Knowability and no ability in climate projections

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1. Introduction

The purpose of this note is to provide a referenced summary of the present scientific understanding about future climate change, tailored towards the kind of global climate factors that are captured in Integrated Assessment Models (IAMs). In outline, it is organized as follows:

i) *Equilibrium climate sensitivity* is the long-term response of global temperature to a doubling of atmospheric CO_2. I review the causes of our current uncertainty, and the prospects for reducing it.

ii) Two other measures of climate change are arguably more important in this context. First the *climate commitment* is a measure of the climate change we already face because of emissions that have already occurred.

iii) The very long timescales associated with attaining equilibrium, especially at the high end of possible climate sensitivity, mean that the *transient climate response* is of greater relevance for climate projections over the next several centuries.

iv) Due to the inherent uncertainties in the climate system, a *flexible emissions strategy* is far more effective in avoiding a given level of global temperature change, than a strategy aims to stabilize CO_2 at a particular level.

v) Many important climate impacts are fundamentally regional in nature. Among climate models, regional climate projections correlate only partially with global climate projections.

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2. Climate sensitivity

Climate sensitivity (here given the symbol T_{2x}, and sometimes called the equilibrium climate sensitivity) is the long-term change of annual-mean, global-mean, near-surface air temperature in response to a doubling of carbon dioxide above preindustrial values. It has long been a metric by which to compare different estimates of the climate response to greenhouse gas forcing (e.g., Charney, 1979). There is a vast literature that has researched climate sensitivity from every possible angle, ranging from state-of-the-art satellite observations of Earth's energy budget, to geological studies covering hundreds of millions of years. A fine review of where things stand can be found in Knutti and Hegerl (2008).

Figure 1 shows a variety of probability distributions (pdfs) of climate sensitivity. A prominent feature of such estimates is that they all exhibit considerable skewness. In other words, while the lower bound is confidently known, the upper bound is much more poorly constrained. There is a small but nontrivial possibility (about 25 %) that the climate sensitivity could exceed 4.5 °C. One concern that has been raised is that the current generation of IPCC climate models (from the fourth assessment, or AR4) does not span the range of climate sensitivity that is allowable by observations (the blue histogram in figure 1 clusters too narrowly around the modes of the other pdfs). The reason for this appears to be that the IPCC climate models do not sample the full range of possible aerosol forcing (Armor and Roe, 2010). This should not be surprising since they are designed to represent the "best" estimate of climate (something akin to the mode of the distribution). However, since these computer models are the only tools available for modeling regional climates, it should perhaps be a concern that they are under sampling the range of possible futures. I next outline briefly how estimates are made from observations and models. The purpose of doing so is to straightforwardly demonstrate the important sources of uncertainty.

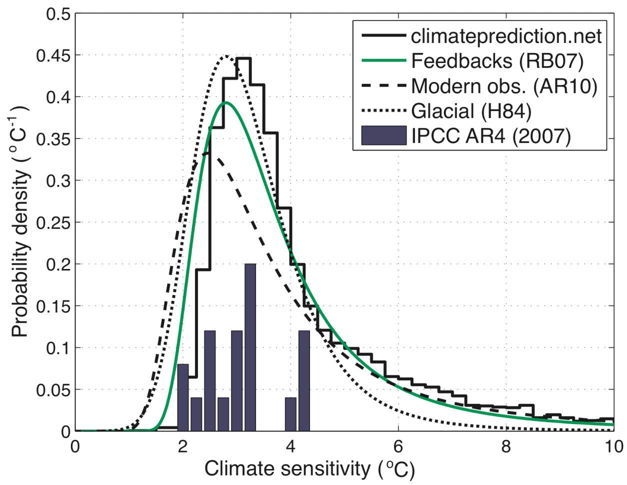

Figure 1. Various estimates of climate sensitivity. In order of the legend: i) from multi-thousand ensembles from one climate model (Stainforth et al., 2005), ii) from feedbacks with climate models (Roe and Baker, 2007), iii) from modern observations (Armour and Roe, 2010), iv) from glacial climates (Hansen et al., 1984), v) A histogram of T_{2x} from 19 main IPCC AR4 models (IPCC, 2007).

