

News in Climate Science since IPCC 2007

Topics of interest in the scientific basis of climate change

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Preface

The IPCC Fourth Assessment Report (2007) assessed worldwide scientific literature on all aspects of climate change, of the period from 2000 to 2006. The report is intensively used for underpinning international climate policy-making. Since 2006, a large and increasing amount of new literature has appeared, raising the question of whether the IPCC knowledge, based on literature until 2006, is still adequate for the UNFCCC Copenhagen Climate summit in 2009.

Our report offers an assessment of scientific peer-reviewed literature, over the period of 2006 to 2009 on a selection of topics from the physical science basis (Working Group I) of IPCC 2007. In addition, peer-reviewed scientific literature has been assessed in which doubts are expressed on the importance of human influence on global warming. Some of these publications have had a lot of attention, either from the scientific community or in the media. Although this report is, by no means, intended as a comprehensive assessment of all areas covered by the physical science basis of IPCC, the authors found no reasons to deviate from the finding of the IPCC in 2007, that global warming since the middle of the 20th century is very likely to be due to human influence on the global climate¹. The selected topics are: (1) decadal variability: past and future; (2) sea level rise, ice sheets, glaciers, and sea ice; (3) climate changes due to solar variability; (4) the carbon cycle; and (5) climate sensitivity and feedbacks.

This report is part of the study ‘News in Climate Science and Exploring Boundaries’ and is to be considered as one of the background reports summarized in ‘A policy brief on developments since the IPCC AR4 report in 2007’. This study is coordinated by the Netherlands Environmental Assessment Agency (PBL) on request of Dutch Environment Minister, Mrs Jacqueline Cramer, aiming to evaluate new scientific insights regarding the IPCC conclusions of 2007, to provide views on possible acceleration of climate change, and analyze policy response options. The study has been prepared in cooperation with the Royal Netherlands Meteorological Institute (KNMI) and Wageningen University and Research Centre (WUR) with contributions from Energy Research Centre of the Netherlands (ECN), ECOFYS, Utrecht University (UU) and Free University of Amsterdam (VU).

Rob van Dorland and Bart Strengers, 30th of November 2009.

¹ *‘There is very high confidence that the net effect of human activities since 1750 has been one of warming. Most of the observed increase in global average temperatures since the mid 20th century is very likely due to the observed increase in anthropogenic GHG concentrations. It is likely that there has been significant anthropogenic warming over the past 50 years, averaged over each continent (except Antarctica)’*, IPCC Fourth Assessment Synthesis Report 2007.

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1. Decadal variability: past and future

Summary

Claims that global warming has stopped since 1998, are at odds with long-term observations.

Eight of the ten warmest years in recorded history have occurred since 2000. The observed global mean temperature trend in the period from 1975 to 2008 was 0.17 ± 0.03 °C per decade, or 0.20 ± 0.04 °C per decade, depending on the data set used. The calculated trends for the 1998-2008 period range from -0.02 ± 0.09 /decade (based on CRU-data, see Figure 1) to 0.12 ± 0.12 °C per decade (based on data from NASA, see Figure 1.1). The recent trend is significantly lower than the long-term trend over the last three decades, for several reasons. This period starts with a strong El Niño in 1998, increasing global temperature with approximately 0.25 °C, and ended with a significant La Niña in 2008, lowering temperature with approximately 0.15 °C. Moreover, the land temperature trend continued unabatedly upwards, while the reconstructed ocean surface temperatures showed a slowdown – for example, in the Southern Ocean, a region with downward trends over the last 11 years. The CRU data have incomplete coverage, with large gaps in the Arctic. The NASA data provides a more complete representation of the Arctic, by taking spatial correlation into account through extrapolating and interpolating in space. The NASA representation suggests the greatest increases in temperature in the Arctic, in the last decade. This difference between CRU and NASA data is the main cause for the difference in trend of the last ten years.

Global temperature anomalies compared to 1961-1990 average

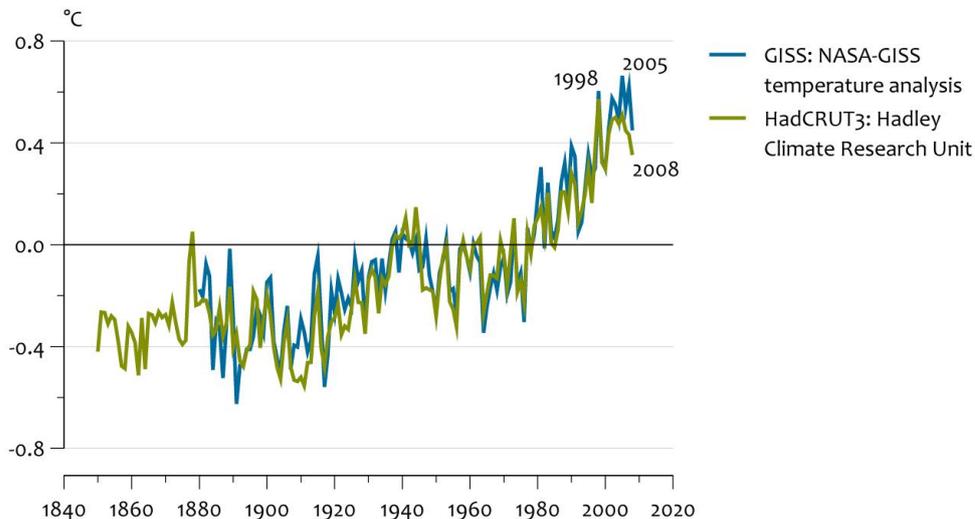


Figure 1.1 Temperature anomalies compared to 1961 to 1990 average, according to the Hadley Centre (Met Office)/Climate Research Unit (HadCRUT3) and NASA Goddard Institute for Space Studies (GISS). The CRUT3 data on land have incomplete coverage, with large gaps in the Arctic. The NASA data provides a more complete representation of the Arctic, which has warmed rapidly, especially over the last decade. This largely explains the significant differences between the data sets of the past 10 years.

The observed temperature trend in Western Europe, over the last few decades, appears about twice as high as the global average. It is suggested that both regional feedbacks and decreased aerosol loading of the atmosphere have played a role.

Since the IPCC AR4, the scientific community has put much effort into investigating the underlying causes of regional climate change, which are of the utmost importance for regional projections. Apart from the relation with global average climate change, the search has been for regional feedbacks amplifying or attenuating the change. In Western Europe, the observed temperature trend over the last decades appears much higher than the global average, by about a factor of two. In winter and spring, higher temperatures are caused by changes in atmospheric circulation, tending to more westerlies in the observations than was calculated by the models. In spring and summer, there is an increase in the amount of solar radiation that reaches the ground, partly due to lower aerosol concentrations. This is underrepresented by state-of-the-art GCMs.

The relatively fast near-surface warming of the Arctic, the ‘Arctic amplification’, due to the well-known surface-albedo feedback, might also be reinforced by changes in atmospheric heat transport into the Arctic surface.

The near-surface warming of the Arctic was almost twice as high as the global average, over the past decades. A recent study examined the vertical structure of temperature change in the Arctic during the late twentieth century, using reanalysis data, and found evidence for temperature amplification well above the surface. Therefore, it was concluded that part of the Arctic temperature amplification cannot be linked to the surface-albedo feedback mechanism, and that changes in atmospheric heat transport into the Arctic surface, in the summer half-year, may be an important cause.

In contrast to the conclusion of the IPCC AR4 that Antarctica was the only continent where no evident global warming had been observed, a recent study concluded that the trend in average surface temperature in both West and East Antarctica were positive for 1957 to 2006.

Also, the average warming of the Antarctic continent is comparable to that in the Southern Hemisphere as a whole. However, the sparseness and short time span of the observations hamper the conclusion of this study. In particular, the extent to which circulation changes in the Southern Hemisphere play a role in temperature trends over Antarctica is still a matter of scientific debate.

Although, since the IPCC AR4, the first steps have been taken in making model-based decadal predictions of forced and natural climate change, this development is still in its infancy.

To produce decadal predictions it is essential that climate models are initialised with the best estimates of the current observed state of the atmosphere, oceans, cryosphere, and land surface. Extended hindcast experiments show the feasibility of decadal predictions. However, future prospects of decadal predictability depend crucially on improved models and data, such as the ARGO network: a network of 3000 drifting buoys, observing the ocean in three dimensions.

For the next few years, expectations with respect to the average global temperature increase have not changed much. The year 2009 is expected to be warmer than 2008. If the sun is entering a stage of very low activity, global temperature increase due to anthropogenic greenhouse gas emissions might be tempered by probably 0.2 °C, within two or three decades.

This coming winter is predicted to see the re-emergence of El Niño, thus, one would expect 2009 to be warmer than 2008. For the coming decades, projected rising levels of greenhouse gases, as well as the behavior of solar activity, are important. The best estimate of the effect of anthropogenic influence is 0.2 °C per decade (IPCC, 2007). If predictions by some solar physicists that the sun is entering a stage of very low activity, for a period of two to seven decades, become reality, global temperature increases might be tempered by about 0.2 °C, within two or three decades (see also Section 2.1.4). However, after recovery of solar activity, the temperature increase will be accelerated, as anthropogenic and solar forcing will both point in the direction of global warming. In case of a ‘normal’ solar behavior, the global temperature rise will be influenced by an approximate 11-year fluctuation of the order of 0.05 °C. These projections may alter due to natural internal factors, such as volcanic eruptions and El Niño events.

1.1. Introduction

The global annual average temperature in 2008 was about 0.1°C lower than in 2007. Nevertheless, 2008 was warmer than any year in the 20th century with the exception of 1998 in the NASA/GISS record, and 1997 and 1998 in the Hadley Centre temperature record. Eight of the ten warmest years since measurements began have occurred since 2000. The last decade was by far the warmest recorded decade globally. The average temperature so far in the 21st century is approximately 0.18°C higher than the average for the 1990s. Claims that global warming has stopped is therefore at odds with long-term observations.

Following IPCC AR4, the scientific community has put a lot of effort into determining the temperature trends at the regional scale and explaining the underlying causes of these changes. The observed temperature trend in Western Europe over the last few decades appears to be much stronger than the global mean, in the order of a factor of two. This is underrepresented by state-of-the-art climate models. In the Arctic too, the near-surface warming has been almost twice as large as the global average in recent decades - a phenomenon known as 'Arctic amplification'. The underlying causes of this temperature amplification remain uncertain, but are extremely important for projections of arctic temperatures. The observed reduction in snow and ice cover in recent decades may have played a role in this, but its contribution relative to other mechanisms is still subject to scientific debate.

