CLIMATE SENSITIVITY: UNCERTAINTY AND LEARNING

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Abstract
The importance of human-induced climate change depends critically on the temperature sensitivity of the climate system, measured by the change in global-average near-surface temperature resulting from a doubling of the preindustrial carbon dioxide concentration, denoted by ΔT_{2x}. Estimates of the value of ΔT_{2x} have been based on mathematical climate models, instrumental measurements of temperature since about the mid 19th century, and surrogate indicators of temperature prior to the instrumental record. Based predominantly on climate model simulations, the Intergovernmental Panel on Climate Change has stated that 1.5°C ≤ ΔT_{2x} ≤ 4.5°C. However, recent studies based on the instrumental temperature record find that there is a significant probability that ΔT_{2x} lies outside this range. Progress in reducing the uncertainty in the value of ΔT_{2x} will require reducing the uncertainty in the radiative forcing, not only by aerosols, but also by the Sun and volcanoes. It will also require a longer record of instrumentally observed near-surface temperatures to enhance the signal of forced climate change relative to natural climate variations.

1. Introduction

The importance of human-induced climate change depends critically on the temperature sensitivity of the climate system, measured by the change in global-average near-surface temperature resulting from a doubling of the pre-industrial carbon dioxide (CO_2) concentration, denoted by ΔT_{2x}. If ΔT_{2x} is small, then the problem of human-induced climate change may not be acute. If ΔT_{2x} is large, then human-induced climate change may be one of the most severe problems of the 21st century.

The earliest estimate of ΔT_{2x} was made by Arrhenius [1896] using an energy-balance model [*Schlesinger et al.,* 1997] which yielded ΔT_{2x} = 5.4°C. Subsequent estimates by such models, radiative-convective models and general circulation models [*Schlesinger et al.,* 1997] gave estimates respectively of 0.24°C [*Newell and Dopplick,* 1979] to 9.6°C [*Möller,* 1963], 0.48°C [*Somerville and Remer,* 1984] to 4.2°C [*Wang and Stone,* 1980], and 1.3°C [*Washington and Meehl,* 1983] to 5.2°C [*Wilson and Mitchell,* 1987]. Based on studies with general circulation models, a U.S. National Research Council (NRC) study chaired by Jule Charney wrote: "We estimate the most probable global warming for a doubling of CO_2 to be near 3°C with a probable error of ±1.5°C" [*NAS,* 1979]. A subsequent NRC study chaired by Joseph Smagorinsky concluded that: "no substantial revision of this {Charney report} conclusion is warranted at this time" [*NAS,* 1982]. The Intergovernmental Panel on Climate Change (IPCC) interpreted the findings of the Charney report to mean that 1.5°C ≤ ΔT_{2x} ≤ 4.5°C [*Houghton et al.,* 1990*.,* 1995*.,*

1996, 2001][1] Estimates of ΔT_{2x} based on paleoclimatic and instrumental temperature data range respectively from 1.3°C [*Hoffert and Covey,* 1992] to 6°C [*Barron,* 1994] and from 0.7 to 10.0°C [*Schlesinger and Ramankutty,* 1992]. None of these estimates provided probability density functions (pdf's) for ΔT_{2x}. An expert elicitation [*Morgan and Keith,* 1995] did provide subjective pdf's for 16 experts whose 90% confidence intervals ranged from (0.1°C to 0.5°C) to (0.1°C to 8°C). More recently, subjective estimates of the ΔT_{2x} pdf were obtained from the instrumental temperature record using Bayesian updating [*Tol and de Vos,* 1998], with the result that the posterior pdf depended strongly on the assumed prior (initial) pdf. The most recent studies based on the instrumental temperature record have found that there is a significant likelihood that ΔT_{2x} lies outside the $1.5°C \le \Delta T_{2x} \le 4.5°C$ range [*Andronova and Schlesinger,* 2001; *Forest et al.,* 2002; *Gregory et al.,* 2002].

In this paper we describe the method we have used to estimate the pdf of ΔT_{2x} based on the instrumental temperature record (section 2), and we examine the uncertainty in ΔT_{2x} due to uncertainties in the radiative forcing (section 3) and the observed temperatures (section 4). In section 5 we examine learning ΔT_{2x} over time. Our conclusions are given in section 6.

