ON THE RESPONSE OF THE CLIMATE SYSTEM TO SOLAR FORCING

Introductory Paper

L. BENGTSSON
Max-Planck-Institut für Meteorologie, Bundesstrasse 55, D–20146 Hamburg, Germany;
Environmental System Science Center, University of Reading, UK
(E-mail: bengtsson@dkrz.de)

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Abstract. The climate response to changes in radiative forcing depends crucially on climate feedback processes, with the consequence that solar and greenhouse gas forcing have both similar response patterns in the troposphere. This circumstance complicates significantly the attribution of the causes of climate change. Additionally, the climate system displays a high level of unforced intrinsic variability, and significant variations in the climate of many parts of the world are due to internal processes. Such internal modes contribute significantly to the variability of climate system on various time scales, and thus compete with external forcing in explaining the origin of past climate extremes. This highlights the need for independent observations of solar forcing including long-term consistent observational records of the total and spectrally resolved solar irradiance. The stratospheric response to solar forcing is different from its response to greenhouse gas forcing, thus suggesting that stratospheric observations could offer the best target for the identification of the specific influence of solar forcing on climate.

Keywords: solar forcing, climate change, climate variability

1. Introduction

Traditionally, long-term observations of the sunspot number have been combined to surface temperature records to provide evidence for a solar forcing-climate relationship (Eddy, 1976). The coincidence of the Maunder minimum with a period of very cold winters in Europe in the late 17th century (Lean, 2000; and others) suggests the existence of such a relationship, which is also supported by numerous model studies (Crowley, 2000; Cubasch and Voss, 2000; Shindell *et al.*, 2001; Rind, 2002). The practice of using estimates of global temperature records, combined to proxy-based reconstructions of the total solar irradiance, TSI, to infer the sensitivity of the climate system to long-term solar variations is, however, questionable. Changes in the solar input associated with the 11-year cycle of solar activity are nowadays well documented, but the actual range of TSI variations over longer time scales is still a subject of debate (see different contributions in this issue) as no direct measurements exist prior to the advent of satellite-based observations, in 1978 (Fröhlich, 2006). Present reconstructions of the TSI over millennial timescales have been subject to several revisions (Hoyt and Schatten, 1993; Lean *et al.*, 1995, 2002; and others), and latest results (Wang *et al.*, 2005) suggest that the amplitude of past

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solar irradiance changes could in fact be significantly smaller than had been previously estimated. Similarly, long-term instrumental records of past climate data are limited in time and space and very few records exists before the middle of the 18th century (Jones and Moberg, 2003). Surface temperature variations in Europe, where most long-term temperature records are found (e.g., Luterbacher *et al.*, 2004), are only weakly correlated with the global surface temperatures (Bengtsson *et al.*, 2006).

Another shadow on the discussed evidence for the solar-climate variability relationship is cast by a possible model underestimation of the reproduced system internal variability (Oldenborgh *et al.*, 2005), which would force the ascription of extreme past climate events to external causes such as changes in TSI. Recent high-resolution models tend indeed to generate larger internal system variability, thus suggesting that an episode of extreme climate may simply result from the combination of internal system processes, independently of changes in external forcing. As outlined in the next paragraph, climate in Europe is indeed markedly influenced by atmospheric circulation patterns (e.g., Hurrell *et al.*, 2003), which are in turn largely determined by non-linear atmospheric processes, unpredictable on longer time-scales (Gerber and Vallis, 2005).