Since IPCC AR4, the climate science community has taken the first steps in decadal predictions of forced and natural change based on coupled climate models. A significant source of predictability on the decadal timescale is associated with external forcing, both natural (e.g. solar variability, volcanic aerosols) and anthropogenic (greenhouse gases, ozone, and aerosols) in origin. A fundamental precept in predictability is the notion that long-lived variations, such as those associated with the Pacific Decadal Oscillation (PDO) or the Atlantic Meridional Overturning Circulation (MOC: the 'warm Gulf stream' that transports heat to the north) can be predicted for a significant fraction of their lifetimes. Thus, there is some confidence that naturally-occurring climate variation with decadal timescales may be predictable at times, given an accurate initial state. These times are likely to be when a significant amplitude variation exists. Three extended hindcast experiments demonstrated enhanced skill from initialization at global scale. However, future prospects of decadal predictability specifically depend on improved climate models and data.

1.2 Trends in temperature

1.2.1 Global temperature records

The instrumental record of surface temperature change is based on a combination of land air, marine air, and ocean surface temperature changes recorded over roughly the past 150 years. Several different datasets exist, the most widely used version (HadCRUT) was produced by the Climatic Research Unit (CRU) of the University of East Anglia in co-operation with the Hadley Centre in the UK (Brohan et al. 2006). A second widely used temperature record is the NASA Goddard Institute of Space Studies (GISS) dataset.

1.2.1.1. HadCRUT data set

In constructing the global surface temperature database, HadCRUT, issues of consistency and the homogeneity of measurements over time have been taken into account. Measures have also been taken to ensure that all non-climatic inhomogeneities (including Urban Heat Island effects) have been removed. Over land regions, more than 3000 monthly station temperature time series measurements are used. Coverage is denser in the more populated parts of the world, particularly, the United States, southern Canada, Europe and Japan. Coverage is sparsest over the interior of the South American and African continents and over the Arctic and Antarctic. The number of available stations was small during the 1850s, but increases to over 3000 stations in the latter half of the 20th century. However,

since the 1970s the number of stations has been declining, but still provides sufficient coverage for the determination of the global average temperature. For marine regions, sea surface temperature (SST) measurements taken on board merchant and some naval vessels are used. Coverage is reduced away from the main shipping lanes and was minimal over the Southern Oceans up until the last decade.

For convenience, global surface temperature data are interpolated onto a regular grid (e.g. 5 degree latitude/longitude grid boxes) and formed into 'anomalies' that represent relative relative departures from a base period (rather than absolute temperatures). These gridded data are, in turn, often spatially averaged to yield large-scale mean temperature estimates, such as hemispheric or global mean temperatures.

Several weak points have come to light in this analysis over the last few years, which will be addressed in the next release. The merging procedure over coastlines, e.g. in Europe, introduces spurious trends. As only regions with observations are taken into account in the spatial averaging procedure, the Arctic regions are also severely underrepresented in the estimate of the global mean temperature.

1.2.1.2. NASA/GISS dataset

The method of analysis of the NASA/GISS temperature record was documented in Hansen and Lebedeff (1987), showing that the correlation of temperature change was reasonably strong for stations separated by up to 1200 km, especially at middle and high latitudes. They obtained quantitative estimates of the error in annual and 5-year mean temperature changes by sampling at station locations a spatially complete dataset of a long run of a global climate model, which was shown to have realistic spatial and temporal variability.

Some improvements in the analysis were made several years ago (Hansen et al. 1999; Hansen et al. 2001), including the use of satellite-observed night lights to determine which stations in the United States are located in urban and peri-urban areas, the long-term trends of those stations being adjusted to agree with long-term trends of nearby rural stations. The current analysis uses surface air temperature measurements from the following datasets: the unadjusted data of the Global Historical Climatology Network (GHCN) (Peterson and Vose, 1997 and 1998), United States Historical Climatology Network (USHCN) data, and the Scientific Committee on Antarctic Research (SCAR) data from Antarctic stations. Many of these stations have inhomogeneities due to changes in the observations, by virtue of the huge number of records used it is assumed that these average out. The sea surface temperature is taken from the Hadley Centre, and is therefore identical to the one used in the HadCRUT reconstruction.

The main difference with the HadCRUT analysis is that NASA/GISS takes into account temperature estimates for remote areas, such as the Arctic and the Antarctic regions. Since the Arctic especially, has warmed considerably in the last few decades, as is seen in the melting of sea-ice, for instance (see section 2.4), the NASA/GISS record shows somewhat higher trends than the HadCRUT record, which is based only on direct temperature measurements. This is accomplished at the price of higher noise levels resulting from the filling of unobserved regions.

1.2.1.3. NOAA/NCDC dataset

A third estimate of the global mean temperature has been made by the National Climate Data Center (NOAA). It is based on the GHCN station data over land and the ERSST dataset over sea. In contrast to the HadCRUT dataset, it does interpolate to unobserved regions, but not for regions that are too far away from any thermometers. The results are therefore often midway between the HadCRUT and GISS time series.

1.2.2 Causes of year-to-year variations in the global temperature

The global mean temperature is not observed, the time series that are used are reconstructions based on the available observations. As noted above, these do not cover the whole globe with sufficient density. The polar regions in particular, and until recently the Southern Ocean, were poorly sampled. The uncertainty from the sampling is estimated to be about 0.1°C. The uncertainty in the trend is greatest where there are large temperature changes in these poorly observed areas and when accurate observations are obtained for previously unobserved regions, such as the Southern Ocean over the last decade.

Even in a perfectly observed world there are likely to be natural oscillations in the global mean temperature. No climate model has ever shown a year-on-year increase in temperatures because of the currently expected level of global warming. A significant factor in such oscillations is the El Niño – Southern Oscillation (ENSO). Whether there is a warm El Niño event, or a cool La Niña event makes an appreciable difference in global mean anomalies - about 0.1 to 0.2°C for significant events. There was a significant La Niña at the beginning of 2008, and that undoubtedly played a role in this year's relative coolness. It is worth pointing out that 2000 also had a similarly sized La Niña but was notably cooler than 2008. In the warmest year on record, 1998, a very strong El Niño event occurred. Its effect has been estimated to give temperatures above the ENSO neutral phase of about 0.25°C.

Besides ENSO, there are other sources which are influencing the Earth's temperature, either due to internal variability or due to external forcing. Variations in weather patterns in Asia, particularly in wintertime, have an impact on global temperatures. This is due not only to the fact that Asia is a substantial landmass area, but also the wide temperature variations on this continent in winter. Ultimately, yearly global temperature variations may be as great as 0.1°C.

The two main external (natural) forcings are volcanic variability and solar activity. There have been no climatically significant volcanoes since 1991, so that is not a factor today. Recently, it has been stated that the sun reached minimum activity in 2008. The impact of the solar cycle on the surface temperature record is somewhat disputed (see chapter 3), but could be in the order of 0.05°C (Lean and Rind 2009, Van Dorland et al. 2006) from solar minimum to maximum activity, with a lag of a year or two. Thus for 2008, a deviation in the direction of cooling of somewhat less than 0.05°C might be expected. However, such a small difference would not significantly shift the rankings.

1.2.3 Trend analysis of global temperatures

A global mean temperature rise has been observed since the mid-seventies. Figure 1.2 shows the trends in the period 1975-2008 (in K/year) for the three temperature records. The linear trends are 0.17±0.03°C /decade for HadCRUT and NCDC and 0.20±0.04°C /decade for NASA/GISS. The model results used for IPCC AR4 are also shown, incorporating the natural (solar, volcanic) and the anthropogenic (greenhouse gases and aerosols) radiative forcings. The average of the climate model results agrees within the error bars for observation. While conversely, the central values from observation agree with the model ensemble (see Figure 1.2).

These trends are not greatly affected by shifting to another starting year or by reducing the period by a few years. However, if the trend analysis period is limited to the last 11 years (1998-2008), trends are reduced considerably. Due to the shorter period, uncertainty in the trend increases in an approximately similar ratio (see Figure 1.3). This period has been chosen because of the strong El Niño in 1998 and the significant La Niña in 2008. The calculated trends for HadCRUT, NASA/GISS and NCDC are - 0.02±0.09, 0.12±0.12 and 0.08±0.08°C /decade respectively. Despite this period with the minimum possible trend in the last three decades, the ensemble of climate models and observations agree within the uncertainty limits, although the model spread seems to be too great. The extra spread over the natural variability is most likely due to varying estimates of the cooling effects of aerosols in

the climate models (in fact the right-most boxes are from models that, due to an oversight, did not vary aerosol concentrations at all).

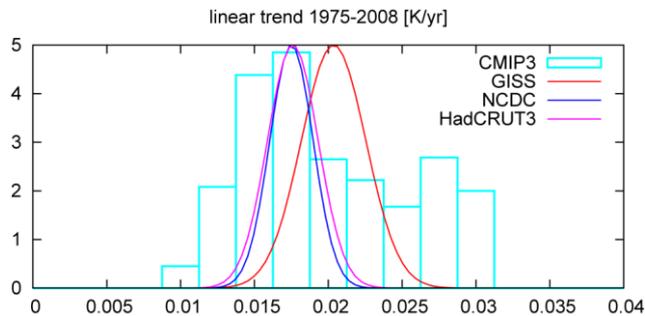


Figure 1.2 linear trends (K/yr) in observations and climate models for the period 1975-2008. Multiple ensemble members of the same model have been weighted by the inverse of the number of ensemble members.

