return period storm surge (expected, on average, once every 50 years) between pre-industrial times and the 2050s. Beyond 1990, greenhouse gas emissions followed a business-as-usual (IS92a) scenario (Lowe & Gregory 2005). Again, changes in storminess, mean sea-level rise and vertical land movement are all considered. The largest changes are predicted for the northwest of the Bay, reaching more than 0.75 m. An interesting aspect of this study was that the largest simulated storm in the Bay did not always lead to the largest surge. This occurred because the track and timing of the storms are also important. The most intense storms did not follow the track that could potentially cause most damage.

Predicting changes in storms and the storm surges they can produce on these 100-year time-scales is an uncertain process, and two of the major challenges facing climate prediction are to quantify (e.g. Woth 2005) and then reduce this uncertainty. The results of this type of modelling are being increasingly used in risk estimate of flooding (e.g. Muir Wood & Bateman 2005).

4. Long-term changes, instabilities and irreversible events

So far we have considered changes in extreme weather events which are a direct consequence of higher mean temperatures, increases in sea level, greater atmospheric water content and so forth. We now consider the possibility of more drastic changes to the climate system. These are generally longer term and in many cases more speculative than the changes outlined above.

(*a*) *Carbon cycle*

As noted earlier, most simulations to date omit feedbacks between climate and the carbon cycle. In the last few years, a number of simulations have been completed which allow the land and ocean carbon cycle to interact with the changes in climate as anthropogenic emissions are increased (Friedlingstein *et al.* in press). All these simulations show a positive feedback, that is, the changes in the carbon cycle lead to an even greater increase in CO_2, even though increasing CO_2 stimulates the growth of certain types of plant leading to an increased storage of carbon on land. All the models in the study showed a decrease in the net uptake of carbon in the ocean, essentially because the ocean is warmer (and warmer water takes up less carbon dioxide) and, in some cases, vertical mixing is inhibited in the warmer climate. There is also a reduced rate of uptake over land, primarily because respiration of carbon from the soil increases with temperature.

In one study (Cox *et al.* 2000), the terrestrial carbon cycle becomes a net source of carbon towards the end of the twenty-first century. This is largely a result of a widespread increase in the rate of soil carbon respiration in the warmer climate also seen in other models. (In this model, the die-back from Amazon vegetation due to reduced precipitation is also important.) The effect of carbon cycle–climate feedbacks in this model increases the average warming over land by 2100 from 5 to 8 °C. The change in the other models considered by Friedlingstein *et al.* (in press) is less severe; most models show a reduced uptake of carbon in the tropical forests, but they generally remain as a sink of carbon. Clearly, there is some way to go before these uncertainties are resolved.

(*b*) *The Atlantic thermohaline circulation*

Much of the Northern Hemisphere is kept considerably warmer than it would otherwise be by the northward penetration of warm, salty water from the subtropical Atlantic to high latitudes, part of the so-called ocean thermohaline circulation or THC (sometimes imprecisely referred to as the 'Gulf Stream'). When the warm, salty water reaches the subpolar regions, it loses its heat to the atmosphere, becoming denser and sinking. Eventually, the cooled water returns southward at depth along the eastern coast of America.

Simple models (e.g. Stommel 1961) suggest that the THC may have two alternative states, one with strong flow (analogous to the present situation) and a second state with very weak circulation. An increase in the strength of the atmospheric water cycle leads to a greater excess of precipitation over evaporation in high latitudes, freshening and reducing the density of surface water, and hence its propensity to sink and weaken the THC. Beyond a certain threshold in the strength of the atmospheric water cycle, only the state with weak THC can exist. This is potentially important because the expected strengthening of the water cycle in response to increasing greenhouse gases might be expected to push the climate system nearer to such a threshold, 'switching off' the THC. If this threshold were passed, the simple models suggest that restoring the water cycle to its original strength may not restart the THC. Evidence of this behaviour from more comprehensive climate models is inconclusive. A 'weak

