with a lattice that has an in-plane orientational epitaxy with the underlying mica lattice. The only reason x-ray analysis can be carried out on these structures is because we are able to signal average over a large collection of prisms that are aligned with one another and epitaxially arranged on the mica support.

This approach to controlling and monitoring the kinetics of crystal growth can be used to study environment-imposed changes in crystal morphology (*28*). Subtle changes in temperature markedly affect the growth of the crystals and the observed morphology of the crystals ultimately formed. Indeed, when the temperature is increased to 35°C, cubic-shaped features emerge at the edges of the prisms while scanning the crystals that were preformed at lower temperature with the PLH-coated AFM tip (Fig. 4). This morphological change is very reproducible and always was induced at the corners or edges of the starting triangular crystals.

This study provides an approach for site-specifically initiating crystal growth on the nanometer-length scale in a way that allows one to monitor growth from crystal seed to more mature structures as a function of environmental conditions (fig. S5). The size of the smallest crystal observed and studied in these experiments (d in Fig. 1) is five orders of magnitude smaller than what could be studied by single-crystal XRD techniques, allowing one to observe morphological changes that would typically go undetected in an x-ray study that focuses on larger structures. Finally, growing crystals of macromolecules is not a trivial process. DPN is now a massively parallel tool (*16*, *29*, *30*), suggesting that this study may open the door for creating combinatorial approaches to identifying the proper conditions to initiate a particular type of crystal growth for a given set of target molecules.

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The Climate Change Commitment

T. M. L. Wigley

Even if atmospheric composition were fixed today, global-mean temperature and sea level rise would continue due to oceanic thermal inertia. These constant-composition (CC) commitments and their uncertainties are quantified. Constant-emissions (CE) commitments are also considered. The CC warming commitment could exceed 1°C. The CE warming commitment is 2° to 6°C by the year 2400. For sea level rise, the CC commitment is 10 centimeters per century (extreme range approximately 1 to 30 centimeters per century) and the CE commitment is 25 centimeters per century (7 to 50 centimeters per century). Avoiding these changes requires, eventually, a reduction in emissions to substantially below present levels. For sea level rise, a substantial long-term commitment may be impossible to avoid.

Oceanic thermal inertia causes climate change to lag behind any changes in external forcing and causes the response to be damped relative to the asymptotic equilibrium response (*1*–*3*). Because of this lag or damping effect, and because of the changes in atmospheric composition (and radiative forcing) that have already occurred, the climate system will continue to change for many decades (centuries for sea level) even in the absence of future changes in atmospheric composition. For global-mean temperature, this is referred to as the "unrealized warming" (*2*), "residual warming" (*4*), or "committed warming" (*5*). Here, I use the term "warming commitment" or, to include sea level rise (*6*, *7*), "climate change commitment."

The assumption of constant atmospheric composition on which the warming commitment idea is based is clearly unrealistic, even as an extreme case of what might happen in the future. An alternative indicator of the commitment to climate change is to assume that the emissions (rather than concentrations) of radiatively important species will remain constant. This Report investigates the constant-composition (CC) warming and sea level commitments, the constant-emissions (CE) commitments, and the uncertainties in each. Uncertainties arise from uncertainties in the climate sensitivity (*2*, *4*), the rate of ocean heat uptake (*2*), the magnitude of past forcing, and the ice melt contribution to sea level change.

The usual (or "equilibrium") CC warming commitment at time t is the difference between the equilibrium warming for forcing at this time (ΔT_e) and the corresponding realized warming (ΔT_r), $\Delta T_e - \Delta T_r$. This is related to the "radiation-imbalance" concept (*8*, *9*). If ΔQ is the forcing to date, and if ΔQ_r is the forcing that gives an equilibrium warming of ΔT_r, then the radiation imbalance is $\Delta Q - \Delta Q_r$ [$\Delta Q - \Delta Q_r$ is approximately equal to the flux of heat into the ocean (*9*)]. Hence

$$\Delta T_e - \Delta T_r = (\Delta Q - \Delta Q_r)(\Delta T2\times/\Delta Q2\times)$$

where $\Delta Q2\times$ is the radiative forcing for a CO_2 doubling (about 3.7 W/m^2) and $\Delta T2\times$ is the corresponding equilibrium global-mean warming. A central estimate of ΔQ (accounting for both natural and anthropogenic forcings) is about 1.7 W/m^2, whereas ΔT_r is about