Forecasts do not greatly differ with respect to the temperature increase in the next few years. It is predicted that this coming winter will see the re-emergence of El Niño, so on this basis 2009 may be expected to be warmer than 2008. If, as predicted by some solar physicists, the sun is entering a phase of very low activity for a period of two to seven decades, global temperature increases may be tempered probably by 0.2°C (0.4°C as a maximum estimate, see chapter 3) within two to three decades. Barring any major volcanic eruptions in the tropics, the quiet sun may temper the projected rate of temperature increase by about $0.2^{\circ}\text{C}/\text{decade}$ due to human activity. Lean and Rind (2009) estimated on the basis of anthropogenic forcing and ‘normal’ solar activity only, i.e. increasing solar irradiance in the next five years, an increase in global surface temperature of $0.15 \pm 0.03^{\circ}\text{C}$ in the period 2009-2014, at a rate 50% greater than predicted by IPCC (2007). But, as a result of declining solar activity in the subsequent five years, average temperature in 2019 is only $0.03 \pm 0.01^{\circ}\text{C}$ warmer than in 2014. It should be noted that these estimates for the next decade exclude temperature trends due to (internal) natural variability.

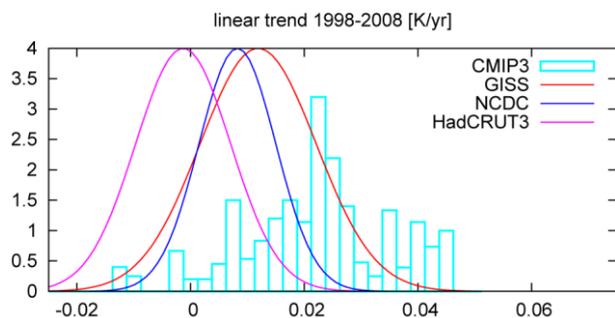


Figure 1.3 linear trends (K/yr) in observations and climate models for the period 1998-2008.

1.2.4 Trends in regional temperatures

Warming has been detected in the global mean temperature and on continental-scale regions, and this warming has been attributed to anthropogenic causes (Stott 2003, IPCC 2007). The observed global warming trend agrees well with predictions (Rahmstorf et al. 2007). However, climate change projections are typically made for much smaller areas. The Netherlands, for instance, corresponds to a

single grid box in most current climate models, but the temperature projections in the KNMI '06 scenarios (van den Hurk et al. 2006, 2007) are based on grid point values of global and regional climate models (GCMs and RCMs, respectively). In this region, temperatures simulated by RCMs do not deviate much from climate models (GCMs), as the prescribed sea surface temperatures (SST) and boundary conditions to a large extent determine the temperature (Lenderink et al. 2007).

1.2.4.1 Western Europe

The warming trend in Western Europe over the last few decades is now so strong that it is discernible in local temperature observations. This opens up the possibility of comparing the trend to the warming predicted by comprehensive GCMs, which until now could not be directly correlated to observations on a local scale, because the variability was too great relative to the trend (or a too small signal-to-noise ratio). The observed temperature trend in Western Europe over the last few decades appears to be much stronger than the global mean, in the order of a factor of two (see Figure 1.4). This is underrepresented in state-of-the-art GCMs (van Oldenborgh et al. 2009).

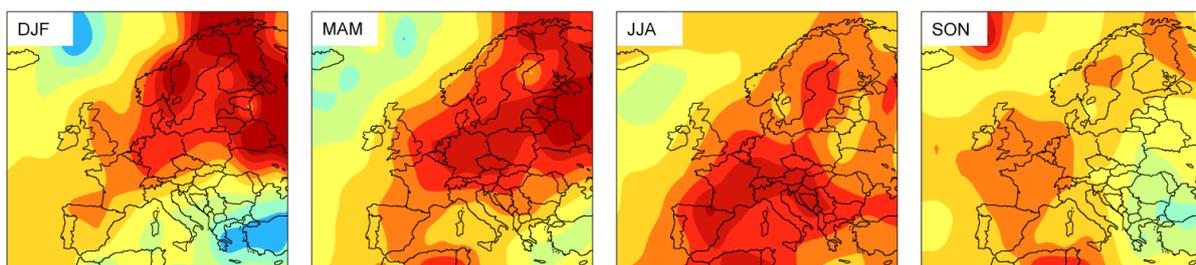


Figure 1.4 Observed trends in surface temperature (colour, [K/K]) March 1950–February 2008, in the merged HadSST2/CRUTEM3 dataset. (a) December–January–February, (b) March–April–May, (c) June–July–August, (d) Sep–Oct–Nov. A value of one denotes a trend equal to global mean warming. Black (red) contour lines indicate that the observed trend is significantly larger (smaller) than the modelled trend (ESSENCE ensemble).

The difference is very unlikely to be due to random fluctuations, either in fast weather processes or in decadal climate fluctuations (Smith et al. 2007, Keenlyside et al. 2008). Changes in atmospheric circulation in winter and spring, tending to more westerlies in the observations than in the models, are important. In spring and summer there has been an increase in the amount of solar radiation reaching the ground. This is partly due to the improved air quality since the mid-1980s, but also seems to be connected with a reduction in cloud cover. This may be associated with the very strong warming trends in southern Europe. Southerly winds then bring warmer and drier air to Western Europe. Note that the prevalence of warm southerly and easterly wind directions has not changed. A common misrepresentation of ocean currents in the Atlantic Ocean also causes climate models to underestimate warming trends in the eastern Atlantic Ocean. This further suppresses warming rates along the Atlantic coasts, but the effect does not reach very far inland.

Smaller contributions to differences between observed and modelled trends come from changes in aerosol effects in spring and summer and from snow cover changes in the Baltic in spring. However, Philipona et al. (2009) and Ruckstuhl et al. (2008) claim that the recent warming amplification relative to the global mean in Western Europe is mainly due to decreased aerosol loading of the atmosphere.

Many of these processes continue to affect regional temperature projections for the 21st century. However, the cause of the recent warming amplification determines the extent of the effect in the future. In case of aerosols are the predominant factor, the warming rate will move to the global average, since IPCC emission scenarios show a slowing down of aerosol changes in the future. In case of the predominance of other factors, as indicated above, climate predictions for Western Europe may show an underestimation of the effects of anthropogenic climate change.

1.2.4.2 Arctic region

Near-surface warming in the Arctic has been almost twice as great as the global average over recent decades (Simon et al. 2005, Serreze and Francis, 2006) — a phenomenon known as ‘Arctic amplification’. The underlying causes of this temperature amplification remain unclear, but are of great importance for temperature projections for the Arctic as well as regions adjacent to the Arctic, such as Western Europe. The reduction in snow and ice cover that has occurred in recent decades (Stroeve et al. 2004) may have played a role in this (Serreze and Francis 2006, IPCC 2007). Climate model experiments indicate that as global temperature rises, Arctic snow and ice cover retreats, causing excessive polar warming (Holland and Bitz 2003, Chapman and Walsh 2007). Firstly, a reduction in the snow and ice cover causes albedo changes and secondly, increased refreezing of sea ice during the cold season and reduced sea-ice thickness both increase the heat flux from the ocean to the atmosphere. Thirdly, changes in oceanic and atmospheric circulation, as well as cloud cover, have also been suggested as causes of Arctic temperature amplification (Zhang et al. 2008, Overland et al. 2008, Langen and Alexeev 2007, Alexeev et al. 2005, Thompson and Wallace 2001, Wu and Straus 2004, Quadrelli and Wallace 2004, Wang and Key 2005). On the basis of climate model results, Winton (2006) and Graversen and Wang (2009) argue, however, that for the most part the Arctic temperature amplification cannot be linked to the surface-albedo feedback mechanism.

Graversen et al. (2008a) examined the vertical structure of temperature change in the Arctic during the late 20th century using re-analysis data, ERA-40 (Uppala et al. 2005) and JRA-25 (Onogi et al. 2007) and found evidence of temperature amplification well above the surface. They concluded that snow and ice feedbacks cannot be the main cause of the warming aloft during the greater part of the year, because these feedbacks are expected to primarily affect temperatures in the lowermost part of the atmosphere, resulting in a pattern of warming that is found in the re-analysis data only in spring. A significant proportion of the observed temperature amplification must therefore be explained by mechanisms that induce warming above the lowermost part of the atmosphere. Graversen et al. (2008a) concluded that changes in atmospheric heat transport into the Arctic in the summer half-year may be an important cause of the recent Arctic temperature amplification.

However, several studies argue that the reported Arctic tropospheric amplification is a non-climatic artefact in the re-analysis data (Thorne, 2008). Grant et al. (2008) show that the vertical structure of the recent Arctic warming is due to the heterogeneous nature of the data source, which incorporates information from satellites and radiosondes. Radiosonde data alone suggest the warming was strongest near the ground. Bitz and Fu (2008) found that in comparison with the observations, the re-analysis exaggerates polar amplification aloft by overestimating the Arctic atmospheric warming and underestimating the Northern Hemisphere atmospheric warming in every season. Specifically, for trends in annual means in the re-analysis for the period 1979–2001, the Arctic warms 2.7 times more than the Northern Hemisphere in the lower middle troposphere, compared with just 1.5 times more in the observations. The smaller warming trends aloft in the observations in winter are more consistent with the amplification of surface warming from ice and snow retreat and the lack of change in the northward atmospheric heat transport for the period 1979–2001.

In their reply, Graversen et al. (2008b) state that the Arctic temperature trend amplification well above the boundary layer – in summer, the maximum amplification is found at a height of around 2 km, and no amplification is encountered near the surface – has been found in two state-of-the-art re-analyses, ERA-40 and JRA-25. Although these datasets show considerable differences regarding the magnitude of the Arctic trends, they both show roughly the same overall vertical structure, on which the Graversen et al. (2008a) conclusions are based. Nevertheless, there is no doubt that more in-situ observations in the Arctic are needed to enhance the quality of future re-analyses, in order to determine the underlying causes of Arctic temperature amplification.

1.2.4.3 Antarctica

Assessments of Antarctic temperature change have emphasized the contrast in recent decades between strong warming of the Antarctic Peninsula and slight cooling of the Antarctic continental interior (Johanson and Fu 2007, Turner et al. 2005). This pattern of temperature change has been attributed to

the increased strength of the circumpolar westerlies, largely in response to changes in stratospheric ozone (Thompson and Solomon 2002). Steig et al. (2009) showed that significant warming extends well beyond the Antarctic Peninsula to cover most of West Antarctica, an area of warming much larger than previously reported. West Antarctic warming has exceeded 0.1°C per decade over the past 50 years, and is strongest in winter and spring. Although this is partly offset by autumn cooling in East Antarctica, the continent-wide average near-surface temperature trend is positive for the period 1957–2006. It should be noted, however, that accurate estimates of the temperature trends are hampered by the sparseness and short duration of the observations on Antarctica.