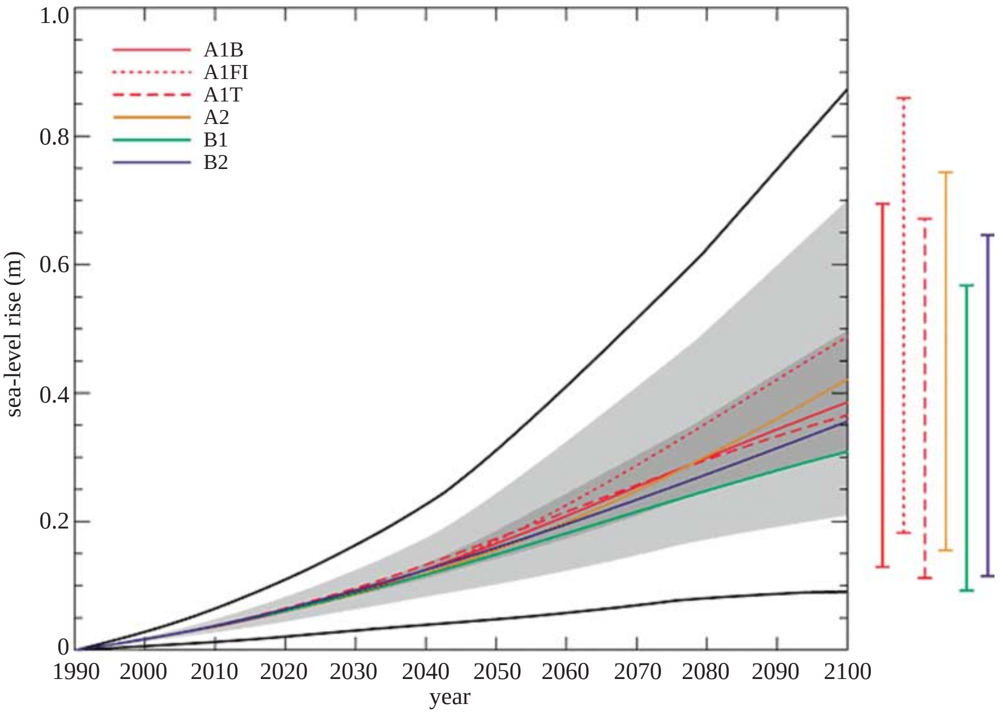

Figure 5. Global average changes in sea-level rise from 1990 to 2100 for the IPCC SRES scenarios (m). The coloured lines show sea-level rise for six illustrative scenarios, using a simple climate model calibrated to nine general circulation models. The dark shading shows the range of predictions for all 35 SRES scenarios. The light shading covers the range of model and emission scenarios. The solid black lines include the additional uncertainty from changes in land ice, permafrost changes and sediment deposition. The bars show the model uncertainty range at 2100 for the illustrative scenarios (from Church *et al.* 2001 with permission from IPCC).

THC' state has been shown to exist in a few such models (Manabe & Stouffer 1999; Rind *et al.* 2001), but when forced with plausible scenarios of increasing greenhouse gases, these models do not suggest the crossing of a threshold to the weak THC state. In these more comprehensive models, further feedbacks involving changes in tropical precipitation and evaporation, transport of salinity anomalies by the wind-driven circulation and local forcing of deep water formation by the wind tend to stabilize the present THC state (e.g. Schiller *et al.* 1997; Thorpe *et al.* 2001; Vellinga *et al.* 2002). For these reasons, a near-shutdown of the THC is considered unlikely over the next 100 years.

Nonetheless, if the THC were to shut down (especially if this were to happen over just a few decades, as is believed to have occurred during the early stages of the present interglacial period), there would be drastic consequences for global climate. These have been simulated, for example, by Vellinga & Wood (2002, in press), who artificially induced a THC shutdown in their model by the addition of a large quantity of fresh water to the North Atlantic. Climatic impacts include cooling and drying through much of the Northern Hemisphere and major shifts in tropical rainfall patterns (notably, substantial drying in Central America and the Indian monsoon zone). In addition, substantial sea-level