2a. Estimates of climate sensitivity from observations.

A linear approximation of the Earth's energy budget is

$$R = H + \lambda^{-1} T, \tag{1}$$

where R is the radiative forcing (units W m^{-2}), H is the heat going into the world's oceans and being stored there, and $\lambda^{-1}T$ is the climate response in terms of the global-mean, annual-mean, near-surface air temperature T, and the climate sensitivity parameter, λ. (e.g., Roe, 2009, Armour and Roe, 2010, and many others). For silly historical reasons the terminology here can be confusing. λ is a more fundamental measure of climate system than T_{2x}, since it does not depend on any particular forcing. λ and T_{2x} are related in the following way. Let R_{2x} be the radiative forcing due to a doubling of CO_2 over pre-industrial values ($\approx$ 4 W m^{-2}). In the long-term equilibrium, ocean heat uptake goes to zero, and so the climate sensitivity is just:

$$T_{2x} = \lambda R_{2x} \tag{2}$$

The point of this algebra is to make it clear that the goal of estimating climate sensitivity from observations is the goal of estimating λ from Equation (1):

$$\lambda = \frac{T}{R - H} \tag{3}$$

We have observations of T, R, and H, whose probability distributions are shown in figure 2. Hereafter we refer to R-H as the climate forcing, since it is the net energy imbalance that the atmosphere must deal with. H and T are actually quite well constrained, as is the radiative forcing associated with CO_2 and other greenhouse gases. As is clear from figure, the major source of uncertainty is R and, in particular, the component of R that is due to aerosols (small airborne particulates that can be either liquid or solid).

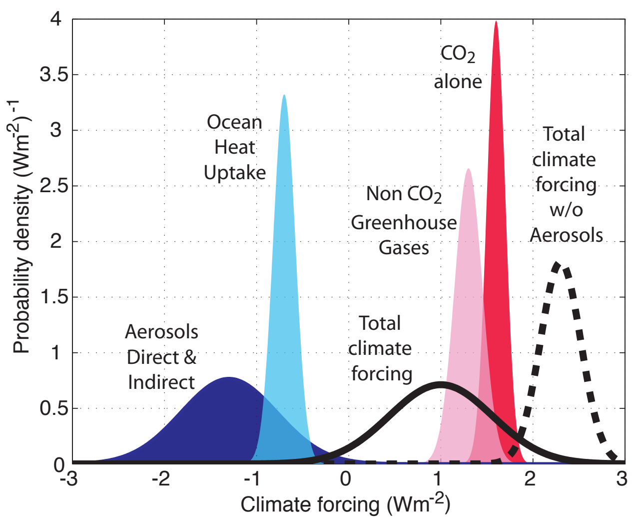

Figure 2: Probability distributions of the terms in the Earth's energy budget, based on IPCC 2007, and updated for newer ocean heat uptake observations. See Armour and Roe, 2010 for details. Total climate forcing is equal to R-H in Eq. 3. Also shown is the total forcing excluding aerosols, which is the climate forcing experienced by the Earth, if all anthropogenic emissions ceased immediately.

The reason that aerosol forcing is hard to constrain is that 1) the spatial pattern and lifetime is extremely complicated to observe (they are primarily in the Northern Hemisphere and downwind of major industrial economies); 2) some aerosols have a cooling effect, some have a warming effect; 3) aerosols alter the thickness, lifetime, and height of clouds – a powerful indirect effect that is hard to measure and attribute properly. The community is confident, however, that the net aerosol effect is almost certainly negative. More information about aerosol uncertainties can be found in Menon (2004).