At middle and high latitudes the variability of the atmosphere is dominated by synoptic weather patterns, with maximum variance on time-scales of a few days to a few weeks. However, because these high-amplitude patterns are essentially non-periodic, considerable high variance contributions can be found in the low frequency part of the signal power spectrum (Hasselmann, 1976; Manabe and Stouffer, 1996). The North Atlantic Oscillation (NAO) (Hurrell *et al.*, 2003) for example, which corresponds to an oscillation in the strength of the geostrophic wind in the eastern North Atlantic (or in the amplitude of the positive pressure gradient between Iceland and the Azores), is presumably generated by changes in the distribution of transient storms (Loptien and Ruprecht, 2005). According to these results, a northerly storm track leads to a high index in the NAO (stronger than average pressure gradient) while a southerly storm track leads to a weaker than average pressure gradient. Other dominant circulation patterns like the Pacific-North America pattern (PNA) are presumably also driven by altered distributions of high frequency weather systems. The most dominant internal mode of variability in the climate system, ENSO (El Niño-Southern Oscillation), is a coupled atmosphere/ocean mode of the tropical Pacific driven by the feedback between surface wind and ocean temperature distribution. ENSO has a major influence in the tropics, but influence also the climate in the extra-tropics, particularly in the western hemisphere. Although ENSO has a dominant time-scale of some four years, the signal has considerable variability. The effect of ENSO is also supported by model studies (e.g., Zebiak and Cane, 1987; Oldenborgh *et al.*, 2005).

As a consequence, the fundamental question of the identification of specific causes of climate change has been found to be a more complex issue than hitherto generally anticipated. Two main issues stand out.

First, the climate system is characterized by internal modes of variability, with time-scales from days to several decades and possibly longer. These modes are characterized by considerable amplitudes, they dominate atmospheric circulation on virtually all time scales and are, as far as we know, largely chaotic and unpredictable beyond a certain time. *The determination of the influence of external forcing on the climate system does thus require a preliminary assessment of the influence of forcing on these modes of variability, as well as of the relative influence of these modes on the global climate system.*

Second, recent studies have demonstrated that the response to external forcing, such as from the sun, aerosols and greenhouse gases, is largely controlled by regional feedback processes, rather than directly by the forcing itself (Boer and Yu, 2003). A consequence of this is that the pattern of response is potentially uncorrelated with the pattern of forcing, making it difficult to separate different forcings as solar forcing and greenhouse forcing may generate the same climate response.

In Section 2 and 3 we will discuss in some detail the possible mechanisms involved in the climate response to external forcing. In Section 4 we will report of a recent study of possible mechanisms explaining the European climate in the period 1500–1900. Finally, general conclusions will be drawn in Section 5.

2. The Mechanisms of Climate Forcing and Response: Geographically-Dependent Feedbacks

External forcing of the climate system is commonly expressed as the difference between incoming and outgoing radiation in Wm^{-2} at the tropopause level (IPCC, 2001, Chapter 6). This definition is justified by the fact that adjustment to altered radiative conditions, such as enhanced solar irradiation, is fast in the stratosphere (of the order of a month), but very slow in the troposphere (of the order of decades) due to the huge heat capacity of the oceans. An increase (decrease) in the solar irradiation at the top of the atmosphere will create an immediate imbalance between incoming and outgoing radiation, thus gradually leading to a warming (cooling) of the troposphere and the Earth's surface. In turn, and because of enhanced thermal radiation from a warmer planet, increased solar input will lead to a slow increase in outgoing infrared radiation that will gradually offset the radiation imbalance at the top of the atmosphere. This is a very slow process, and several centuries will be needed before a complete radiation balance is restored. The effect is analogue in case of warming resulting from increased greenhouse gases concentration levels, but in this case the immediate imbalance is due to *reduced outgoing radiation*. Hansen *et al.* (1997) have undertaken a series of model experiments to explore the response of climate to various forcing factors, including increases in CO_2 levels and in solar irradiance. Although the model used by Hansen *et al.* (ibid) was significantly simpler than present General Circulation Models (GCMs), the latter nevertheless respond rather similarly (e.g., Boer and Yu, 2003). Following