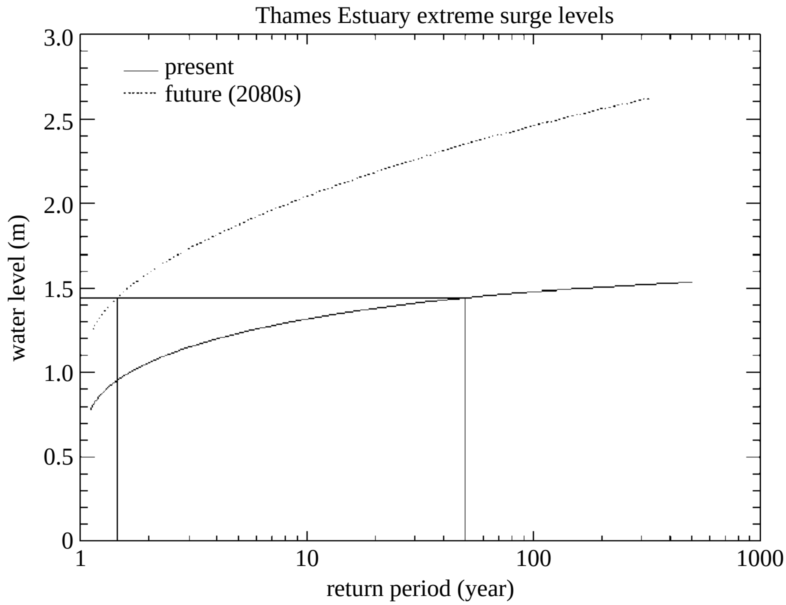

Figure 6. Return period curve for the southern North Sea near the Thames Estuary. Changes in storminess, mean sea-level rise and vertical land movement were included. The future results are for a medium-high scenario following the SRES A2 emissions. The changes are expressed relative to present-day tide.

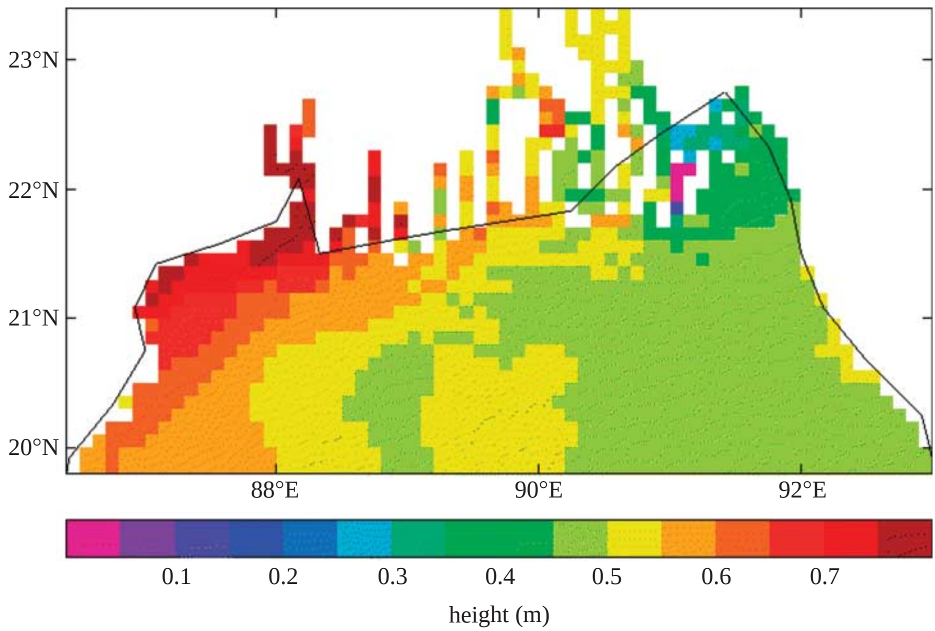

Figure 7. Predicted changes in the height of a 50-year return period event measured relative to present-day tide in the Bay of Bengal between the 2050s and the pre-industrial era. The result includes changes in storminess, an increase in mean sea level and vertical land movement.