The extent to which circulation changes in the Southern Hemisphere plays a role in temperature trends over Antarctica is still a matter of scientific debate. An observed trend in the Southern Hemisphere annular mode (SAM), i.e. an increase in the westerlies in recent decades, has involved an intensification of the polar vortex. Steig et al. (2009) found from simulations using a general circulation model that the essential features of the spatial pattern and the long-term trend could be reproduced and suggested that neither can be directly attributed to increases in the strength of the westerlies. Instead, regional changes in atmospheric circulation and associated changes in sea surface temperature and sea ice are required to explain the enhanced warming in West Antarctica.

The source of this trend in SAM may be related to stratospheric ozone losses, greenhouse gas increases, and natural variability. Arblaster and Meehl (2006) examined the annular mode trends in the 20th century using a state-of-the-art global coupled model forced with the observed time series of greenhouse gases, tropospheric and stratospheric ozone, sulfate aerosols, volcanic aerosols, solar variability, and various combinations of these. By comparing the model simulations with observations, it was found that ozone changes were the largest contributor to the observed summertime intensification of the southern polar vortex in the latter half of the 20th century, with increases in greenhouse gases also being an essential factor in reproducing the observed trends at the surface. This conclusion is in line with the recent findings of Perlwitz et al. (2008).

Although stratospheric ozone losses are expected to stabilize and eventually recover to pre-industrial levels over the course of the 21st century, stratospheric cooling due to increases in greenhouse gases will continue to intensify the polar vortex throughout the 21st century. However, using Chemistry-Climate Model Validation (CCMVal) models, Son et al. (2008) found that the net effect of the expected disappearance of the ozone hole and increases in greenhouse gases in the first half of the 21st century will decelerate the tropospheric westerlies in the Southern Hemisphere summer on the poleward side, in contrast with the prediction of most models used in the IPCC AR4 report. As anthropogenic radiative forcing will cause widespread temperature increases over the entire Southern Hemisphere, it is expected that temperatures over East Antarctica will increase, but perhaps at a lower rate than the hemispheric average.

1.3 Decadal Predictability

1.3.1 Background

Climate projections used for the coming century in the IPCC AR4 report were probabilistic using a multi-model ensemble approach. In these projections changes in the external anthropogenic and natural forcing were used, but potential extra information contained in the initial state and the natural variability of the climate system was neglected. The climate system exhibits a natural variability on a decadal timescale and an understanding of the initial state of the climate system together with the decadal variations in the external forcings should therefore narrow the spread of climate projections for the coming decades.

Since AR4 the climate science community has taken the first steps in making decadal predictions of forced and natural change based on coupled climate models. These models are initialized with the best

estimates of the current observed state of the atmosphere, oceans, cryosphere, and land surface. The initial states are influenced by both the current phases of modes of natural variability, and by the accumulated impacts of anthropogenic radiative forcing to date.

1.3.2 Natural decadal variability

The most prominent natural, coupled ocean-atmosphere variability is the El Niño-Southern Oscillation (ENSO). ENSO events are characterised by warming of the central and eastern tropical Pacific Ocean with cooling over portions of the subtropics and the tropical western Pacific. Historically, El Niño events occur about every 3 to 7 years and alternate with the opposite phase of below-average temperatures in the eastern tropical Pacific (La Niña). The nature of ENSO events has varied considerably over time, however, and in recent years many studies have documented decadal and longer-term variability in ENSO (e.g. Trenberth et al. 2007). During El Niño, warm sub-surface water is spread out over the surface, and a reduction of evaporative cooling also warms the tropical Indian and North Atlantic Oceans. Together, these effects cause a noticeable increase in the global mean temperature. ENSO also affects rainfall and temperature patterns over a large part of the globe (van Oldenborgh and Burgers 2005).

Other decadal to inter-decadal variability is especially prominent in the North Pacific (e.g. Trenberth and Hurrell 1994). In this area, fluctuations in the strength of the wintertime Aleutian Low pressure system co-vary with North Pacific SST (Figure 1.5, top), called the 'Pacific Decadal Oscillation' (PDO) (Mantua et al. 1997). Another dominant basin-wide mode in the Pacific is the Inter-decadal Pacific Oscillation (IPO) (Power et al. 1999, Folland et al. 2002). It is not yet clear whether these modes are in fact independent of decadal ENSO variability.

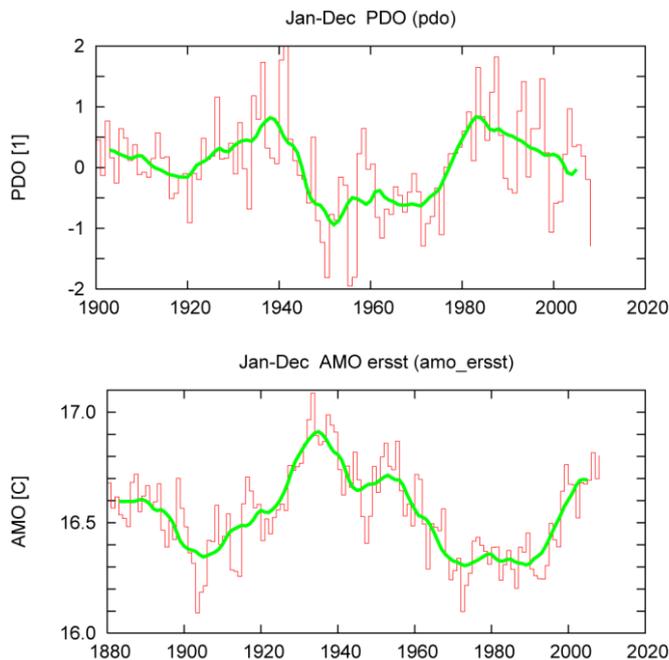


Figure 1.5 Annual mean values of the PDO and AMO indices derived from the ERSST dataset. The green line denotes a 10-yr running mean.

The dominant mode of Atlantic decadal variability is the 'Atlantic Multi-decadal Oscillation' (AMO) (Figure 1.5, bottom). The robustness of the signal has been addressed using paleo-climate records for the last four centuries (e.g. Delworth and Mann 2000). This describes the oscillation between warmer seawater temperatures in the Northern Atlantic from 1930-1950 and from 1990-now, and lower

temperatures from 1900-1920 and from 1970-1990 (see Figure 1.5). These variations affect rainfall in the Sahel region and northwestern Africa, the number of hurricanes over the Atlantic Ocean (Giannini et al. 2003, Lu and Delworth 2005, Hoerling et al. 2006) and temperatures in eastern North America. In models the AMO is connected to oscillations in the strength of the Atlantic Meridional Overturning Circulation. Although AMO is important for the North Atlantic region, its effect on global temperature is limited.

1.3.3 Sources and limitations of decadal predictability

1.3.3.1 External forcing

A significant source of predictability on the decadal time-scale is associated with external forcing, which is both natural (e.g. solar variability) and anthropogenic (greenhouse gases, ozone, and aerosols) in origin. Future external forcing from greenhouse gases will provide significant regional predictability (e.g. Lee et al. 2006), since the increase in concentrations over the next 30 years is about the same, no matter which emission scenario is followed (Hibbard et al. 2007). Future changes in anthropogenic aerosols will be concentrated along industrial regions and affect the climate with a strong regional pattern. Unpredictable volcanic eruptions may be a significant 'wild card' in decadal climate predictions. Similarly, only very general features of the solar cycle can be projected.

1.3.3.2 Natural internal variability

A fundamental precept in predictability is the notion that long-lived variations, such as those associated with decadal ENSO or the Atlantic MOC, can be predicted for a significant fraction of their lifetimes. Thus, there is some confidence that naturally occurring climate variation with decadal time-scales may, at times, be predictable under the proviso of an accurate initial state (Collins 2002, Smith et al. 2007, Keenlyside et al. 2008, Pohlman et al. 2009). These times are likely to occur when there is a significant amplitude variation in the dominant climate factors. At other times, particularly in the nascent phase of variation growth, predictability is less likely. A prerequisite for this is that the climate models accurately represent this low frequency climate variability and that a good estimate of the initial state is available. A challenging problem is that the internal variability of climate models show large differences, even between models with slightly different configurations. The paucity of observational data precludes a definitive statement about which simulation best approximates nature. However, no model replicates time series like the observed AMO.

1.3.3.3 Initialization with ocean data

Natural internal decadal predictability originates mainly in the large heat capacity and inertial momentum of the ocean. This implies that the ocean initial state in particular, must be accurately described. Initialization has three main components to it: the observing system, the assimilation method and the model. The three components are combined to produce initial conditions for the climate model.

a. The observing system

Historically, the sub-surface ocean has been sparsely observed, and some of the data appear to be significantly biased (Domingues et al. 2008, see section 2.2.3), which makes the development and testing of ocean initialization schemes difficult. The construction of reliable initial ocean states is therefore one of the most challenging issues in decadal prediction.

Recent and planned improvements to the observational network offer significant potential for improvements in future forecasting skill. The deployment of a global array of profiling floats by the ARGO programme provides, for the first time, contemporaneous measurements of both temperature and salinity over the upper 2 km of the global ocean, potentially offering a step change to initialize ocean heat and density anomalies.

Several re-analyses of historical ocean observations have been carried out and are being evaluated in the CLIVAR GSOP (Global Synthesis and Observations Panel) intercomparison project. Temperature

and salinity from two of these have already been used to initialize models for decadal forecasts. (Smith et al. 2007, Pohlman et al. 2009). However, the strength of the Atlantic Meridional Overturning Circulation still is completely inconsistent with these re-analyses, as direct observations of this value have only been made since 2004 (Cunningham et al. 2007, Kanzow 2007).

b. The assimilation method

The observations have to be turned into a full three dimensional state of the ocean. The best way to do this is to use an ocean model to propagate the information from the (sparse) observations to other regions. This procedure is known as data assimilation. At the moment it is not yet clear which method yields the best results.

c. The model.

Because of model errors the sub-surface ocean state associated with the initial condition may be significantly different from the climate of the free-running coupled model. This causes a 'coupling shock' as the coupled model rapidly adjusts away from the observed climate estimate towards the coupled model climate. There are several approaches to minimize or correct for this 'coupling shock', such as bias correction and anomaly initialization (Smith et al. 2007, Keenlyside et al. 2008, Pohlman et al. 2009). These are pragmatic approaches and due to the non-linearity of the climate, also questionable. The relative merits have yet to be quantified for decadal time-scales.