2. Method Used to Estimate ΔT_{2x}

We use a simple climate model (SCM) to simulate the change in hemispheric-mean temperatures from 1765 to the present for prescribed radiative forcing, ΔT_{2x}, and the radiative forcing by sulfate aerosol in a reference year (1990), ΔF_{SO4}.

2.1. Simple Climate Model

The original, global version of the SCM was developed by Schlesinger in 1984, based on the model's original formulation by Hoffert *et al.* [1980], and was used by Schlesinger and colleagues to simulate the global-mean temperature evolution for the different GHG scenarios of the IPCC 1990 report [*Bretherton et al.,* 1990], and for greenhouse-policy studies [*Schlesinger and Jiang,* 1991; *Schlesinger,* 1993; *Hammitt et al.,* 1992; *Lempert and Schlesinger,* 2000; *Lempert et al.,* 1994, 1996]. The hemispheric version of the model was developed to study the influence on the climate system of anthropogenic sulfate aerosol [*Schlesinger et al.,* 1992] and putative solar-irradiance variations [*Schlesinger and Ramankutty,* 1992], and has been used to discover a 65-70 year oscillation in the observed global-mean surface temperatures, but which was found to occur only over the North Atlantic Ocean and its bordering continental regions [*Schlesinger and Ramankutty,* 1994, 1995]. A hemispheric version of the model that explicitly calculates the individual temperature changes over land and ocean in each hemisphere [*Ramankutty,* 1994] has been used to investigate the influence on climate of volcanoes [*Ramankutty,* 1994] and the influence of global warming on sea level [*Yohe and Schlesinger,* 1998]. The hemispheric version of the model has been used to determine the causes of the

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[1] Strictly speaking, this interpretation is not correct. According to Webster's Revised Unabridged Dictionary "probable error (of an observation, or of the mean of a number), that within which, taken positively and negatively, there is an even chance that the real error shall lie. Thus, if 3{sec} is the probable error in a given case, the chances that the real error is greater than 3{sec} are equal to the chances that it is less. The probable error is computed from the observations made, and is used to express their degree of accuracy" [*Porter,* 1998]. Thus 3.0±1.5°C means that there is a 50% probability that $1.5°C \le \Delta T_{2x} \le 4.5°C$, and there is a 50% probability that ΔT_{2x} lies outside this range.

temperature changes observed since 1856 [*Andronova and Schlesinger,* 2000] and to estimate the pdf of ΔT_{2x} [*Andronova and Schlesinger,* 2001].

For prescribed ΔT_{2x} the SCM determines the changes in the temperatures of the surface air and ocean, the latter as a function of depth from the surface to the ocean floor [*Schlesinger et al.,* 1997]. The ocean in the model is subdivided vertically into 40 layers, with the uppermost being the 53 m deep mixed layer and the deeper layers each being 100 m thick. Also, the ocean is subdivided horizontally into a polar region where bottom water is formed, and a nonpolar region where there is upwelling. In the nonpolar region, heat is transported upwards toward the surface by the water upwelling there and downwards by physical processes whose effects are treated as an equivalent diffusion. Heat is also removed from the mixed layer in the nonpolar region by a transport to the polar region and downwelling toward the bottom, this heat being ultimately transported upward from the ocean floor in the nonpolar region. The atmosphere in each hemisphere is subdivided into the atmosphere over the ocean and the atmosphere over land, with heat exchange between them.

2.2. Radiative Forcing

The anthropogenic radiative forcing consists of greenhouse-gas (GHG) forcing beginning in 1765 [*Harvey et al.,* 1997] due to the increasing concentrations of CO_2, methane, N_2O, chlorofluorocarbons and tropospheric ozone [*Stevenson et al.,* 1998], and the direct (clear air) plus indirect (cloudy air) radiative forcing by tropospheric sulfate aerosols (SO_4) beginning in 1857 [*Harvey et al.,* 1997]. Stratospheric-ozone forcing due to ozone depletion is ignored here as it is small [*Forster,* 1999].

Volcanic radiative forcing estimated by *Andronova et al.* [1999] is predominantly due to the scattering of incident solar radiation back to space by stratospheric sulfate aerosols created from SO_2 gas injected into the stratosphere by major volcanic eruptions, for which we have optical-depth data beginning in 1850 [*Sato et al.,* 1993].