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0.7°C. Given $\Delta T2\times$ = 2.6°C (*10*), a central value for the current equilibrium warming commitment is about 0.5°C, with a corresponding radiation-imbalance estimate of 0.7 W/m^2. These results are in accord with other estimates in the literature, but uncertainties are large.

Because it would take an infinite time for the unrealized warming to appear, a more useful definition makes the unrealized warming a time-dependent quantity, namely, the evolving changes in global-mean temperature that would result if atmospheric composition were kept constant at its present state (*4*). This is the definition I use here. Temperatures under this new definition tend asymptotically to the previous equilibrium commitment definition. The new definition can be applied equally to the CC and CE commitments and can be used for both temperature and sea level.

To quantify the changes in global-mean temperature and sea level that would occur if either atmospheric composition or the emissions of radiatively important gases were kept constant at today's levels (the year 2000 is used to define "today"), I used the simple coupled gas-cycle/climate model MAGICC (*10*–*12*). MAGICC has been calibrated against a range of coupled atmosphere/ocean general-circulation models (*13*, *14*) and was used in the Intergovernmental Panel on Climate Change (IPCC) Third Assessment Report (TAR) and earlier IPCC reports to produce the standard projections of global-mean temperature and sea level change. For access to MAGICC, see (*15*).

For sea level rise commitments, a change has been made in the way the melt contribution from land-based Glaciers and Small Ice Caps (GSICs) is calculated. In the TAR, the GSIC formula was only meant to be applied through the year 2100 (I project to the year 2400 here). Because of an empirical area-correction term used in the TAR (*16*), GSIC results are unrealistic beyond 2100, and the correction term imposes an artificial melt maximum (*17*). The modified formulation (*17*) matches the TAR results well through the year 2100 and then tends asymptotically to the initially available GSIC ice mass (taken as 40-cm sea level equivalent).

The other TAR sea level rise terms are (*16*) thermal expansion (a direct output of the climate model), mass-balance changes for the Greenland and Antarctic ice sheets, long–time-scale changes in these ice masses due to past climate change, deposition of sediments on the ocean floor, and runoff from the thawing of permafrost. In the TAR formulation, the last three components (referred to here as "unforced contributions") are independent of past forcings. To quantify nonexpansion uncertainties, I used methods employed in the TAR.

For the CC and CE temperature commitments, the primary sources of uncertainty are past radiative forcing, the climate sensitivity, and the rate of ocean heat uptake. For past forcing, I considered the effect of natural forcings from solar irradiance changes (*18*) and volcanic eruptions (*19*), and uncertainties in aerosol forcing. For climate sensitivity, I used a central value of $\Delta T2\times$ = 2.6°C and a range of 1.5° to 4.5°C, approximately equal to the 90% confidence interval (CI) (*10*). For ocean mixing, I used vertical diffusivities (K_z) of 1.3, 2.3, and 4.1 cm^2/s, also representing the 90% CI and median values (*10*).

A breakdown of the natural and anthropogenic components of the CC commitment, together with uncertainties arising from ocean mixing (K_z) uncertainties, is given in table S1. Past natural forcing (inclusion of which is the default case here) has a marked effect. The natural forcing component is surprisingly large, 64% of the total commitment in 2050, reducing to 52% by 2400. The effect of ocean mixing uncertainties is small, at most 7%.