Thus, from Eqs. 2 and 3, the probability distribution of climate sensitivity comes from combining a relatively narrow distribution (the well-known temperature change) in the numerator with a relatively broad distribution (the much less well-known climate forcing (i.e., R-H)) in the denominator of Eq. 3. It is this combination that produces the skewed distribution seen in figures 1 and 3c. The graphs in figure 3 are the fundamental reason why we can say with great confidence that it is very likely that observed forcing has not been large enough to imply a climate sensitivity of less than about 1.5 ^{o}C. On the other hand, uncertainties in observed forcing also mean that we cannot confidently rule out the disconcerting possibility that the modern warming has occurred with small climate forcing, which would imply very high climate sensitivity. Note that the curves in figure 1 and 3 are consistent with the probabilities given in the 2007 IPCC report.

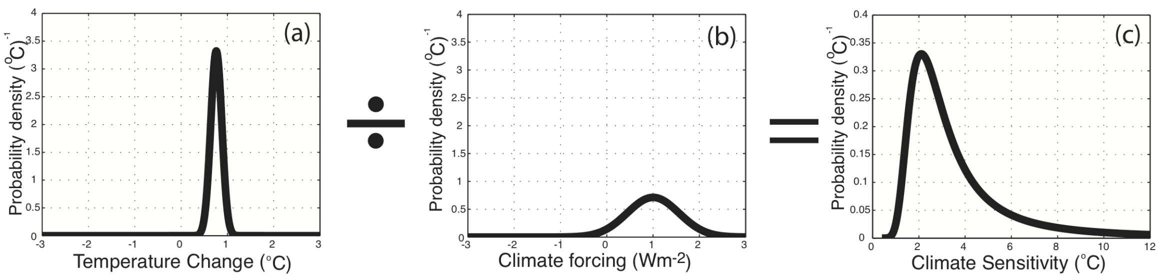

Figure 3: The calculation of climate sensitivity from observations involves combing a relatively narrow probability distribution of T (panel a) in the numerator, with a relatively broad distribution of F= H-R (panel b) in the denominator of Eq. (3). This leads to the skewed distribution of climate sensitivity (panel c). Note the pdfs must be combined properly - it is not just a simple division - but the point is hopefully clear.

2b. Estimates of climate sensitivity from models.

Climate sensitivity also can be estimated from climate models. Figure 1 shows three such efforts. The first is the spread of T_{2x} among the main IPCC AR4 models. One issue is that the mainstream IPCC AR4 climate models are not designed to explore the edges of the probability distribution, but instead are designed with the most likely combination of model parameters, and parameters are 'tuned' to reproduce observed climate history. Clear evidence of that tuning comes from the correlation of climate sensitivity and imposed aerosol forcing in the models in such a direction that twentieth century observations tend to be reproduced (Kiehl, 2007, Knutti, 2008). Such tuning is not problematic if models are interpreted as reflecting combinations of climate sensitivity and aerosol forcing that are consistent with observed constraints (Knutti, 2008). However AR4 models do not fully span the range of aerosol forcing allowed by observations (Kiehl, 2007; IPCC, 2007). This is the likely reason that the AR4 models under sample of the full range of possible climate sensitivity, as seen in figure 1.

Climate sensitivity can also be estimated by using thousands of integrations of the same climate model with the parameters varied by reasonable amounts, a strategy pursued by the *climateprediction.net* effort (figure 1, e.g., Stainforth et al., 2005). This work also found a skewed pdf of T_{2x}. Roe and Baker (2007) explain this in terms of a classic feedback analysis, summarized in figure 4. The relationship between feedbacks and response also produces a skewed distribution because of the way that positive feedbacks have a compounding effect on each other (e.g., Roe, 2009). The range of feedbacks as diagnosed within the AR4 models produces a pdf of climate sensitivity that is quite consistent with the pdf estimated from observations (figure 1). This should be expected since it is observations that ultimately provide constraints on the models.

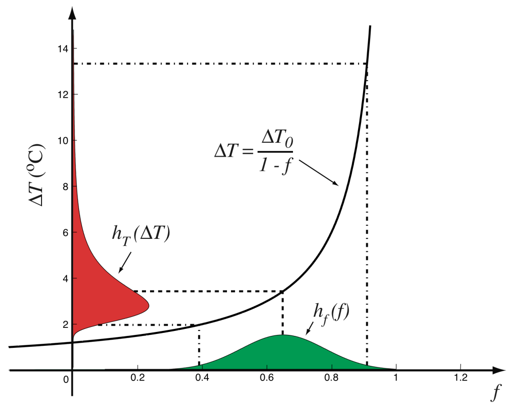

Figure 4: Model feedbacks and climate sensitivity. The black curve shows the mapping between climate feedbacks (x-axis, green curve), and climate response (y-axis, red curve). See Roe and Baker, 2007 for details.