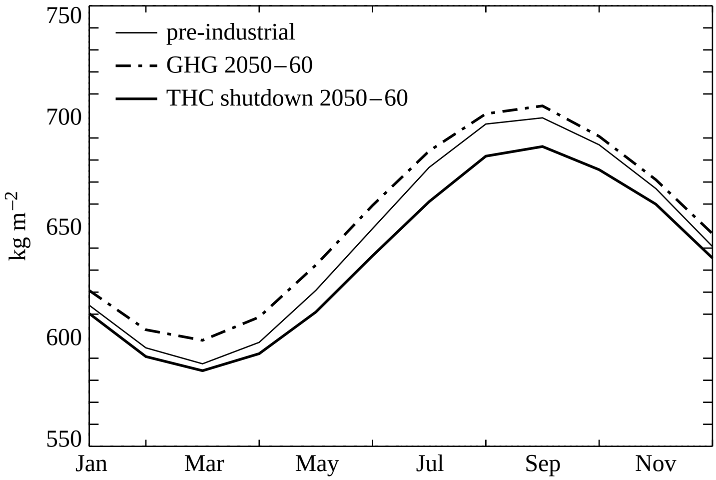

Figure 8. Simulated annual cycle of soil moisture content, averaged over the region between 0 and 25°N, in three model simulations: long-term average with fixed pre-industrial greenhouse gases (thin solid), 2050s under a plausible scenario of increasing greenhouse gases (dot-dash) and a *hypothetical* scenario in which the THC shuts down in 2050 following the same scenario of increasing greenhouse gases (thick solid). Increasing greenhouse gases produce an enhancement of tropical precipitation that results in increasing soil moisture, but a THC shutdown would reduce precipitation in this region. The net result in this case would be a reduction of soil moisture. GHG, greenhouse gas.

rise around the North Atlantic and reduced uptake of carbon by the oceans could be expected. Some of the changes would be in the opposite direction to those expected from the direct effects of greenhouse gases (e.g. in the northern tropics, drying due to the THC shutdown would overwhelm wetting due to greenhouse gas increase; see figure 8). Other changes would reinforce the effects of the greenhouse gases (e.g. adding to sea-level rise around the North Atlantic and the summer drying over Southern Europe that is seen in most climate simulations).

In most simulations of the response of the THC to increasing atmospheric greenhouse gas concentrations, the ocean surface warms and freshens, with the freshening being due to increased high-latitude precipitation (Wu *et al.* 2005). This decreases the surface water density, resulting in a weakening of the THC, but not a shutdown as described previously. In this (more likely) scenario, the warming accompanying the increase in greenhouse gases generally overwhelms the cooling due to the slow down in the circulation. However, reduced warming over the ocean limits the increased flux of moisture from ocean to land that would otherwise occur (Manabe *et al.* 1992). As a consequence, the weakening of the THC exacerbates the drying over southern Europe. Thus, even if a complete shutdown is unlikely, it is still important to predict accurately the amount by which the THC may weaken in future. Unfortunately, there is a large range in the weakening predicted by different models, reflecting an incomplete quantitative knowledge of the processes that determine the THC response. In the third assessment report of the intergovernmental panel on climate change, models produced weakening of the THC ranging between about 0 and 50% (Cubasch *et al.* 2001). This high level of uncertainty is currently being addressed by a systematic, international multi-model analysis, aimed at determining the key processes responsible for the range of responses among models (Gregory *et al.* 2005). Ultimately, it is hoped that this understanding will allow critical observational tests to be devised that will narrow the range of plausible THC projections.

(*c*) *Sea-level rise and deglaciation of the Greenland ice sheet*

The impacts of sea-level rise during the twenty-first century were discussed in §3*b*. However, sea level presents an additional challenge, in that once a change has been initiated it may continue well beyond the next century, and possibly beyond the next millennium, even if surface temperatures are stabilized rapidly. The thermal expansion component of sea-level rise is slow to stabilize because it depends on the adjustment of the slowly responding deep ocean. The land ice contribution on these long time-scales is likely to be dominated by the large ice sheets, which also respond slowly. Recent work at the Hadley Centre has looked at the likelihood of causing a partial or complete deglaciation of the Greenland ice sheet (Gregory *et al.* 2004; Lowe *et al.* 2006), and how long such a process might take (Ridley *et al.* 2005). Using a simple climate model, Gregory *et al.* (2004) demonstrated that even quite modest future levels of greenhouse gas concentration might lead to deglaciation. Lowe *et al.* (2006) used an ensemble of more complex climate models to show that if carbon dioxide concentrations stabilize at levels of around 650 p.p.m.v., then around half of the models in an ensemble of plausible but slightly different climate models could exceed the warming threshold for partial or complete Greenland deglaciation. Even if CO_2 were stabilized at a much lower level of 450 p.p.m.v., then around 5% of the models in the ensemble would still result in the deglaciation threshold being exceeded there. Ridley *et al.* (2005) demonstrated that more than half of the ice sheet mass (amounting to more than 3.5 m of sea-level rise) was lost after a millennium of climate forcing corresponding to atmospheric carbon dioxide being stabilized at four times the pre-industrial level. The peak rate of melting provided enough freshwater into the model ocean to have a small but noticeable effect on the model's ocean circulation, temporarily reducing the THC by a few percent. However, this was not enough to lead to widespread Northern Hemisphere cooling.