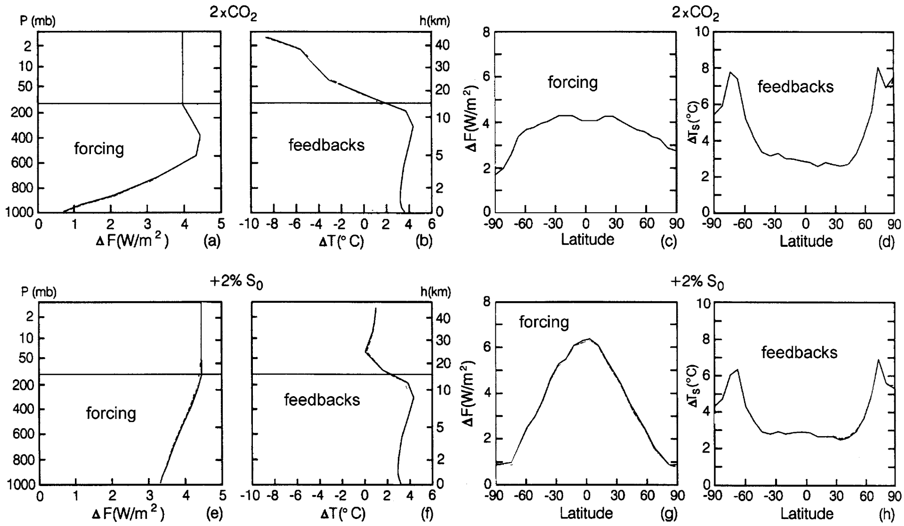

Figure 1. Forcing and feedback in $2 \times CO_2$ and $+2\%$ increase in TSI. (a) and (e) shows the altitude distribution of the net radiative flux and (b) and (f) are the corresponding equilibrium temperature changes. (c) and (g) are the zonal mean forcing versus latitude. (d) and (h) are the corresponding zonal mean surface temperature response. From Hansen *et al.* (1997).

Hansen *et al.* (ibid) we show the climate response to greenhouse gases and solar forcing in Figure 1.

If solar irradiance increases by 2%, the instantaneous flux change at the top of the atmosphere is 4.7 Wm^{-2}. A fraction of this energy is absorbed in the stratosphere and radiated back to space without affecting tropospheric temperatures. After convective radiative adjustment of the stratosphere, lasting a few months or so, the adjusted forcing entering the troposphere amounts to 4.5 Wm^{-2}. This adjusted forcing is comparable in amplitude to the adjusted forcing corresponding to a doubling of CO_2. Restoration of the energy balance without alteration of the atmospheric lapse rate implies a warming of the Earth's surface by about 1.3 K. As indicated by Hansen *et al.* (ibid), feedback processes in the climate system increase the temperature further to between 3 and 4 K, although the range of this interval is strongly model-dependent and is likely to be even larger. While the equilibrium temperature for the doubling of CO_2 and increasing the solar irradiation by 2% is the same in the troposphere, the situation is however quite different in the stratosphere.

In the case of forcing by greenhouse gases, the stratosphere cools because of increased radiation to space (higher concentration of greenhouse gases). The cooling of the stratosphere will not be compensated by radiation from below as this radiation is absorbed in the troposphere and only a weak thermal radiation from the temperature minimum at the tropopause enters the stratosphere. However, the stratosphere responds quite differently to solar warming. In contrast to the case with greenhouse gases, a slight warming occurs instead in the stratosphere as a result of increased solar input, partly due to increased absorption by stratospheric ozone.

Both solar and greenhouse gases forcing amplitudes are largest at lower latitudes (Figure 1). This feature is easily understandable in the case of solar forcing, as solar input to the Earth is a function of latitude. Although less marked, a similar behavior is also observed in the case of greenhouse gases forcing. In contrast, understanding the latitudinal system response to the different forcings cannot be achieved without an analysis of the model results. The immediate warming increases the amount of water vapor in the troposphere (the atmosphere tends to conserve relative humidity), thus enhancing the warming as water vapor is a dominant greenhouse gas. The amount of water vapor in the atmosphere is not uniform as the response will naturally be largest in the areas where the relative humidity is high. This occurs in the Intertropical Convergence Zone (ITCZ) and in the storm track regions. Changes in cloud cover and cloud distribution will strongly affect both short- and long-wave radiation. While low clouds have been shown to cool the atmosphere because of increased reflection of solar radiation, high clouds warm the atmosphere because of enhanced absorption of terrestrial radiation. The total effect of clouds leading to an overall net cooling of some 20 Wm^{-2} in the present climate (Stephens, 2005). Whether changes in clouds will give rise to a warming or a cooling of the climate system is however still unclear, and present models behave differently in this respect. In addition, atmospheric and ocean circulation do transport heat and water vapor to higher latitudes, thus affecting climate and the temperature distribution even further. As a general result, *surface warming is actually larger at high latitudes than at low latitudes in spite of the fact that the forcing has a maximum at low latitudes* (Boer and Yu, 2003).