2d. Prospects for improved estimates of climate sensitivity.

Can a narrower range of climate sensitivity be expected soon? One can ask: how might more accurate observations or better climate models change the estimate of T_{2x}?

Reducing uncertainty in either forcing or feedbacks would produce a narrower range. However it is the nature of these skewed distributions that the mode of T_{2x} moves to higher values as the range of forcing or feedbacks is narrowed, leaving the cumulative probability of $T_{2x} > 4.5^{o}C$ stubbornly persistent (Allen et al., 2007; Roe and Baker, 2007; Baker et al., 2010).

It should also be made clear that there are formidable scientific challenges in reducing uncertainty in climate model feedbacks, or in observing the aerosol forcing better. Progress will occur, but it is likely that it will be incremental.

Another line of attack is to try to combine multiple estimates of climate sensitivity in a Bayesian approach that might, in principal, significantly slim the fat tail of T_{2x} (e.g., Annan and Hargreaves, 2006). However, as with all Bayesian estimates, the value of the analysis is critically sensitive to 1) the independence of different observations; and 2) structural uncertainties within and among very complex models (e.g., Henriksson et al., 2010; Knutti et al., 2010). An objective assessment of these factors has proven elusive, rendering the information obtained by the exercise hard to interpret, and there is an acute risk that it produces overconfident estimates.

Overall it is probably prudent to anticipate that there will not be dramatic reductions in uncertainty about the upper bound on climate sensitivity (Knutti and Hegerl, 2008). On the timescale of several decades, Nature herself will slowly reveal more of the answer. We will learn about the transient climate response (see below) more quickly than the equilibrium climate sensitivity.

Those interested in understanding the above arguments in greater depth would do well to read the work of Prof. Reto Knutti (at ETH in Switzerland) and his collaborators. His research is of extremely high caliber, and quite accessible for a non-specialist.

3. The climate commitment.

What if all human influence on climate ceased overnight? Such a scenario—called the *climate commitment*—informs us of the climate change we already face due only to past greenhouse gas emissions. Framing the question this way has proven to be useful in providing a conceptual lower bound on future climate warming.

Early definitions of the climate commitment simply fixed CO_2 concentrations at current levels (e.g., Wigley, 2005; Meehl et al., 2005), but maintaining current levels actually requires continued emissions. Lately the focus has been more appropriately on the consequences of establishing zero emissions (e.g., Solomon et al., 2009). Two important, though sometimes overlooked points should be made. Firstly the geological carbon cycle means that, although much of the anthropogenic CO_2 ultimately gets absorbed by the ocean, some fraction — about 25 to 40% — remains in the atmosphere for hundreds of thousands of years (e.g., Archer et al., 2009). Secondly aerosols, have a short lifetime in the atmosphere (days to weeks). Thus when human influence ceases, aerosols are