Overall results and uncertainties associated with aerosol forcing and the climate sensitivity are shown in Fig. 1 (CC case) and Fig. 2 (CE case). Aerosol forcing is characterized by the forcing in 1990. The central values (and uncertainty ranges) are those used for global-mean warming projections in the TAR (*10*, *13*): –0.4W/m^2 (–0.3 W/m^2 to –0.5 W/m^2) and –0.8 W/m^2 (–0.4 W/m^2 to –1.2 W/m^2) for direct and indirect sulfate forcing, and –0.1 W/m^2 (–0.2 W/m^2 to +0.1 W/m^2) for the sum of biomass and fossil and organic carbonaceous aerosols. The central value for total aerosol forcing is –1.3 W/m^2 (range, –0.6 W/m^2 to –1.9 W/m^2). Results depend primarily on the total aerosol forcing rather than the specific breakdown into different forcing categories. Extreme combinations, such as high climate sensitivity with low aerosol forcing, have very low probability (*20*).

In the CC case (Fig. 1), both climate sensitivity and aerosol forcing uncertainties are of similar importance. The eventual (equilibrium) commitment could be larger than 1°C (for high sensitivity and low aerosol forcing; low aerosol forcing means a higher value for past total forcing). This result is consistent with Wetherald *et al*. (*5*) because the Geophysical Fluid Dynamics Laboratory (GFDL) model used by these authors has a high sensitivity (4°C) (*14*). At the other extreme, the eventual commitment could be less than 0.2°C (for low sensitivity, virtually independent of the magnitude of aerosol forcing).

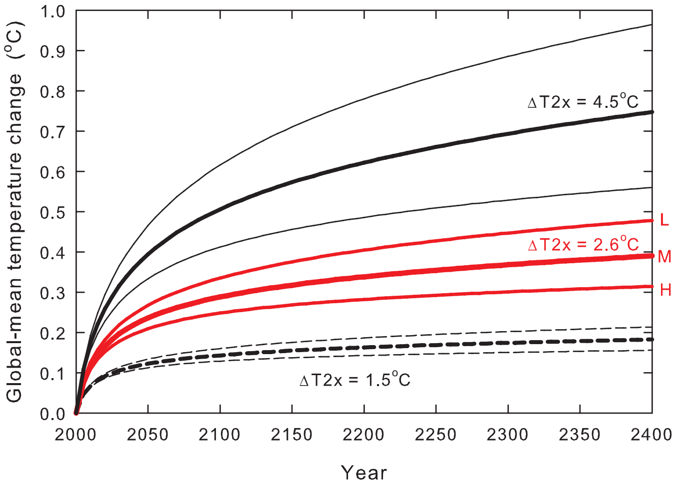

Fig. 1. CC warming commitment (constant concentrations after 2000) for different climate sensitivities and aerosol forcing levels (L, M, and H on the right of the figure indicate low, mid-, and high magnitudes for aerosol forcing, respectively). Values for the central sensitivity value (2.6°C equilibrium warming for a CO_2 doubling) are shown in red.

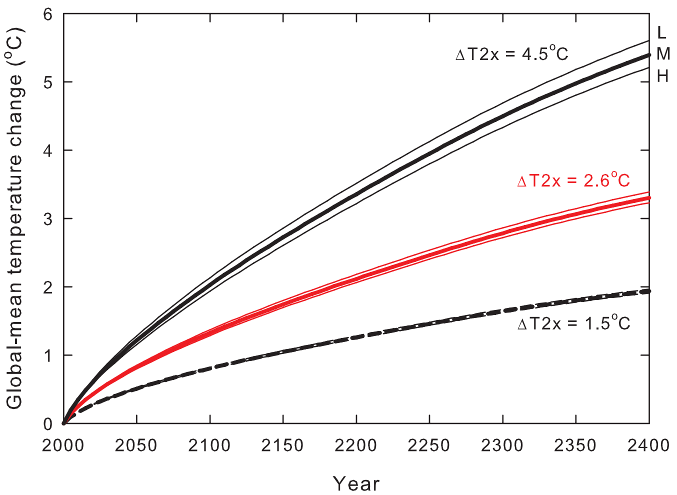

Fig. 2. CE warming commitment (CEs after 2000) for different climate sensitivities and aerosol forcing levels (L, M, and H on the right of the figure indicate low, mid-, and high magnitudes for aerosol forcing, respectively).