Figure 1 showed results obtained with a simplified mixed layer model experiment. More realistic models accounting for the full heat exchange of the atmosphere with the oceans however yield different results (Figure 2). Because of the response to the dominant westerlies over the Southern Oceans, an active exchange of heat takes place between the ocean and the atmosphere, leading to an efficient mixing of heat into the deep ocean (Held, 2005). This slows down the warming process, so at the Southern Hemisphere with its strong persistent winds ocean surface warming is much slower than at northern high latitudes. At the Northern Hemisphere, reduced surface warming is restricted to limited parts of the North Atlantic. As a result, most transient climate change models provide a response pattern as shown in Figure 2 (see Räisänen, 2002 for a more complete discussion).

3. Dynamical Contributions

The second problem related to attribution of the causes of climate change is the chaotic variability of atmospheric and ocean circulation patterns over longer time scales. This contribution is considerable, and best illustrated by repeated general

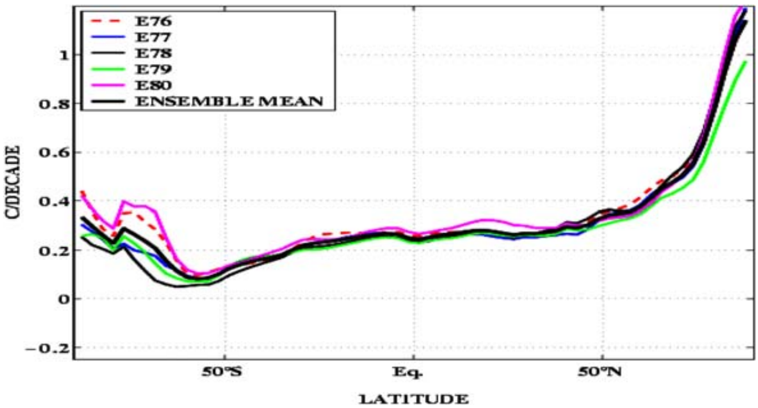

Figure 2. 80-year mean of a five different transient climate change experiments with the Bergen climate model showing the zonal average temperature increase during a time when the CO_2 forcing is increasing by 1%/year compared to a case when CO_2 is constant. The ensemble mean is also indicated. Courtesy H. Drange, Bergen Climate Centre.