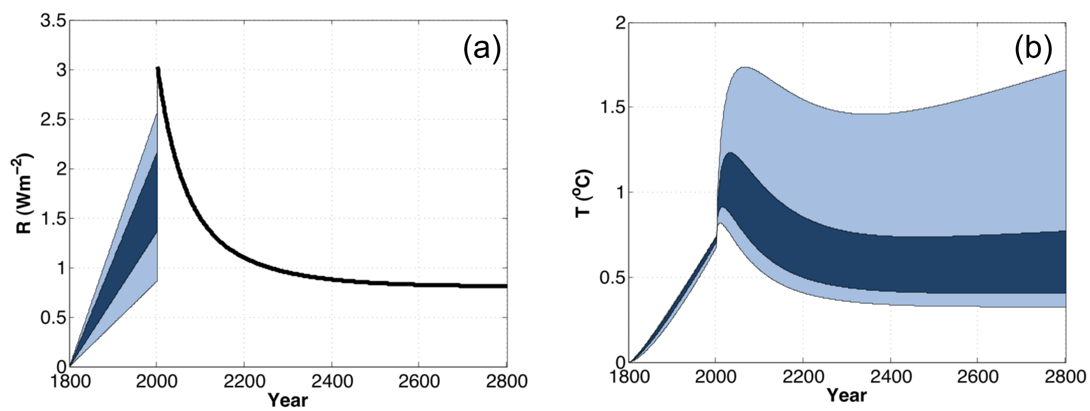

Figure 5: Idealized representation of the climate commitment following a cessation of all human influence on climate. Based on Armour and Roe, 2010. Panel (a) shows a simple view of how uncertainty in forcing has grown since 1800, as allowed by IPCC 2007 observed uncertainties. After emission cease (here at yr 2000) the uncertain aerosols quickly vanish, there is a jump in forcing due to sudden unmasking of the (relatively well-known) radiative forcing due to CO_2 and other greenhouse gases, which then declines slowly over time (black line). Panel (b) shows the temperature over this period, from a simple climate model. For each possible trajectory of past climate forcing history, a different value of climate sensitivity is implied, in order that the accurately known past warming is reproduced (low past forcing requires high climate sensitivity, and vice versa). The light blue curve shows the 90% confidence range, as permitted by uncertainties in observations, which ultimately grows to be 0.3 to 6^{o}C at equilibrium. The dark blue curve is the `likely' IPCC range (68%). It is this range that is spanned by the main IPCC AR4 models because they under sample the allowed range of past forcing. Note that these calculations here only include uncertainties due to aerosols. The spread would be larger if uncertainties in GHG and ocean heat uptake were included. Nonetheless the graph highlights that uncertainty in future temperatures is a result of uncertainty in past forcing.

rapidly washed out of the atmosphere and the effect of this is to unmask additional warming due to the much more slowly declining CO_2 (illustrated in figure 2 and 5).

Figure 5 shows an idealized calculation of the climate commitment from Armour and Roe (2010), which contains more details. The purpose of showing this is to highlight that our uncertainty about future temperature comes primarily from our uncertainty about past forcing. After ceasing all emissions, the degree and trajectory of future warming depends on the state of the current climate forcing. We face the disconcerting possibility that our ultimate climate commitment already exceeds 2 ^{o}C, because of our current inability to rule out that past warming occurred with relatively little climate forcing. In other words, the lower flank of the pdf of the past climate forcing distribution (figure 5a) controls the upper flank of the pdf of the future temperature response (figure 5b).

3a. Climate forcing and climate sensitivity are not independent.

Perhaps the most important point to emphasize for the application to integrated assessment models (IAMs) is that climate sensitivity and climate forcing are not independent of each other. For any projections made of the future, a starting point for the current climate forcing must be assumed. We are currently quite uncertain about what that starting point is. If aerosol forcing is strongly negative, there is a strong implication that climate sensitivity is high. If aerosol forcing is weak, climate sensitivity must be low. Uncertainties in climate forcing and climate sensitivity must not be assumed to be independent.

4. The transient climate response.

Equilibrium climate sensitivity relates to a hypothetical distant future climate after the system has equilibrated to a stipulated forcing. The transient climate response over the course of a few centuries may be a more directly useful property of the climate system. A formal definition of the transient climate sensitivity has been proposed as the global-average surface air temperature, averaged over the 20-year period centered on the time of CO_2 doubling in a 1% yr^{-1} increase experiment, which occurs roughly at 2070. While this metric may be more relevant for the future, a negative trade-off is that its exact value depends on this artificially defined trajectory of emissions.

For reasons discussed below, the transient climate response is much better constrained than climate sensitivity. In the words of the IPCC, it is very likely (> 9-in-10) to be greater than 1°C and very unlikely (< 1-in-10) to be greater than 3 °C. Thus the community is much more confident about the evolution of the climate over the coming century than it is about the ultimate warming.

4a. The immensely long timescales of high sensitivity climates.