Since AR4 there have been many extended hindcast experiments (Smith et al. 2007, Keenlyside et al. 2008, Pohlman et al. 2009, the ENSEMBLES project). They have shown enhanced skill in initialization at global scale (Smith et al. 2007) and over the North Atlantic (Keenlyside et al. 2008, Pohlman et al. 2009). These studies took a very similar approach: initializing a global climate model using observed anomalies and running it forward ten years, while accounting for changes in external forcing (natural and anthropogenic). However the initialization technique, data and models were different and gave rise to different results. Whereas Smith et al. (2007) demonstrated that initialization leads to better predictions for global mean temperature, the results of Keenlyside et al. (2008) and Pohlman were less convincing. The latter two studies showed enhanced skill in the North Atlantic region while the hindcasts of Smith et al. (2007) covered only recent warming and were inconclusive.

These findings are being extended in the ENSEMBLES, THOR and COMBINE European projects and numerous other national projects, so that we can expect that the added value of initialization of decadal forecasts will become clearer over the next few years. The value due to the changes in greenhouse gas and aerosol concentrations are self-evident and lead to skill over almost the entire globe.

1.3.4 Future prospects

The future prospects for decadal predictability specifically depend on improved climate models and data. In this respect, as part of the Coupled Model Intercomparison Project phase 5 (CMIP5; Taylor et al. 2009), modelling centres around the world are planning coordinated suites of decadal hindcast and prediction experiments covering the period 1960-2035. The results of these experiments will be used for the AR5 of the IPCC. Paleo-climate reconstructions will also play an important role.

The sensitivity to model formulation (Palmer et al. 2005) has led to a number of efforts that have demonstrated that a multi-model ensemble strategy is currently the best approach to adequately resolve forecast uncertainty and forecast probability distribution in seasonal-to-interannual predictions (Kirtman and Min 2009, Palmer et al. 2008, Doblas-Reyes 2005, Hagedoorn et al. 2005, Palmer et al. 2004). This is probably also true for decadal predictions. Future decadal forecasts will therefore be probabilistic in nature.

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2. Sea level rise, ice sheets, glaciers, and sea ice

Summary

Since the IPCC AR4, many studies have been published concerning the various contributions to sea level rise, as an important aspect of climate change, in terms of impact on society.

For the 1961-2003 period, the global sea level rise budget due to melting land ice is still not closed; in the IPCC AR4 it is stated that the explained (attributed) sea level rise is 0.7 mm/yr less than the observed sea level rise.

There is a tendency towards a larger Greenland and Antarctic loss of ice mass than presented in the IPCC AR4 report. The emerging picture is that we have moved from a more or less steady ice mass, towards conditions of significant retreat, for both the Greenland and Antarctic ice sheets.

Current mass changes in Antarctica are dominated by the retreat of specific basins in West Antarctica. The rapid retreat along the south-eastern side of Greenland, reported in the IPCC AR4 report, has stopped; nevertheless, the total mass loss in Greenland has increased. The lubrication effect (i.e. meltwater increasing the ice flow velocity) is more widespread than previously assumed, but probably not important for the contribution to sea level rise from the Greenland ice sheet. Current estimates of the contributions from both Greenland and Antarctica are 0.5 mm/yr each and, according to observations in the period 2002 to 2009, are accelerating over time. These contributions are much higher than presented in the IPCC AR4 report, which estimated a sea level drop, in future scenarios, due to changes in Antarctica. There is no convincing evidence to adjust estimates of the contributions of small glaciers to global sea level rise.

Data of ocean heat content have been updated and reanalysed: the heat content has increased in the 1969-2003 period, and reached a plateau in 2004 to 2008, rather than displaying an earlier reported period of cooling.

The heat content in the upper 700m has increased over the period from 1969 to 2003, by 0.24 to 0.41 x 10²² J/year. The earlier reported cooling in the 2004-2008 period has been assigned to two systematic biases in the ocean temperature data used. There is no convincing evidence to adjust estimates on the contribution of thermal expansion to global sea level rise.

For 2100, a plausible and physically-based high-end projection for average global sea level rise is higher than the global estimates reported by the IPCC AR4, being 0.25 to 0.76 metres (for the A1FI scenario), relative to 1990 levels, versus 0.55 to 1.1 metres, implying a rise along the Dutch coast of 0.40 to 1.05 metres.

These estimates are higher than those reported by the IPCC AR4, but lower than estimates on the rate of sea level rise during the Last Interglacial stage, of up to 1.4 to 1.9 metres per century. From the paleo climatological evidence, it is known that the rate of sea level rise can be much more than a metre per century, during periods with similar amounts of ice on Earth and temperatures around what might be expected for the near future. However, it is unclear to which extent this can be attributed to enhanced solar radiation in the Northern Hemisphere summers and to higher temperatures, both in combination with changed heat transports.

Sea ice retreat and thinning in the Arctic continue much faster than reported in the IPCC AR4.

Arctic ice coverage in the summer of 2007 reached a record minimum, with the ice surface area extent declining by 42%, compared to the average of the 1979-2000 period (see Figure 2.1). In 2008, the ice extent was slightly larger than in 2007, but still 34% below this average. In 2009, the ice extent was 690,000 square kilometres larger than the second-lowest extent in 2008, but still 1.68 million square kilometres below the 1979-2000 September average (in September the ice extent is at its minimum). Since 1979, the Arctic sea ice extent has been declining at a rate of 11 per cent, per decade. Not only did the ice extent reduce but so did the average thickness of the sea ice, which decreased strongly,

increasing the vulnerability for further changes. The oldest ice has essentially disappeared, and 58% of the Multi-Year Ice (MYI) now consists of relatively young two- and three-year-old ice, compared to 35% in the middle of the 1980s. In addition, submarine sonar measurements (covering the central ~38% of the Arctic Ocean), showed an overall average winter ice thickness of 1.9 metres, in 2008, compared to 3.6 metres in 1980. Changes in atmospheric and ocean circulation increased the vulnerability of the sea ice, over the last ten years. Changing atmospheric patterns, such as the NAO circulation, would probably affect the sea ice in the Arctic, implying the possibility of a partial recovery over the next years, although this might be negated by the effect of increasing temperatures.

Minimum arctic ice extent

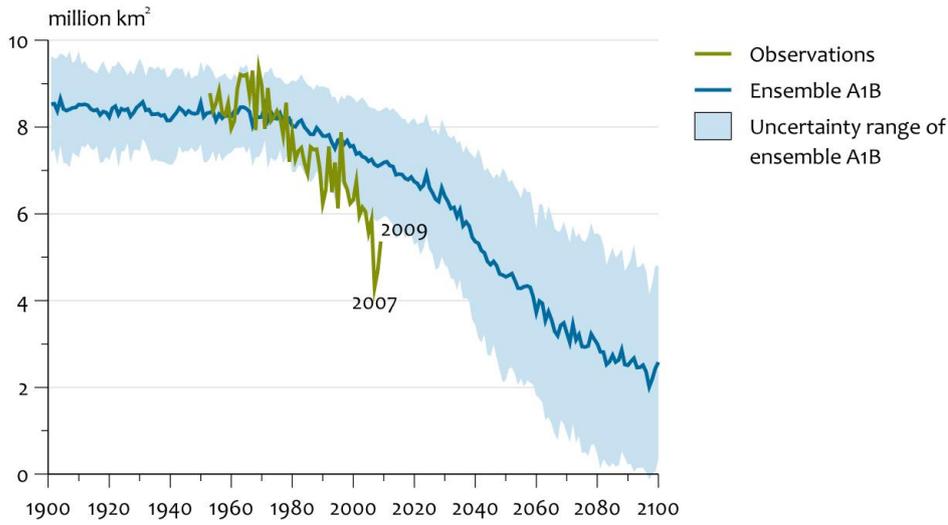


Figure 2.1 Minimum Arctic Sea Ice extent. Observations (green) vs. the average according to models using an 'average' SRES scenario (A1B) and their uncertainty ranges, as reported in the IPCC AR4. Sea ice retreat and thinning in the Arctic continue much faster than in the IPCC range

2.1. Introduction

By the time the IPCC AR4 report (IPCC, 2007) was published a wealth of papers had appeared with recent observations on ice sheets, indicating that many more processes are important for the mass balance of ice sheets than hitherto addressed in the current generation of models. For this reason the IPCC was extremely careful in its projection of sea level rise, because the components related to the recently observed dynamical thinning of ice sheets were excluded. In a separate attempted projection this component was estimated to be between -0.01 and $+0.17$ m by the end of this century. Since the IPCC report many studies have addressed this topic. Here, new insights with respect to the dynamical thinning will be discussed. Another important development since IPCC AR4 is the increasing focus on local sea level rise rather than global average sea level rise. Besides sea level rise, we will also address changes in the sea ice which are important for future climate evolution due to the albedo feedback, for example.

2.2 Components of sea level rise

2.2.1. Observations of sea level rise

Observations of sea level rise taken by satellite measurements since the IPCC AR4 report do not show any major change in the current rate of sea level rise, which is around 3 mm/yr (Cazenave et al. 2009). For comparison, the global mean rate of sea level rise over the 20th century deduced from tide gauges is 1.7 ± 0.3 mm/year (Holgate and Woodworth 2004, Church and White 2006). IPCC AR4 reported a global mean rise of 1.8 ± 0.5 mm/year for the period 1961-2003, and an accelerated rise revealed by satellite radar altimetry of 3.1 ± 0.7 mm/year for the period 1993-2003. Along the coast of the Netherlands, no acceleration has been observed over the last few decades; sea level rise is at a steady rate of about 2.5 ± 0.6 mm/yr (data by RWS National Institute for Coastal and Marine Management, the Netherlands).

The global eustatic sea level rise budget for the period 1961-2003 is still not closed; in the IPCC AR4 report it is stated that the explained (attributed) sea level rise is 0.7 mm/yr less than the observed sea level rise. It has been suggested recently by Oerlemans et al., (2007), that the contribution of small glaciers over the 20th century is somewhat larger than previously estimated. The attribution of sea level rise is even more problematic over a timescale of hundreds of years because it was presumed that sea level rise started at the end of the 18th century before the temperature rise began, which is counterintuitive (Jevrejeva et al. 2008).