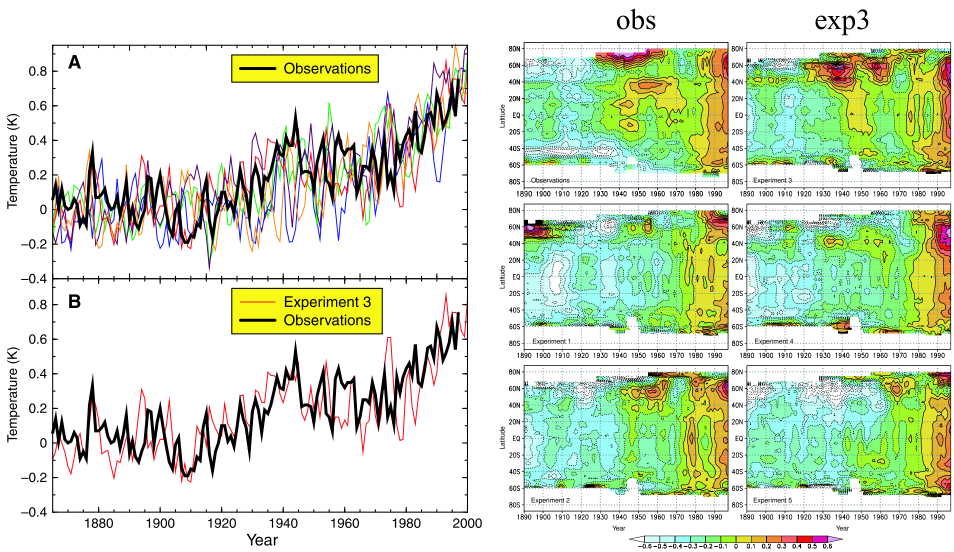

Figure 3. Monte-Carlo simulation with a couple climate model forced by observed greenhouse gases and sulphate aerosols. One out of five simulations almost perfectly reproduces the observed global temperature variability for the period 1860–2000. From Delworth and Knutson (2000).

circulation models simulations using slightly different initial conditions but exposed to identical forcing. In such an ensemble experiment, consisting of five simulations for the period 1860–2000 (Delworth and Knutson, 2000), one of the simulations agreed almost perfectly with the observed global surface temperature (Figure 3). This realization reproduced the warm events in the 1930s and 1940s, and as in observational records, the warming was most pronounced in the Arctic. The other four simulations also reproduced similar anomalies, but at other times. However, the underlying slow warming trend due to the forcing induced by observed greenhouse gases and aerosols increases could be found in all the ensemble experiment members. Considering the above discussed similarities between the system response to solar and greenhouse gases forcing, it is to be expected that a similar result would be achieved if a gradual increase in solar forcing had been considered in the model experiment instead of greenhouse gas forcing. In the next Section, we report the results of a modeling study that do highlight the characteristics of internal climate variability in Europe.

4. A Recent Model Experiment

Recently, Bengtsson *et al.* (2006) investigated the possible causes of observed surface temperature variations in Europe for the period 1500–1900. During this period, only minor changes in greenhouse gases concentrations and in other anthropogenic influences on climate occurred, so that temperature variations could with reasonable confidence be attributed to natural variations. These include solar forcing, and the effect of increased stratospheric aerosol concentration levels following intense volcanic eruptions. The reconstructed observational record of surface temperature variations over the European land area for this period is probably one of the most reliable and best documented in the world, with reconstructions achieving even seasonal resolution (Luterbacher *et al.*, 2004; Xoplaki *et al.*, 2005). It is interesting to note that extreme values of seasonally averaged temperature occurred at different periods. The coldest winter and spring temperatures over the considered time period occurred at the beginning of the 18th century and at the end of the 18th century , respectively, while the coldest summer and autumn took place at the beginning of the 20th century (Table I). Intuitively, this does not suggest major external forcing as the cause of the noted extreme temperatures, but rather natural variations in the climate system.

This interpretation is strongly supported by the result of a climate simulation performed with a coupled general circulation model (Jungclaus *et al.*, 2006) exposed to

TABLE I

Compiled data from Luterbacher *et al.* (2004) and Xoplaki *et al.* (2005) covering the period 1500–2003. From Bengtsson *et al.* (2006).

	Coldest year	Coldest decades	Warmest year	Warmest decades
Winter	1708/09	End 1600	1990	1990s
Spring	1785	Around 1700	1989	1990s
Summer	1902	Beginning 1900	2003	1990s
Autumn	1912	End 1800 Early 1900	1938	1930s

TABLE II
Mean value, standard deviation, and extreme temperature for model and observed data.