Solar radiative forcing larger than the 0.1% variation of the solar irradiance observed by satellites since 1978 over two 11-year sunspot cycles is hypothesized to have occurred before 1978 based on the observed variations of other characteristics of the sun and sun-like stars. Two models of solar forcing are examined here. One was constructed by *Lean et al.* [1995] for the period 1610-1994 based on sunspot areas and locations, He 1083 nm emission, group sunspot numbers, and Ca emissions from the sun and sunlike stars. This model was updated through 1998 by *Fröhlich and Lean* [1998]. The other model was constructed by *Hoyt and Schatten* [1993] for 1700-1992 based on the fraction of sunspot area occupied by the penumbra, solar-cycle length, equatorial rotation rate, decay rate of the solar cycle, and the mean level of solar activity. We shifted the irradiances of the HS solar model by 4 W/m^2 such that they matched the 1978 solar irradiance observed by satellite [*Fröhlich and Lean,* 1998], and we extended the irradiances of the HS solar model from 1992 through 1998 using the data from *Fröhlich and Lean* [1998].

2.3. Optimal Estimation of ΔT_{2x} and ΔF_{SO4}

Because the observed temperatures are departures from the 1961-90 mean while the simulated temperatures are the change from 1765, we add to the latter a constant to convert the temperature changes to temperature departures. The constant is determined analytically by minimizing the root-mean-square error (RMSE) between the simulated and observed global-mean temperature (GMT) departures. This is repeated for many different pairs of ΔT_{2x} and ΔF_{SO4}, and the locus

of points $\Delta T_{2x} = f_{GMT}(\Delta F_{SO4})$ for the minimum RMSE is determined. For this function, ΔT_{2x} increases as ΔF_{SO4} becomes more negative. This occurs because the net positive forcing due to the GHGs and SO_4 decreases as ΔF_{SO4} becomes more negative, hence ΔT_{2x} must increase to reproduce the observed GMT departures.

Similarly, a constant is added to the simulated changes in interhemispheric temperature difference (ITD) to convert them to temperature departures. We use the ITD because the historical emission of SO_2, the gaseous precursor to the SO_4 aerosol, has been larger in the northern hemisphere (NH) than in the southern hemisphere (SH). Thus the SO_4-induced cooling is larger in the NH than in the SH. The constant is determined analytically by minimizing the RMSE between the simulated and observed ITD departures. This is repeated for many different pairs of ΔT_{2x} and ΔF_{SO4} values, and the locus of points $\Delta T_{2x} = f_{ITD}(\Delta F_{SO4})$ for the minimum RMSE is determined. For this function, ΔT_{2x} decreases as ΔF_{SO4} becomes more negative. This occurs because the ITD increases in magnitude as ΔF_{SO4} becomes more negative, hence ΔT_{2x} must decrease to reproduce the observed ITD departures.

The $(\Delta T_{2x}, \Delta F_{SO4})$ solution is given by the intersection of the two functions, $\Delta T_{2x} = f_{GMT}(\Delta F_{SO4})$ and $\Delta T_{2x} = f_{ITD}(\Delta F_{SO4})$. In practice, this optimal-estimation solution is determined by an optimization program. As an example, for the radiative-forcing model (RFM) consisting of greenhouse gases (G), tropospheric ozone (T), sulfate aerosol (A) and the *Lean et al.* [1995] solar forcing (S), we obtain $\Delta T_{2x} = 2.7°\mathrm{C}$ and $\Delta F_{SO4} = -1.1\ \mathrm{Wm}^{-2}$. The differences between the observed and simulated GMTs and ITDs represent either the unforced natural variability of the climate system, the variability due to some radiative forcing not included in the GTAS radiative-forcing model, or both. In 1994 we analyzed a similar GMT residual using singular spectrum analysis and discovered a 65-70 year oscillation therein, but which was found to occur only over the North Atlantic Ocean and its bordering continental regions [*Schlesinger and Ramankutty,* 1994, 1995]. This finding has been given support by analyses of paleoclimate data (tree rings, ice cores, ice melt, lake varves, coral, historical data) [*Mahasenan et al.,* 1997; *Mann et al.,* 1995, 1999] and by simulation studies with coupled atmosphere-ocean models [*Delworth et al.,* 1993, 1997; *Delworth and Knutson,* 2000]. This oscillation is the cause of the observed global warming during the first half of the twentieth century and its subsequent reversal until the mid-1970s [*Andronova and Schlesinger,* 2000].