A key factor in the long-term evolution of the climate is the diffusive nature of the ocean heat storage (figure 6b). In order to reach equilibrium the ocean abyss must also warm, and because of the relatively sluggish circulation of the deep ocean, the upper layers must be warmed before the lower layers, and the more the temperature change must be, the longer diffusion takes to work. A simple scaling analysis (e.g., Hansen et al., 1985) shows that:

$$\text{Climate adjustment time} \propto (\text{climate sensitivity})^2$$

Thus if it takes 50 yrs to equilibrate with a climate sensitivity of 1.5 °C, it would take 100 times longer, or 5,000 yrs to equilibrate if the climate sensitivity is 15 °C. Although Nature is of course more complicated than this, the basic picture is reproduced in models with an (albeit simplified) ocean circulation. Figure 6a shows one such calculation from Baker and Roe (2009), though there are others (in particular see Held et al., 2010).

If IAMs are to be used to project out more than a few decades, it is critical that they represent this physics correctly. A single adjustment time for climate, or a deep ocean that is represented as a uniform block, cannot represent this behavior.

The extremely high temperatures found in the fat tail of climate sensitivity cannot be reached for many centuries for very robust physical reasons. Failure to incorporate this fact will lead to a strong distortion of the evolution of possible climate states, and of the subsequent IAM analyses based on them.

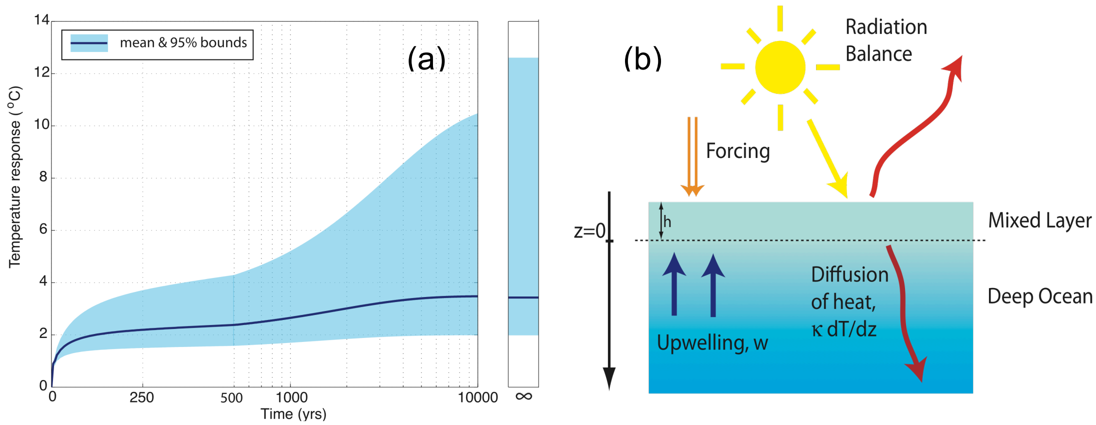

Figure 6: (a) The evolution of possible climate trajectories in response to an instantaneous doubling of CO_2 given the existing uncertainty in climate sensitivity. From Baker and Roe, 2009. Note the change to a logarithmic x-axis after 500 years. Low climate sensitivity is associated with rapid adjustment times (decades to a century). High climate sensitivity has extremely long adjustment times – thousand of years. This results from the fundamentally diffusive nature of the ocean heat uptake, illustrated schematically in panel (b). Such behavior is also reproduced in more complete physical models. See Held et al. (2010), for example.

5. CO_2 stabilization targets are a mistake.

A prominent part of the conversation about action on climate change has centered on what the right level of CO_2 should be in the atmosphere (e.g., Solomon et al., 2010). Some advocate for 350 ppmv (e.g., Hansen et al. 2008), though we are already past 380 ppmv and climbing, others contemplate the consequences of 450 ppmv (e.g., Hansen, et al., 2007), still others 550 ppmv (Pacala and Soccolov, 2004; Stern, 2007).