The global eustatic sea level rise for the period 1993-2003 can be attributed to thermal expansion and cryospheric changes as presented in IPCC AR4. Although it matches the observations of the tide gauges, this could be coincidental as the period is short in terms of decadal variations in the ocean (Wunsch et al. 2007).

2.2.2. The cryospheric component

Satellite based estimates of mass loss of the major ice sheets have been available only since about 1995. Recently major progress has been made due to the use of gravitational measurements from the GRACE (Gravity Recovery and Climate Experiment) satellites. Although there is still some discrepancy between the different GRACE studies, there appears to be growing agreement on the mass loss of the major ice sheets. Wouters et al. (2008) (WO08 in Figure 2.2) explains most of the differences between the GRACE estimates and therefore we consider it to be the best estimate of mass loss over the last five years based on GRACE data. When taken with the independent estimate of Rignot et al. (2006), a mass loss close to 0.5 mm/yr for Greenland seems to be the most appropriate

figure at this moment. Recent observations presented by Velicogna (2009) show that the mass loss is accelerating (V09 in Figure 2.2).

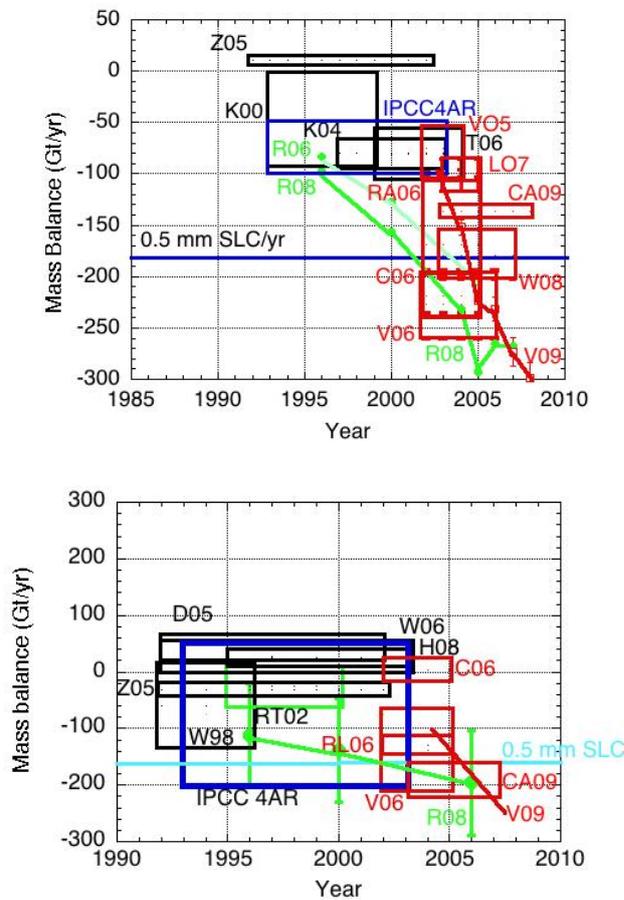


Figure 2.2 Updated mass balance for Greenland (upper panel) and Antarctica (lower panel). Green: Insar (radar altimeter) ice velocity measurements combined with surface mass balance estimates. Red: Gravitational measurements (GRACE). Black: radar altimetry. Blue: IPCC AR4 assessment. Coding indicates initial of first author and year of publication. Z05: Zwally et al. 2005, K00: Krabill et al. 2000, K04: Krabill et al. 2004, R06: Rignot et al. 2006, T06: Thomas et al. 2006, C06: Chen et al. 2006, V06: Velicogna et al. 2006, RA06: Ramilien et al. 2006, L07: Luthcke et al. 2007, R08: Rignot et al. 2008, W008: Wouters et al. 2008, V09: Velicogna 2009, W98: Wingham et al. 1998, RT02: Rignot et al. 2002, D05: Davis et al. 2005, H08: Helsen et al. 2008, D04: Davis et al. 2004, W06: Wingham et al. 2006,.

A similar value can be estimated for Antarctica. Note that the different colours in Figure 2.1 are independent estimates of the mass loss of the ice sheets with all their respective advantages and disadvantages. Not all estimates are in agreement with each other, but there is a tendency towards a larger mass loss than is presented in the IPCC AR4 report. The picture which emerges is that we have moved from a reasonably steady mass towards conditions of significant retreat in both Greenland and Antarctica. The picture for Antarctica is dominated by the mass loss in West Antarctica, particularly in the glaciers of the Pine Island and the Thwaites basins that run off into the Amundsen Sea. This mass loss can be attributed to increased flow speeds of tidal water glaciers, particularly for Antarctica, but possibly also for Greenland (Pritchard et al. 2009). In Greenland increased surface melt dominates the observed mass loss. Using a high resolution regional atmospheric model, Ettema et al. (2009)

calculated an increase in the runoff of 3% per year over the period 1990-2007. The cumulative mass loss calculated by Wouters et al. (2008) is in very close agreement with the results of Ettema et al. (2009). Results based on radar altimetry data over the period 2003-2007 confirm the undeniable loss of mass for Greenland (Slobbe et al. 2008).

2.2.2.1 Antarctica

At the time of writing the IPCC AR4 report projected that mass loss in Antarctica would be dominated by increased accumulation scenarios, possibly partly counteracted by dynamic changes leading to an increasing volume in Antarctica in the near future. This picture has changed dramatically, particularly due to the recent mass loss in West Antarctica. In the Amundsen Sea Embayment the net imbalance in the basin is roughly 50%, implying that the mass loss is twice as great as the mass gain (Thomas et al. 2004). Recent observations indicate a mass loss from this area of about 90 Gt/yr or 0.3 mm/yr sea level rise. So there is clear and persuasive evidence that West Antarctica is shrinking in size while there is no convincing evidence for increased accumulation in East Antarctica. For this reason West Antarctica dominates the picture of mass loss (Vaughan 2008). The rapid retreat observed is in areas where the ice is in direct contact with the ocean. Melt rates can be extremely high in these regions and the ice sheet is vulnerable if the bottom topography towards the centre of the ice sheet is below sea level. In those areas retreat can only be halted if the bottom topography rises again. For the Thwaites and Pine Island basin it may be estimated that this will be the case only after removing 1.5m of sea level rise (Bamber et al. 2009). Katsman et al. (2009) estimate the contribution to sea level rise from those areas in 2100 to be 49 cm at most, because the acceleration of the flow also needs time. Hence, there are at present no grounds for doom scenarios (Vaughan 2008, Pfeffer et al. 2008) involving a full collapse of West Antarctica within a century, leading to a 5 metre rise in sea level.

Apart from the contribution to sea level change, evidence is observed for changing conditions in the ice shelf areas. Major ice shelves have collapsed over the last 10 years and most recently in spring 2009, the Wilkins ice bridge failed. It is now clear that the disappearance of ice shelves leads, temporarily at least, to a speed up in the glaciers behind the shelves. This is a notion which has long been disputed by theoretical glaciologists. On the Antarctic Peninsula, the loss of the Larsen B ice shelf resulted in the glaciers behind it speeding up by factors of 2-8 (Pritchard and Vaughan 2007). This will eventually also lead to a temporary increase in the rate of sea level rise. The Peninsula is one of the most rapidly warming areas in the world with warming rates of over 0.5 K per decade (Vaughan et al. 2003).

2.2.2.2 Greenland

As with Antarctica, most new insights for the Greenland ice sheet focus on ice dynamics. Several important mechanisms are recognized for Greenland. The first process which may be important for enhanced dynamical thinning of the ice sheet is related to the percolation of melt water into the bedrock. Either by enhanced melting or surface lake drainage, this melt water may increase the flow velocity of the ice and this is therefore called the 'lubrication effect'. As a result more ice will be transported to lower areas leading to further melt. Zwally et al. (2002) noted the importance of the lubrication process for Greenland along the EGIG line close to the equilibrium line, i.e. the point on the ice sheet where surface ice melt is equal to ice accumulation. More recently, satellite interferometry data (Joughin et al. 2008) confirm that this process is taking place over a large area of the western ablation zone of the ice sheet related to lake drainage (Das et al. 2008) or increased surface melt (Van de Wal et al. 2008). GPS observations (Van de Wal et al. 2008) show that the acceleration might be up to a few hundred percent for short periods during summer; far greater than earlier measurements indicated. However, the measurements also suggest that despite the slightly increasing ablation rates, velocities have decreased over a 17 year period. This indicates that the ice sheet adjusts its drainage system to the increased melt water on a decadal scale. It implies that the lubrication

process is not very important in the enhanced thinning of the ice sheet and therefore will have a limited effect on ice sheet response to climate warming over the next few decades.

The second process which is important to the dynamical thinning of ice sheets, is the interaction between the ice and the ocean (e.g. Nick et al. 2009). In Greenland there are certain regions bordering on the ocean with an inland area which is below sea level (Pfeffer et al. 2008). These regions are vulnerable to enhanced ice mass loss. Particularly the Jakobshavn Isbrae area is studied extensively as it has a large area below sea level.

The rapid retreat along the southeastern side of the Greenland ice sheet reported in the IPCC AR4 report seems to have come to a halt (Howat et al., 2007). In any event, the bedrock of the ice sheet in this particular region is above sea level so that any further retreat will stop ocean ice interactions and ice sheet shrinkage.

2.2.2.3. Small glaciers and ice caps

A recent paper by Bahr et al. (2009) estimates a possible contribution to future sea level rise from small glaciers which is significantly higher than reported in the IPCC AR4 report or in an update by Meier et al. (2007), yielding a contribution of 0.1 to 0.25 m in 2100. However the timescale of the estimate made by Bahr et al. (2009) may be subject to dispute. Major uncertainties in the estimates for small glaciers can be attributed to uncertainties in the total reservoir of ice on Earth. Recent papers by Bahr et al. (2009) and Hock et al. (2009) are based on larger estimates of the total amount of ice and hence lead to more melt over the next century. A thorough re-assessment of the ice volume is required before the merits of these recent studies can be assessed.

2.2.3. The ocean component

Besides changes in the cryosphere, the global mean sea level may change due to a change in the average density of the ocean (steric changes). Global mean changes in ocean density are dominated by the thermal expansion of ocean water, referred to as the thermosteric contribution. Halosteric changes induced by salinity variations may be important locally but play only a minor role in the global mean (Bindoff et al. 2007).