2.4. Probability Density Functions for ΔT_{2x} and ΔF_{SO4}

The GMT and ITD differences are only a single realization of the unforced temperature noise of the climate system. To measure the "noise-induced" uncertainty in our estimate of ΔT_{2x} and ΔF_{SO4}, we have used the bootstrap re-sampling method for correlated data developed by *Solow* [1985] to generate 5000 realizations of the noise, each realization of n = 142 years duration [*Andronova and Schlesinger,* 2001]. The bootstrap method is based on the same principle as Monte Carlo simulation, with the only difference being that the bootstrap uses the empirical distribution of the sample itself from which to resample. Each bootstrap sample is chosen by sampling n values at random with replacement from the original observations. The basic idea of the bootstrap re-sampling method for correlated data involves transforming the original correlated data to uncorrelated quantities, forming a bootstrap sample from these quantities, and then transforming back to a quasi bootstrap sample from the original data without destroying the pattern of autocorrelation exhibited by the original sample.

For any RFM such as GTAS, we construct an ensemble of 5000 surrogate observational temperature records for each hemisphere by adding the temperature signal for the RFM to each of the 5000 realizations of the noise. For each ensemble member we used the same procedure to estimate ΔT_{2x} and ΔF_{SO4} that we did for the single real observational record. We find that over the 16 RFMs (G, GA, GS, GT, GV, GAS, GAV, GSV, GTA, GTS, GTV, GASV, GTAS, GTAV, GTSV, GTASV) which contain 80,000 surrogate observational temperature records that the mean and 90% confidence range for ΔT_{2x} are 3.40°C and 1.0°C to 9.3°C, and there is a 54% likelihood that ΔT_{2x} lies outside the IPCC range of 1.5 to 4.5°C. The corresponding quantities for ΔF_{SO4} are –0.91 Wm^{-2} and –0.54 Wm^{-2} to –1.30 Wm^{-2}.

3. Uncertainty in ΔT_{2x} Due to Uncertainty in Radiative Forcing

Figure 1 illustrates the uncertainty in the optimal estimate of ΔT_{2x} due to uncertainty in radiative forcing. For GT, the radiative forcing is due greenhouse gases and tropospheric ozone, and $\Delta T_{2x} = 1.14°C$. This value is very close to the value obtained when there is no net feedback, $\left(\Delta T_{2x}\right)_o = G_o \Delta F_{2x}$, where $G_o = T/\left(1-\alpha_p\right)S_o$ is the gain of the climate system with zero feedback [*Schlesinger,* 1985*,* 1988*,* 1989][2]. Taking the global-mean temperature $T = 288$ K, planetary albedo $\alpha_p = 0.3$, and solar irradiance $S_o = 1367\ Wm^{-2}$ yields $G_o = 0.3\ °C/Wm^{-2}$. For $\Delta F_{2x} = 3.71\ Wm^{-2}$, $\left(\Delta T_{2x}\right)_o = 1.12°C$.