	(a) Luterbacher (2005)			(b) Model		
	Annual	DJF	JJA	Annual	DJF	JJA
Mean	8.13	−2.31	16.83	7.39	−2.30	16.96
Stand. dev.	0.41*	1.15	0.47	0.55	1.23	0.55
Min.	6.61 (1695)	−5.69 (1695/96)	15.62 (1821)	5.41 (16)	−7.45 (15)	15.55 (170)
Max.	9.40 (1822)	−0.09 (1842/43)	18.18 (1757)	9.41 (78)	1.47 (52)	19.08 (459)
Max.–Min.	2.79	5.60	2.56	4.00	8.92	3.53

(a) Observed land surface temperature 1500–1900 for the European land area (35°N–70°N, 25°W–40°E) (b) the same for the ECHAM5/ MPI-OM integrated and initialized under pre-industrial forcing. Units in C. From Bengtsson *et al.* (2006). Years of extreme temperatures are indicated within the parentheses. Model years are taken from the 500-year calculation using pre-industrial greenhouse forcing only.
*Standard deviation for the last 100 years is 0.53, suggesting that the variance is underestimated in the earlier data records.

an atmospheric composition characteristic of the pre-industrial period, but without any changes in TSI nor in atmospheric aerosol concentrations (Bengtsson *et al.*, 2006). Irrespective of this, the model simulated the same average, standard deviation and extreme seasons for winter and summer (Table II) as had been derived from the observational temperature reconstruction (Luterbacher, 2005, private communication), thus supporting the idea that long-term variability of climate in Europe is dominated by natural processes. Further inspection of the model results shows that the simulated extremely cold winters and extremely warm summers have large similarities with observational reconstructions, including 30-year long periods of colder and warmer temperatures (not shown). The only difference of significance was that the occurrence of longer periods of extreme summer temperature was slightly higher in the observational reconstruction. This suggests a possible external influence, either from volcanic aerosols or from solar forcing. We shall emphasize here that these results do not imply the exclusion of any influence from low-frequency TSI variations on the climate system, but do rather highlight the difficulty faced when attempting to infer the relative influence of concurring forcing on the climate system from observed temperatures only.

5. Concluding Remarks

The present contribution highlights some of the principal issues that greatly complicate the attribution of climate change and climate variations to a specific forcing. One of the reasons for this difficulty is the large difference observed between the places of forcing and response patterns, the latter being mainly determined by feedback processes in the climate system. These feedback processes explain why warming due to both higher TSI or greenhouse gases is largest at high latitudes, in spite of the fact that forcing there is smaller than at low latitudes. There are also clear indications from model studies (e.g., Boer and Yu, 2003) that the climate response patterns are *additive in their geographical distribution even if the forcing patterns are different.* This stresses the fact that it is the feedback mechanisms which dominate the pattern of response irrespective of the forcing patterns.

Another difficulty is related to the high level of unforced variability in the climate system, resulting from the combination of internal processes. Such processes are chaotic in nature and unpredictable beyond a certain period of time, as predictability is conditioned by the growth of small errors in the initial system state.

The general conclusion of the present discussion is that a reliable attribution of the lower atmospheric response to TSI changes is not achievable on the basis of past temperature records only, as internal modes of variability of the climate system, at least over the last several hundred years, can give rise to variations with amplitudes similar to the ones expected from solar input changes (Bengtsson *et al.*, 2006). We are therefore inclined to conclude that the effect of solar forcing in the troposphere is still an open issue.

The results of this study highlight the need for long term monitoring of both total and spectrally resolved solar irradiance measurements in the upper atmosphere, where the time scale of response to changes in the solar input is significantly shorter than in the troposphere. Furthermore, the stratospheric response to external forcing factors is different depending on the type of considered forcing, thus providing a discrimination criterion among the various forcings of climate.

The following volume chapters, grouped under the title "Detection and attribution of climate change," will underline different aspects of the problem discussed in the present paper. We purposely did not address here the question of how forced stratospheric changes in turn affect the troposphere. This is an area of great interest and importance, which is explored in detail in Section IV of the present volume. A well documented example of such processes is the role of ozone depletion and its influence on the stratospheric circulation over Antarctica (Thompson *et al.*, 2005). Recent studies have demonstrated how this circulation also has affected the troposphere (Thompson *et al.*, ibid), thus contributing to the large warming observed on the Antarctic Peninsula during winter and spring.

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