2.2.3.1. Thermal expansion

Rahmstorf (2007) argues that given the long response time of the ocean to changes in atmospheric conditions, the initial rate of sea level rise may be expected to be proportional to the temperature increase. By fitting the results of a relatively simple climate model, the CLIMBER 3- model, for the 20th century, he found a proportionality constant for the thermosteric sea level of 1.6 mm/yr/K. When this semi-empirical relationship is used to explore future sea level rise it yields a maximum global mean thermosteric rise of 51 cm in 2100. This is substantially greater than the central estimate of 26 cm presented in the IPCC AR4 report. It is also 12 cm (30%) larger than the actual rise modelled by the climate model of Rahmstorf (2007), indicating that the semi-empirical method may be too simple to cover the processes involved in ocean heat uptake (Gregory et al. 2001, Raper et al. 2002). A fit of 21st century data for thermosteric sea level rise from six climate models analyzed by Katsman (pers. comm.) yields a considerably smaller proportionality constant of 0.9 ± 0.3 mm/yr/K. Notably, all the models analyzed gave a smaller proportionality constant than applied by Rahmstorf (2007), indicating that in the latter model thermal expansion shows a high sensitivity to atmospheric temperature rise. Using this lower constant results in a lower estimate for the thermosteric sea level rise of 29 cm by the end of the 21st century for a similarly linear high end temperature increase of 6.4 K in line with the IPCC AR4 report, but obviously a large difference compared with Rahmstorf (2007). So based on the approach proposed by Rahmstorf (2007), but adopting the coefficients determined by Katsman et al.

(2008) implies that there is no convincing evidence that the IPCC AR4 estimates for thermal expansion need to be adjusted.

2.2.3.2. *Heat uptake by oceans*

With over 1000 times the heat capacity of the atmosphere, the world ocean is the largest repository for changes in global heat content (Levitus et al. 2005). Monitoring ocean heat content is therefore fundamental to detecting and understanding changes in the Earth's heat balance. Past estimates of the global Ocean Heat Content Anomaly (OHCA) provide strong evidence of global warming. Climate models exhibit similar rates of ocean warming, but only when forced by anthropogenic influences, i.e. the warming due to the increase in atmospheric greenhouse gases (Gregory et al. 2004, Barnett et al. 2005, Church et al. 2005, Hansen et al. 2005).

While there has been a general increase in the global OHCA during the last fifty years, there have also been substantial decadal fluctuations, including a short period of rapid cooling (4×10^{22} J of heat lost in the 0–700 m layer) from 1980 to 1983 (Levitus et al. 2005). Most climate models, however, do not show decadal variability of this magnitude (Gregory et al. 2004, Barnett et al. 2005 their Figure S1, Church et al. 2005, and Hansen et al. 2005) and it has been suggested that such fluctuations in the observational record may be due to inadequate sampling of ocean temperatures (Gregory et al. 2004).

Gouretski and Koltermann (2007) pointed out systematic errors in measurements affecting the interpretation of Ocean Heat Content. Based on this work new corrections have been applied to the original data by Levitus et al. (2009). As a result the strong interdecadal variability of the global heat content reported in earlier publications (e.g. Levitus et al. 2005) is reduced in magnitude but the linear upward trend in Ocean Heat Content remains similar to earlier estimates. The linear trends in the Ocean Heat Content of the upper 700 m (OHC700) are $0.40 \pm 0.05 \times 10^{22}$ J/yr for the period 1969–2008 and $0.27 \pm 0.04 \times 10^{22}$ J/yr for the period 1955–2008.

These results may be compared with estimates by Domingues et al. (2008) and by Ishii and Kimoto (2009). All three estimates of OHC700 exhibit a similar linear trend over the overlapping period of 1969–2003, but there are differences in the year-to-year variability. These estimates all use different processing methods and there are some differences in the data used. Domingues et al. (2008), for example, do not use any mechanical bathythermograph (MBT) data in their study. In Levitus et al. (2009), Ishii and Kimoto (2009), and Domingues et al. (2008) the trends are 0.32 ± 0.05 , 0.24 ± 0.04 , and $0.41 \pm 0.06 \times 10^{22}$ J/yr, respectively. The general agreement of the three results suggests that the global linear trend is qualitatively robust in terms of the processing methods although there are differences in the magnitude of the trend.

2.2.4 *Local sea level rise*

Since IPCC AR4 several publications have appeared that discuss local sea level rise. Several effects lead to a non-uniform contribution of relative sea level rise over the world ocean (e.g. Milne et al. 2009). Besides the variation in ocean dynamics (Hu et al. 2009) and regional variations in thermal expansion (e.g. Katsman et al. 2008, Yin et al. 2009), gravity-elastic effects are important for local absolute sea level rise.

When ice masses on land melt, the fresh water released is not distributed evenly over the oceans (e.g. Farrell and Clark 1976, Milne et al. 2009). Land-based ice sheets and glaciers exert a gravitational pull on the surrounding ocean, and hence ice mass changes affect the Earth's gravity field. In addition, the gravity field is altered by the elastic deformation of the solid Earth under the shifting loads of ice and ocean water. As a result of these local gravitational and elastic changes, a shrinking land-ice mass yields a distinct pattern of sea level rise. These 'fingerprints' can be used to translate the global mean contributions from glaciers and ice sheets into local contributions.

As a result, near the Dutch coast the contribution from the Greenland ice sheet is reduced by a factor of four near our coast, whereas Antarctic mass loss is enhanced by 20% (Mitrovica et al. 2001). The contribution made by small glaciers to local sea level rise is reduced by about 20%. The net effect of gravitation therefore strongly depends on the ratio of mass loss between Greenland and Antarctica.

For the Greenland as well as the Antarctic ice sheets, there appears to be agreement between the values published by various authors. Farrell and Clark (1976), Clark and Primus (1988) and Mitrovica et al. (2001), for example, reach similar results, although they use different methods and numerical approaches. The results are also in close agreement with simple analytical calculations made more than a century ago (Woodward, 1888). By contrast, the results presented by Plag and Jüttner (2001) deviate strongly. The reasons for this are not fully understood. While the impact of the differences in approach and methodology to solve the sea-level equation have been analysed by the mainstream of authors, (Mitrovica and Peltier 1991), a thorough comparison of this kind is lacking for the results of Plag and Jüttner (2001).

Comprehensive observations of vertical land motion and sea level close to rapidly changing large ice masses can serve as a basis for a thorough validation of the published elastic fingerprints, but such observations are only just starting to become available (Khan et al. 2009).

2.3 Possibility of high rates of sea level rise

2.3.1. Paleo observations on sea level rise

While the sea level has varied over 120 m during glacial/interglacial cycles, there has been little net rise over the past several millennia until the 19th and early 20th centuries, when geological data and tide gauge data indicated an increase in the rate of sea level rise. Estimates of sea level rise under past conditions are of interest for future sea level predictions. For this reason it is most interesting to consider, in particular, the past interglacial period (Eemian, approximately 120 thousand years ago), because the geometrical conditions of the continents (i.e. their position on the globe and the shape and position of mountain ridges) are comparable to the present-day. Although it was only slightly warmer than today (1- 2K), estimates are that the sea level was 4 to 6 metres higher (Overpeck et al. 2006, Duplessy et al. 2007). Crucial to the discussion however, is not the sea level itself, but rather the rate of sea level rise, being the rate of change over time. This latter quantity is more difficult to assess from paleo-climatological studies as uncertainty in the dating increases the uncertainty in estimates of the rate of sea level rise.

In a paper by Rohling et al. (2008) based on marine data from the Red Sea, it is suggested that the maximum sea level rise during the Eemian might have been as much as 1.7 m per century, which is remarkable as Greenland and Antarctica/West Antarctica were the only sources of melt water (e.g. Kopp et al. 2009). More importantly, the Red Sea records suggest that the onset of such rapid increases in sea level can occur within a few hundred years. From a human perspective, this is not 'instantaneous', but still fast enough to have a major impact on society. A recent paper by Blanchon et al. (2009) based on coral reefs in Mexico, even claims rates of sea level rise for the Eemian to be as much as 3 m per century. In taking the Eemian as analogue for future warming we should realize that the Eemian was a period where warming was caused by higher solar radiation in summer at high Northern Latitudes than today, which therefore cannot directly be compared with a near future climate state driven by greenhouse gas increases, but what the data show at least is that high sea level rates, above the present-day sea level, have occurred in a geological context.

2.3.2. A physical based approach

The scenario for sea level rise presented in the Intergovernmental Panel on Climate Change (IPCC) Fourth Assessment Report (AR4) does not include the full range of possible changes, in particular with regard to the contributions from Antarctica and Greenland. Our understanding of recently observed dynamic ice sheet behaviour is limited (Alley, 2008) and it is not adequately represented in the current generation of ice-sheet models. More severe scenarios that partly address this shortcoming have been proposed (Rahmstorf 2007, Pfeffer et al. 2008, Grinsted et al. 2009).

Katsman et al. (2009) presented a high-end scenario for local sea level rise. They assumed that all contributions except that of Antarctica depend (to some extent) on the rise in global mean atmospheric temperature, and use a global average temperature projection of 2 °C to 6 °C in 2100, similar to the *likely* range for the A1FI projections, and a global mean rise of 2.5 to 8 °C for 2200 (Lenton, 2006). The high-end contribution of GIS combines the model-based assessment of surface mass balance changes (IPCC, 2007) with an additional contribution from fast dynamical processes, derived from observationally-based assumptions about the temperature sensitivity of tidewater glaciers and basal sliding. For Antarctica a separate dynamical estimate was based on changes in the mass budget in specific basins.

To arrive at a local sea level rise scenario, regional ocean circulation changes in the North Atlantic Ocean are taken into account (Katsman et al., 2008), assessed from climate model simulations (Meehl et al. 2007b). Other factors taken into account include local gravitational changes and the effects of elastic deformation of the Earth's crust leading to a distinct pattern of sea level rise (see section 2.2.4) when a land-based ice mass shrinks. Eventually, the individual components are quadratically summed (as in AR4), assuming that the uncertainties are independent, and rounded off to the nearest 5 cm. For 2100, a plausible high-end projection of a global mean sea level rise of 0.55 to 1.10 metres and a rise along the Dutch coast of 0.40 to 1.05 metres is obtained (Katsman et al., 2009). Note that this analysis results in global and local estimates above the IPCC projections of 0.25 to 0.76 metres but less than estimates of 1.4 to 1.9 metres per century for the rate of sea level rise during the last interglacial period (Lisiecki and Raymo 2005, Rohling et al. 2008, Shackleton: Zagwijn, 1983).