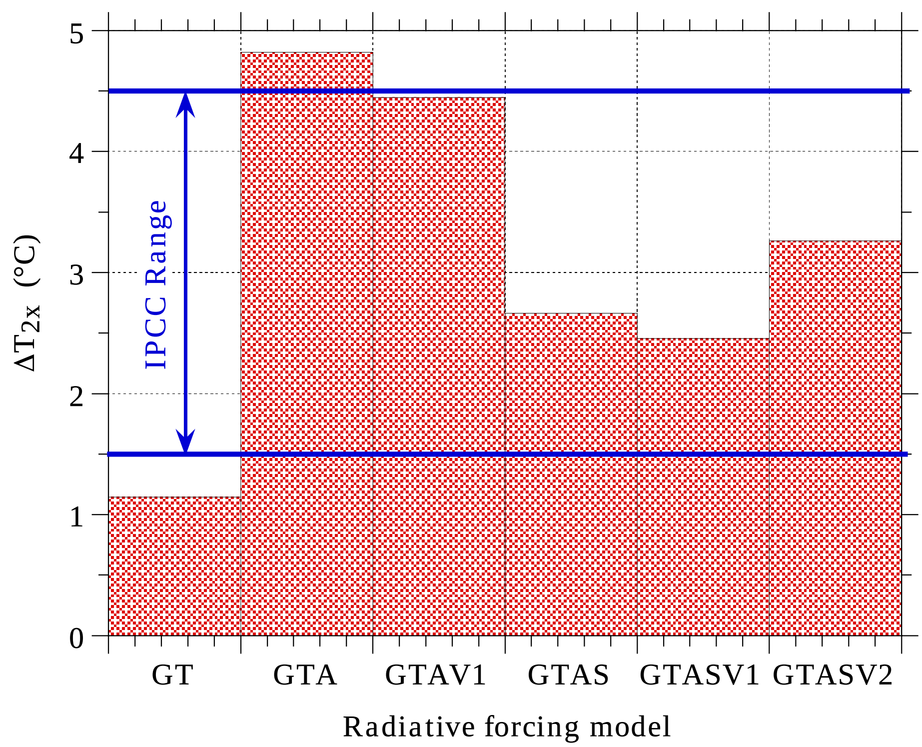

Figure 1. Dependence of the estimated ΔT_{2x} on radiative forcing model.

[2] Strictly speaking, the climate sensitivity is the gain of the climate system with feedback, $G_f = G_o/\left(1-f\right)$, where f is the feedback (*op cit.).*

For GTA, negative sulfate aerosol radiative forcing is added to the positive GT forcing and ΔF_{SO4} is determined by the optimum estimation together with ΔT_{2x}. For GTA, $\Delta T_{2x} = 4.8°C$. This fourfold increase in ΔT_{2x} is required such that the observed GMT can be reproduced by the SCM for the smaller net radiative forcing that results from the partial cancellation of the positive GT forcing by the negative sulfate forcing. In this light, the result for GT may be interpreted as the case for which the positive radiative forcing by carbonaceous aerosol balances the negative GT forcing by the sulfate aerosol forcing.

Including the radiative forcing calculated by *Andronova et al.* [1999] for the volcanic optical depths of *Sato et al.* [1993], GTAV1 in Fig. 1, reduces ΔT_{2x} by about 8%.

Including the solar-irradiance radiative forcing of *Lean et al.* [1998], GTAS in Fig 1, reduces ΔT_{2x} from its value for GTA by about 40%. A similar reduction is also obtained for the solar forcing of *Hoyt and Schatten* [1993] (ΔT_{2x} not shown). Because the solar-irradiance forcing constructed by both *Lean et al.* and *Hoyt and Schatten* increases over the period of instrumental temperature observations, the solar-irradiance forcing is positive. Adding it to the GTA forcing increases the net positive forcing. Accordingly, to reproduce the observed temperatures by the SCM requires the reduction of ΔT_{2x}. It is extremely important to learn whether or not the sun's irradiance varied as has been constructed. If it did not and changed only by the 0.1% observed by satellite since 1978 over a little more than two 11-year solar-activity cycles, then ΔT_{2x} is twice as large (GTA) as it would be if the sun did vary as constructed (GTAS).

Including volcanoes with GTAS decreases ΔT_{2x} by about 8%, as before

We have recently calculated the radiative forcing for the volcanic optical depths compiled by *Robertson et al.* [2001] and estimated their effect on ΔT_{2x} [*Andronova and Schlesinger,* 2003]. These optical depths differ from those of *Sato et al.* [1993] in both their chronology and intensity. When their radiative forcing is included with GTAS as shown by GTASV2 in Fig.1, ΔT_{2x} is increased by about 22%. This is in contrast to the 8% decrease in ΔT_{2x} when the *Sato et al.* [1993] volcanoes were included with GTA. Because volcanoes occur in only one hemisphere or the other, their radiative forcing is not the same in both hemispheres. Thus volcanoes influence not only the GMT but also the ITD. Accordingly, volcanoes can either decrease or increase ΔT_{2x}.

From these results it is clear that reduction of the uncertainty in the estimation of ΔT_{2x} requires reduction of the uncertainty in the radiative forcing by aerosols, the sun and volcanoes.

Table 1. Estimates of ΔT_{2x} and ΔF_{SO4} from three datasets for GA.