Warming commitments for the CE case are much higher and do not tend to any asymptotic limit even on a time scale of millennia (largely because, at CE, CO_2 concentrations continue to grow for many centuries). Climate sensitivity uncertainties are the dominant source of commitment uncertainty. By 2400, the warming ranges from about 2°C (for low sensitivity) to almost 6°C (high sensitivity). The clear message here is that, if we are to avoid future warming of this magnitude, emissions of radiatively active gases will have to be reduced to substantially below present levels.

For the sea level rise commitment results, we have an additional source of uncertainty in the ice melt and unforced contributions to sea level rise. Table S2 shows uncertainties in the CC commitment arising from ocean mixing uncertainties and gives a breakdown of the sea level rise commitment into contributions due to past natural forcing, past anthropogenic forcing, and unforced contributions.

Uncertainties in the CC sea level commitment resulting from uncertainties in ocean heat uptake arise in two ways. First, the rate of ocean heat uptake affects the rate of atmospheric warming, which affects the rate of melt of land-based ice. Second, the rate of ocean heat uptake directly affects oceanic thermal expansion. For the temperature/melt effect, larger K_z leads to a larger warming commitment, which increases sea level rise. For expansion, larger K_z leads to a greater expansion commitment, also increasing sea level rise. For the commitment, both effects act in concert, in contrast to absolute changes that have compensating effects that reduce overall sensitivity to ocean mixing uncertainties. However, the overall commitment uncertainties arising from K_z uncertainties are small at 7 to 9%.

The breakdown into natural forcing, anthropogenic forcing, and unforced effects shows that unforced effects make a substantial contribution (for the CC case, less so for the CE case). For sea level changes arising from past forcings, anthropogenic forcing dominates.

Climate sensitivity and sulfate aerosol forcing uncertainties for the sea level commitment are shown in Figs. 3 and 4. The CC case, constant concentrations from 2000, is shown in Fig. 3. The central commitment estimate (mid-aerosol forcing, $\Delta T2\times$ = 2.6°C, mid-melt) is a continuing rise of around 10 cm/century, of which about 40% is due to unforced effects (table S2) (*15*). Aerosol uncertainty effects (low, middle, and high values for 1990 forcing) are indicated by the letters to the right of the figure. Results are also shown for sensitivities of $\Delta T2\times$ = 1.5° and 4.5°C (mid-melt assumed). Climate sensitivity and aerosol forcing uncertainties are equally important, as shown by the overlapping ranges for different sensitivities. At the extreme high end (high sensitivity, low aerosol forcing, high melt), the rate of rise is almost 30 cm/century (26% due to unforced effects). At the extreme low end, the rate of rise is negligible (with zero unforced changes).

For the CE case (Fig. 4), CE from 2000, the central commitment estimate is a continuing rise of almost 25 cm/century. Approximately 15% of this is due to unforced effects. The lower and upper bounds are around 7 cm/century and more than 50 cm/century. These projections do not include the more catastrophic possibilities of accelerated melt in Greenland or the collapse of the West Antarctic ice sheet, as discussed in the TAR (*16*).

I considered the conventional (CC) commitment of changes that occur if atmospheric composition is held fixed at present (2000) levels, and the CE commitment for which emissions are fixed at their present levels. These commitments have been quantified for both global-mean temperature and sea level rise. Time-dependent changes are considered rather than just the usual asymptotic or equilibrium commitment.

The CC warming commitment rises steadily to an eventual warming of about 0.2° to more than 1°C. The contribution from past natural forcings exceeds that from past anthropogenic forcing. The corresponding CE warming commitment has no limit even on a time scale of many centuries, primarily because, at CE, CO_2 concentrations continue to rise for a millennium or more. The CE warming commitment in 2400 ranges from 2° to almost 6°C, with most of the commitment due to past anthropogenic forcing. Both climate sensitivity and past aerosol forcing uncertainties are important in determining the CC commitment, whereas climate sensitivity is the main source of uncertainty for the CE commitment.