A further question related to Greenland is that of whether once triggered, the deglaciation would be irreversible if the greenhouse gas forcing were later reduced. Without the ice sheet, the surface is warmer, which may explain why the ice sheet does not reform. However, 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 another active area of current research.

These large potential increases in mean sea level from thermal expansion and ice sheet melting would inundate many inadequately protected coastal areas. Even where this was not the case they would make extreme high water events much more likely. Thus, although these predictions are uncertain, they demonstrate that even with a sizeable mitigation effort, future coastal populations will need to adapt to a very large eventual rise (Nicholls & Lowe 2004). The west Antarctic ice sheet could cause additional sea-level rise on long time-scales (Church *et al.* 2001). However, predictions of changes to this ice sheet are even more uncertain and are not discussed here.

(d) Clathrates

It has been estimated that 10^{16} kg of methane are trapped below the sea floor and in permafrost in ice-like solids known as clathrate hydrate (see Buffet 2000 for a recent review). There has been speculation that an increase in ocean bottom temperature could lead to the destabilization of marine hydrates, particularly methane, which is a strong greenhouse gas. This may be slightly offset by the stabilizing effect of an increase of bottom pressure due to addition of mass to the ocean through melting land ice. It is thought that a large release occurred in the Late Paleocene 55 million years ago, explaining the anomaly of δ^{13}C at that time (e.g. Zachos *et al.* 1994). Xu & Lowell (2001) (based on a one-dimensional model) argue that the pressure effects are small, and the temperature effect on hydrate accumulation and dissociation near the sea floor is significant only at higher temperatures, such as those that occurred during the Paleocene. If this is the case, then any such effect is unlikely in the next hundred or so years. However, this is a topic of considerable uncertainty.

5. Concluding remarks

Despite the many uncertainties in modelling climate and the wide range of emission scenarios, substantial climate change is almost certain to occur over the next few decades. Increases in global mean temperature of 1.4–5.8 °C are predicted to occur by the end of the twenty-first century (IPCC 2001). The mean changes will lead to higher temperature extremes, greater extremes in rainfall intensity and greater extremes in storm surges and coastal flooding.

There are also longer-term changes which must be regarded as more speculative. The terrestrial biosphere, which currently is a sink for atmospheric carbon dioxide, could become a source. The overturning circulation in the North Atlantic could slow down markedly or stop, although indications are that any local cooling would be more than offset by the warming effect of increased greenhouse gases. However, a slowing of the circulation would enhance the predicted drying of the land surface in southern Europe in summer. If the local warming around Greenland exceeds 3 °C (true for most model simulations and emission scenarios) and is sustained, then the Greenland ice sheet could be removed over several thousand years. There is some evidence from model simulations that if this were to occur, then it would not reform even if greenhouse gas concentrations returned to their current values. Finally, there has been speculation that the oceanic warming might lead to the release of methane trapped in the ocean floor, further enhancing warming.

The simulation of changes in extreme climate events is a difficult problem, as they are often associated with highly nonlinear local events that are particularly hard to model. Many of the processes which contribute to climate change, including those regarding the formation and dissipation of cloud and modelling vegetation, are insufficiently understood, presenting a challenge for many years to come. However, ongoing research is improving our understanding of the current range of modelling results in these areas, and beginning to provide observational constraints that will allow that range to be narrowed in future. This will contribute to the future development of more soundly based risk analyses for such events by the wide range of communities that will be affected by climate change.

We thank Peter Cox at the Centre for Ecology and Hydrology for discussions concerning carbon cycle–climate feedbacks and Roger Flather at the Proudman Oceanographic Laboratory for providing the European and Bay of Bengal surge models. Jason Lowe, Richard Wood and Michael Vellinga are supported by the UK Department of Environment, Food and Rural Affairs under contract PECD7/12/37.

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