Dataset/Quantity	Period	ΔT_{2x} (°C)	ΔF_{SO4} (Wm^{-2})
Jones et al. [1999]	1861 – 1997	4.8564	–0.2183
Jones and Moberg [2003]	1861 – 2000	4.2437	–0.2068
Folland et al. [2001a]	1861 – 2000	1.2771	+0.0095

4. Uncertainty in ΔT_{2x} Due to Uncertainty in the Observed Temperatures

All of the results presented so far have been obtained using the hemispheric temperatures of *Jones et al.* [1999] which extend from 1856 through 1997. At the end of 2002 we decided to extend our analyses through 2000. At that time the only hemispheric temperature analysis that extended through 2000 was that of *Folland et al.* [2001a] which covers the period 1861 to 2000. Subsequently the analysis by *Jones and Moberg* [2003] from 1856 to 2000 has been published.

Table 1 shows our estimate of ΔT_{2x} and ΔF_{SO4} based on these three datasets. While the values obtained using the data of *Jones et al.* [1999] and *Jones and Moberg* [2003] (JAM) are comparable, the values obtained using the data of *Folland et al.* [2001] (FEA) are remarkably different. In particular, ΔF_{SO4} is positive for the *FEA* data, while ΔF_{SO4} is negative for the data of *Jones et al.* [1999] and *JAM.* As a result of this, ΔT_{2x} for the *FEA* data is only 30% of the value for the *JAM* data.

Figure 2 shows that the difference in the sign of ΔF_{SO4} for the *FEA* and *JAM* data is due to the difference in their temperatures for the Southern Hemisphere, with the *FEA* values being less negative than the *JAM* values during the late 19th and early 20th centuries. This difference leads to a positive ITD for the *FEA* data, which yields a positive ΔF_{SO4} and small ΔT_{2x}.

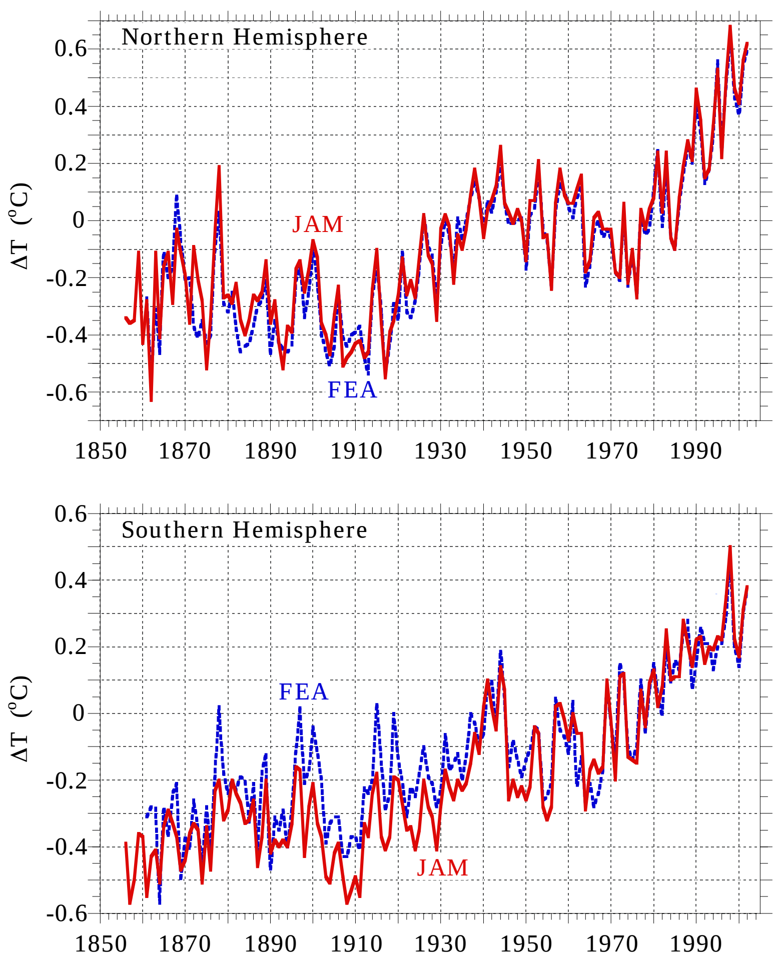

Figure 2. Dependence of the estimated ΔT_{2x} on radiative forcing model.