For sea level rise, both the CC and CE commitments lead to almost linear increases in sea level out to at least 2400 and probably much longer. For the CC commitment, sea level rises at about 10 cm/century (uncertainty range, near zero to about 30 cm/century). Except at the low end of the range, a substantial fraction of this increase arises from unforced contributions to sea level rise (40% in the cen-

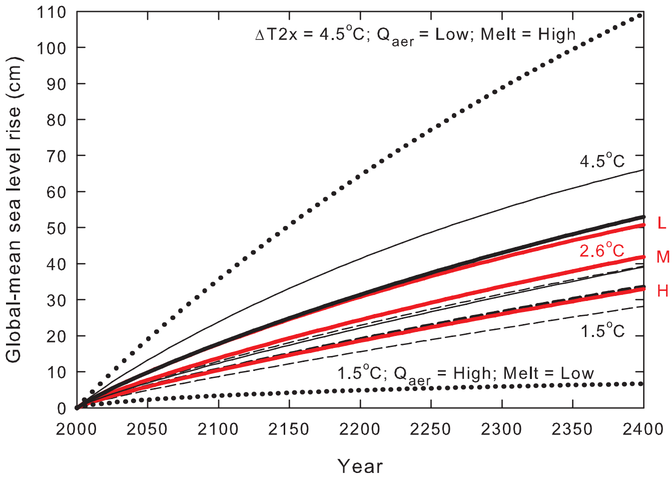

Fig. 3. CC sea level rise commitment (constant concentrations after 2000) for different climate sensitivities and aerosol forcing levels (L, M, and H on the right of figure indicate low, mid-, and high magnitudes for aerosol forcing, respectively). The central curves assume best-estimate values for all ice melt parameters. For these curves, aerosol forcing and climate sensitivity uncertainty ranges overlap. For example, the mid-aerosol-mid sensitivity ($\Delta T2\times$ = 2.6°C; central red curve) results are very similar to the high-aerosol-high sensitivity ($\Delta T2\times$ = 4.5°C; lowest full black curve) results and the low-aerosol-low sensitivity results ($\Delta T2\times$ = 1.5°C; top dashed black curve). Extremes spanning sensitivity, aerosol, and melt uncertainties are shown by the bottom and top dotted curves.

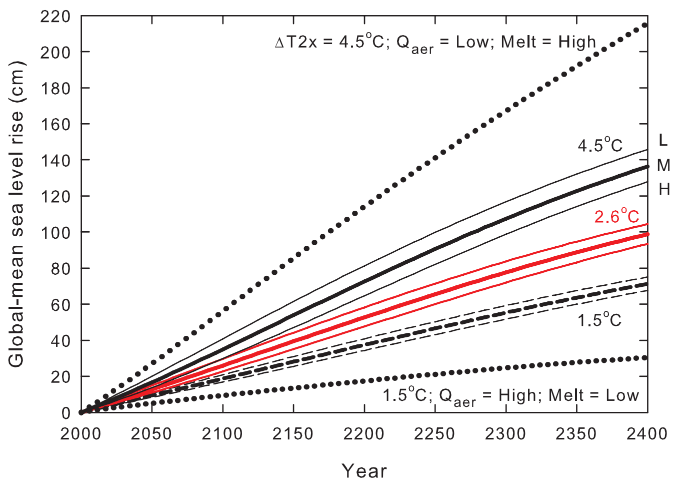

Fig. 4. CE sea level rise commitment (CE after 2000) for different climate sensitivities and aerosol forcing levels (L, M, and H on the right of the figure indicate low, mid-, and high magnitudes for aerosol forcing, respectively). The central curves assume best-estimate values for all ice melt parameters. Extremes spanning sensitivity, aerosol, and melt uncertainties are shown by the bottom and top dotted curves.

tral case). For the CE commitment, sea level rises at about 25 cm/century (uncertainty range, 7 to more than 50 cm/century). The fractions arising from unforced contributions to sea level rise are less than those in the CC case.