However decreeing and setting in stone a particular target for CO_2 is fundamentally the wrong approach, and a vastly inefficient way to avoid a particular climate scenario. This point was made very elegantly and powerfully in a study by Allen and Frame (2007), reproduced in figure 7. Panel a) shows a scenario of what could happen if we decided today to stabilize CO_2 at 450 ppmv by 2100, and then waited for the climate to evolve. Our current best guess is that would lead to an equilibrium temperature change of 2 ^{o}C, taking us to the edge of what some have called dangerous climate change. However because of our current uncertainty in climate sensitivity, the envelope of possible climate states is quite broad by 2150. In other words, our hypothetical choice that we made today still leaves us exposed to a quite broad envelope of risk. Note, though, that figure 7a is consistent with figure 6 – temperatures in the fat tail of high climate sensitivity are still very, very far from equilibrium at 2150.

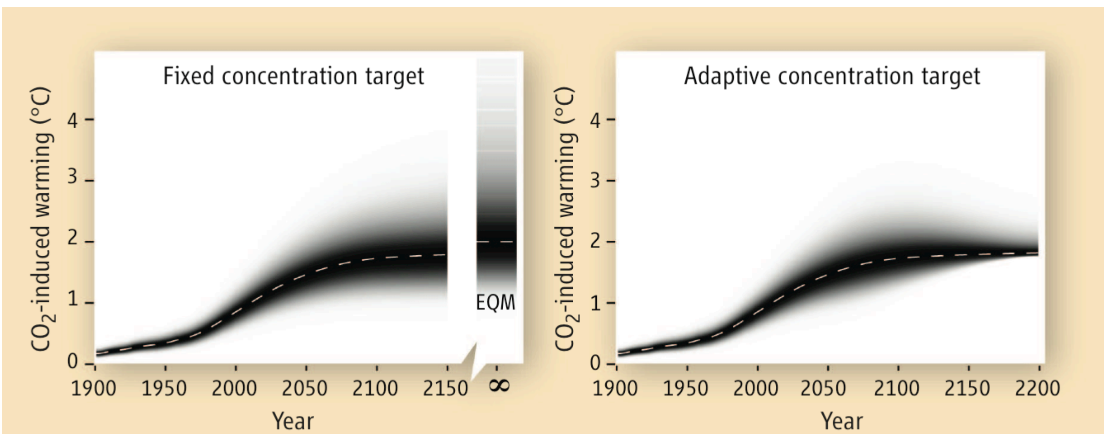

Figure 7: reproduced from Allen and Frame (2007). Carbon dioxide–induced warming under two scenarios simulated by an ensemble of simple climate models. (Left) CO_2 levels are stabilized in 2100 at 450 ppm; (right) the stabilization target is recomputed in 2050. Shading denotes the likelihood of a particular simulation based on goodness-of-fit to observations of recent surface and subsurface-ocean temperature trends. The darker the shading, the likelier the outcome.

Panel b) of figure 7 considers an alternative strategy in which we still act according to our best guess today, but re-compute a new concentration target at 2050, based on the fact that 40 years have elapsed and Nature has given us more information about what trajectory we are on. Figure 7b makes it clear that this adaptive strategy is vastly more effective in achieving a desired climate target (in this case a global temperature change of 2 ^{o}C).

Because the link between CO_2 levels and global temperature is uncertain, and because it is prudent to anticipate only incremental advances in our understanding, it is common sense to pursue a strategy that has built-in flexibility rather than declaring a fixed concentration.

6. How well do global projections correspond to regional projections?

Many of the most important climate impacts – changes in hydrology, storminess, heat waves, snowpack, etc. – are fundamentally regional in nature. How reliable is global climate change as a predictor of regional climate change? Since this is a question about the future, we are forced to use climate models. Figure 8 analyzes how well global climate sensitivity correlates with local climate change (in this case annual mean temperature and precipitation change in 2100), comparing among eighteen different IPCC models (IPCC, 2007).