The difference in the data sets comes from the two averaging methods used: (1) a "simple" averaging technique used by JAM, and (2) an "optimal" averaging technique used by FEA. The difference between the two methods was highlighted in the IPCC Third Assessment Report [Fig. 2.7, *Folland et al.,* 2001b]. When data are plentiful, there is no systematic bias between the two methods. The systematic difference in the Southern Hemisphere averages between the two methods is clearly a cause for concern. At this stage definitive statements as to which method gives the best hemispheric time series cannot be made. However, it is recognized that this subject needs to be studied further. These concerns will be addressed in detail in the near future.[3]

5. Learning ΔT_{2x} Over Time

The uncertainty in ΔT_{2x} due to the natural variability of the hemispheric temperatures can be diminished in the future as additional observations become available. We illustrate this learning in Figure 3 where estimates of ΔT_{2x} for the GTA radiative forcing are shown in the form of box plots at 10-year intervals from 1940 to 2000, each with the observed temperatures starting in 1856. It is seen that the 5% confidence value for ΔT_{2x}, shown by the bottom of the box, changes very little with time, from about 1.2°C to 2.0°C. There is a larger variation of the 50% confidence value for ΔT_{2x}, shown by the solid line within the box, from about 6°C in 1950 to 3°C in 2000. The 95% confidence value for ΔT_{2x}, shown by the top of the box, in general decreases, from almost 27°C in 1940 to 12.5°C in 2000. Superposed on this downward trend in the 95% confidence level is an oscillation, apparently as a result of the previously mentioned temperature oscillation over the North Atlantic Ocean.

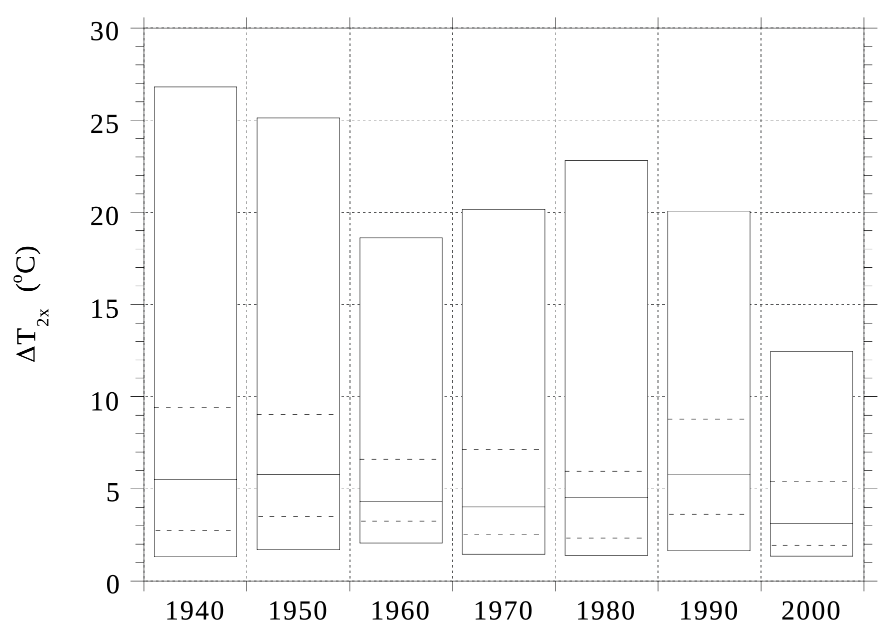

Figure 3. Box plots of the estimation of ΔT_{2x} over time.

[3] Personal communication on 10 September 2003 from N. Rayner, C. Folland, P. Jones and P. Thorne.

6. Conclusion

Progress in reducing the uncertainty in the value of ΔT_{2x} will require reducing the uncertainty in the radiative forcing, not only by aerosols, but also by the Sun and volcanoes. If the radiative forcing by aerosols cannot be learned exogenously but only endogenously from the observed temperature changes, then the uncertainty in the southern hemisphere temperatures must be reduced. The uncertainty in climate sensitivity due to climate noise can be reduced by learning over time, that is, by performing future estimations using longer observational records. Thus, it is quite likely that the formulation and negotiation of policies to abate human-induced climate change will, for the foreseeable future, continue to be made against a backdrop of deep uncertainty. Such policy formulation and negotiation under uncertainty can be facilitated by robust adaptive decision strategy [*Lempert et al.,*1996, 2003; *Lempert and Schlesinger,* 2000, 2002;].

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