The CE results reinforce the common knowledge that, in order to stabilize global-mean temperatures, we eventually need to reduce emissions of greenhouse gases to well below present levels (*21*). The CC results are potentially more alarming, because they are based on a future scenario that is clearly impossible to achieve and so represent an extreme lower bound to climate change over the next few centuries. For temperature, they show that the inertia of the climate system alone will guarantee continued warming and that this warming may eventually exceed 1°C. For sea level, a continued rise of about 10 cm/century for many centuries is the best estimate. Although such a slow rate may allow many coastal communities to adapt, profound long-term impacts on low-lying island communities and on vulnerable ecosystems (such as coral reefs) seem inevitable.

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Materials and Methods

Tables S1 and S2

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How Much More Global Warming and Sea Level Rise?

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Two global coupled climate models show that even if the concentrations of greenhouse gases in the atmosphere had been stabilized in the year 2000, we are already committed to further global warming of about another half degree and an additional 320% sea level rise caused by thermal expansion by the end of the 21st century. Projected weakening of the meridional overturning circulation in the North Atlantic Ocean does not lead to a net cooling in Europe. At any given point in time, even if concentrations are stabilized, there is a commitment to future climate changes that will be greater than those we have already observed.

Increases of greenhouse gases (GHGs) in the atmosphere produce a positive radiative forcing of the climate system and a consequent warming of surface temperatures and rising sea level caused by thermal expansion of the warmer seawater, in addition to the contribution from melting glaciers and ice sheets (*1*, *2*). If concentrations of GHGs could be stabilized at some level, the thermal inertia of the climate system would still result in further increases in temperatures, and sea level would continue to rise (*2*–*9*). We performed multimember ensemble simulations with two global coupled three-dimensional climate models to quantify how much more global warming and sea level rise (from thermal expansion) we could experience under several different scenarios.

The Parallel Climate Model (PCM) has been used extensively for climate change experiments (*10*–*15*). This model has a relatively low climate sensitivity as compared to other models, with an equilibrium climate sensitivity of 2.1°C and a transient climate response (TCR) (the globally averaged surface air temperature change at the time of CO_2 doubling in a 1% CO_2 increase experiment) of 1.3°C. The former is indicative of likely atmospheric feedbacks in the model, and the latter includes ocean heat uptake and provides an indication of the transient response of the coupled climate system (*6*, *12*). A second global coupled climate model is the newly developed Community Climate System Model version 3 (CCSM3), with higher horizontal resolution (atmospheric gridpoints roughly every 1.4° as compared to the PCM, with gridpoints about every 2.8°) and improved parameterizations in all components of atmosphere, ocean, sea ice, and land surface (*16*). The CCSM3 has somewhat higher sensitivity, with an equilibrium climate sensitivity of 2.7°C and TCR of 1.5°C. Both models have about 1° ocean resolution (0.5° in the equatorial tropics), with dynamical sea ice and land surface schemes. These models were run for four- and eight-member ensembles for the PCM and CCSM3, respectively, for each scenario (except for five members for A2 in CCSM3).

The 20th-century simulations for both models include time-evolving changes in forcing from solar, volcanoes, GHGs, tropospheric and stratospheric ozone, and the direct effect of sulfate aerosols (*14*, *17*). Additionally, the CCSM3 includes black carbon distributions scaled by population over the 20th century, with those values scaled by sulfur dioxide emissions for the rest of the future climate simulations. The CCSM3 also uses a different solar forcing data set for the 20th century (*18*). These 20th-century forcing differences between CCSM3 and PCM are not thought to cause large differences in response in the climate change simulations beyond the year 2000.

The warming in both the PCM and CCSM3 is close to the observed value of about 0.6°C for the 20th century (*19*), with PCM warming 0.6°C and CCSM3 warming 0.7° (averaged over the period 1980–1999 in relation to 1890–1919). Sea level rises are 3 to 5 cm, respectively, over the 20th century as com-

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