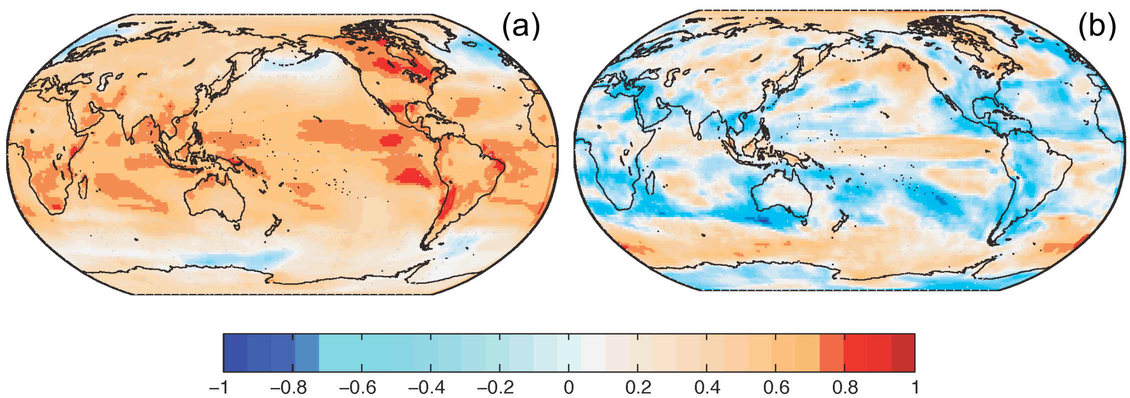

Figure 8: a) correlation among 17 IPCC climate models of their global equilibrium climate sensitivity and their local annual-mean temperature change in 2100,; b) same as a), but for annual-mean precipitation. Calculation made by N. Feldl from IPCC archived model output based on the A1B emissions scenario, and similar plots for other variables are at *http://earthweb.ess.washington.edu/roe/GerardWeb/Publications.html.*

It takes a correlation of $r \sim 0.75$ before half of the variance (i.e., r^2) of the local climate change is attributable to the global climate change. Only a very few patches of the planet achieve even this level of correlation in annual temperature (Figure 8a) and nowhere reaches this measure in annual precipitation (Figure 8b). This highlights that the connection between regional and global climate change is not that strong. This result should not be surprising: though models may all agree on the *sign* of the climate change in a given region, there is a great deal of scatter and individual model vagaries in projecting the *magnitude* of the climate change. Research into the limits of regional predictability is only just beginning. A useful starting point is Hawkins and Sutton (2009).

Summary.

1) The most important point to drive home is that uncertainty is not ignorance. The planet has warmed in the recent past, and will continue to warm for the foreseeable future. That this is a result of our actions is beyond rational dispute. The overwhelming preponderance of the IPCC 2007 report is extremely reliable,

and reflects an objective characterization of the best current understanding about climate. All of the following points are consistent with (and in many cases drawn from) that report.

2) A traditional measure of the planet's response, equilibrium climate sensitivity is uncertain, primarily because of uncertainty in the radiative forcing due to aerosols. This precludes us from calibrating our models of climate with greater accuracy.

3) However a focus on climate sensitivity may be misplaced because of the tremendously long timescales associated with reaching equilibrium – thousands of years in the case of the fat tail of high climate sensitivity.

4) If all human influence were to cease today, the rapid loss of anthropogenic aerosols from the climate would unmask CO_2 warming, and the planet's temperature would increase as a result. The degree of warming is quite uncertain.

5) For related reasons, a strategy that aims to stabilize concentration of greenhouse gasses at a particular level is a mistake, because the degree of warming is still unpredictable. A strategy that aims for a flexible emissions will be much more effective at preventing a particular level of warming.

6) IAMs have to make choices about how to represent climate forcing associated with human activity. We are quite uncertain about what this level is right now. It is crucial to appreciate that uncertainty in climate sensitivity and uncertainty in climate forcing cannot be treated as independent.

7) Many climate damages both to humans and to the biosphere result from regional climate factors. Unfortunately, there is relatively little agreement among climate models about how global climate changes relate to local climate changes, and this is especially true in some of the most vulnerable subtropical regions. Thus the meaning of analyses that use only global temperature changes to assign climate damages is unclear.

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