2.3.3. A semi-empirical approach

A semi-empirical approach was applied by *Rahmstorf* (2007) to predict sea level rise in the next century (see also section 2.2.3.1). A series of papers discussed the statistical details of the approach (Holgate et al. 2007, Schmith et al. 2007 and Rahmstorf 2007b) showing that, depending on the details of the determination of the fits, arguments can be made for and against the existence of a fit between temperature change and sea level change over the last century. It can also be argued whether data over the past century can be used to explore the options for the future. Temperatures increased only ~0.7 K over the last century, whereas the IPCC AR4 estimates for 2100 vary from 1.4 to 5.8 K. This implies that the statistical correlation is applied widely outside the range for which it has been derived. This is usually not considered a sound scientific approach. Moreover, our understanding of sea level rise is somewhat better than simply correlating atmospheric temperature to sea level rise. Sea level rise is basically due to four factors: reduction of small glaciers, thermal expansion and changes in the ice sheets of Greenland and Antarctica. Atmospheric temperature plays a role in most of them, but many other climatological parameters have to be considered as well (change in ocean water temperature, changes in precipitation, decreasing size of glaciers, etc). There is no evidence at all that the recent changes observed in Greenland and Antarctica are directly related to changes in the atmospheric temperature. However, the poorly understood mechanisms behind these recent changes may well dominate the high sea level estimates. As indicated above, based on kinematic considerations rather than temperature change data, Katsman et al. (in prep.) predict that Antarctica and Greenland will account for more than half a metre of sea level rise in the next century. Their more cautious estimate of a high-end global sea level rise of 0.50 to 1.15 m (including gravity effect, dynamical ocean

changes and vertical land movement for the Dutch coast) might therefore be a better way of moving forward.

2.4 Sea ice

2.4.1 Northern Hemisphere

It was concluded in IPCC AR4 that mostly ice-free late-summers may be seen in the Arctic by the end of this century, but the dramatic decline in Arctic summer sea ice cover in the past three years (see Figure 2.1) suggests that ice-free summers may start to occur much sooner (Stroeve et al. 2007). Through its role in regulating the exchange of energy between the ocean and atmosphere, it is anticipated that ice loss will influence temperature, atmospheric circulation and weather patterns.

Ice coverage in summer 2007 reached a record minimum, with ice extent declining by 42% relative to conditions in the 1980s (i.e. -10.2% per decade) (Comiso et al., 2008). The much-reduced extent of the oldest and thickest ice, in combination with other factors, such as ice transport, that assist the ice-albedo feedback by exposing more open water, help to explain this large and abrupt rate of ice loss. In 2008 ice extent was slightly greater than in 2007, but still 34% less than the average in the period 1979-2000. In 2009 the ice extent was again almost equal to the low level of 2008 (see Figure 2.1). The reduction in the maximum winter extent is smaller, showing a decrease of 2.9% per decade (Stroeve et al., 2007).

Maslanik et al. (2007b) combined satellite-derived estimates of sea-ice age and thickness to produce a proxy ice thickness record for 1982 to the present. These data show that in addition to the well-documented loss of perennial ice cover as a whole, the amount of oldest and thickest ice within the remaining multi-year ice pack has declined significantly. The oldest ice types have essentially disappeared, and 58% of the Multi-Year Ice (MYI) now consists of relatively young 2 and 3-year-old ice compared to 35% in the mid-1980s. Submarine sonar measurements (covering the central $\sim 38\%$ of the Arctic Ocean), also showed an overall mean winter thickness of 1.9 metres in 2008 compared to 3.6 metres in 1980 (Kwok and Rothrock, 2009). Using a coupled ice-ocean model, Lindsay et al. (2009) determined that the mean September ice thickness has been declining at a rate of 0.57 metres per decade over the last 29 years.

Kwok et al. (2009) estimated the volume of the Arctic sea ice from ten Ice, Cloud, and land Elevation Satellite (ICESat) campaigns spanning a 5-year period between 2003 and 2008. They found ice losses in 2008 relative to the mean volume in 2003/2004 of 42% and 21% for the autumn and winter, respectively. This seasonal contrast in the volume losses is high: the larger volume loss during the autumn is likely to be due to the later formation of the seasonal ice cover associated with the record minimums in summer ice coverage in recent years.

This extreme and abrupt loss of ice cover in 2007 and the very low levels in 2008 and 2009, following the extensive and sustained reduction in the oldest and thickest ice beginning in the late 1980s, is consistent with the premise that younger, thinner ice is likely to be more sensitive to melt and to area loss due to ridging and rafting, with a variety of implications for the basic nature of the Arctic Ocean (e.g. McPhee et al. 1998). The change in ice thickness from the late-1980s to mid-1990s shows the effects of the highly positive phase of the Northern Annular Mode (NAM), with strong counter-clockwise atmospheric wind transport in the eastern Arctic moving older ice northward to be replaced by first-year ice, and exporting old ice south through Fram Strait between Greenland and Iceland. The changes in age and thickness from the mid-1990s to the mid-part of the current decade reflect a different circulation pattern, where cross-Arctic transport from the Pacific side of the Arctic to the Atlantic side is prevalent (e.g. Maslanik et al. 2007a). The result is loss of ice in the Chukchi and Beaufort seas earlier in the period, combined with losses in the East Siberian Sea during the last four years. Over the past ten years, sea ice thickness associated with older ice has increased in the Laptev

Sea and the Fram and Nansen basins, which were regions of ice loss from the late-1980s to mid-1990s. A return to strongly positive NAM conditions or similar patterns would presumably remove much of this ice, further reducing the extent of the thickest portion of the perennial pack.

Another significant change is the role of the Beaufort Gyre - the dominant wind and ice drift regime in the central Arctic. In the past, ice within the Gyre circulated for years within the Arctic Basin in a clockwise pattern as it aged and thickened (e.g. Tucker et al. 2001). Since the late-1990s however, ice has typically not survived the transit through the southern portion of the Beaufort Gyre, severing the previously continuous, clockwise journey of the MYI. The western Arctic Basin is therefore acting as a new area of 'ice export' in which MYI is removed through a combination of transport and melt. Hence, rather than helping to replenish the old, thick ice, a strengthened Gyre under current conditions instead assists in the transition to younger, less extensive perennial ice cover. As a result of the change in the age of the ice, other properties change as well. For example, ice strength generally increases with age due to brine drainage (e.g. Kovacs 1996), so a shift towards younger ice means that, on average, the ice cover can more easily compress in upon itself via ridging and rafting, producing more open-water area which in turn reduces surface albedo and absorbs more heat, comparable to conditions in 1997–1998 that may have contributed to substantial ice loss in the Beaufort Sea (McPhee et al. 1998). This process adds a dynamical component that strengthens the ice-albedo feedback.

The dramatic reduction in the minimum extent of Arctic sea ice in recent years has been accompanied by surprising changes in the thermohaline structure of the Arctic Ocean, with a potentially important impact on convection in the North Atlantic and the Meridional Overturning Circulation of the world ocean (McPhee et al. 2009). Extensive aerial hydrographic surveys carried out in March–April 2008, indicate major shifts in the amount and distribution of freshwater content when compared with winter climatological values, including substantial freshening on the Pacific side of the Lomonosov Ridge. Measurements in the Canada and Makarov Basins suggest that the total freshwater content there has increased by as much as 8500 cubic kilometres in the area surveyed, effecting significant changes in the sea surface dynamic topography, with an increase of about 75% in steric level difference from the Canada to Eurasian Basins, and a major shift in both surface geostrophic currents and freshwater transport in the Beaufort Gyre.

By combining satellite measurements of sea-ice extent and conventional atmospheric observations, Francis et al. (2009) found that varying summer ice conditions are associated with large-scale atmospheric features during the following autumn and winter well beyond the Arctic boundary. Mechanisms by which the atmosphere 'remembers' a reduction in summer ice cover include warming and destabilization of the lower troposphere, increased cloudiness, and slackening of the poleward thickness gradient that weakens the polar jet stream. Therefore this ice atmosphere relationship suggests a feedback into the general circulation in the northern hemisphere, although care should be taken in interpreting trends over short periods (Graversen et al. 2008).

As the retreat acts as a feedback for the climate system it is important to quantify the rate of retreat. A model study by Ridley et al. (2007) found a large range in the sensitivity of Arctic sea-ice retreat to global temperature change, from 11 to 18% per degree C. Differences in the sensitivities can be attributed to differences in the amount of ocean and atmospheric heat transported from low to high latitudes dominating over local radiative contributions to the heat budget.

If summer sea ice continues on its downward trajectory, which is likely to occur as greenhouse gases accumulate in the atmosphere (Stroeve et al. 2007), then the large-scale atmospheric winter patterns associated with below-normal summer ice cover are also likely to continue. Because summers with below-normal ice extent tend to precede winters characterized by a neutral-to-negative NAO phase, the persistently positive NAO of the late-1980s and early-1990s that is blamed for removing much of the Arctic's thick sea ice (Rigor and Wallace 2004) appears less likely to recur. Furthermore, the prevalence of a near-neutral NAO phase may promote a recovery in the ice cover (Serreze et al. 2007, Rigor et al. 2002), although perhaps not enough to counteract the thermodynamic consequences of increasing greenhouse gases.

In light of the rapid recent retreat of Arctic sea ice, a number of studies have discussed the possibility of a critical threshold (or ‘tipping point’) beyond which the ice–albedo feedback causes the ice cover to melt away in an irreversible process. The focus has typically centred on the annual minimum (September) ice cover, which is often seen as particularly susceptible to destabilization by the ice–albedo feedback. Eisenman and Wettlaufer (2009) examined the central physical processes associated with the transition from ice-covered to ice-free Arctic Ocean conditions. They show that although the ice–albedo feedback promotes the existence of multiple ice-cover states, the stabilizing thermodynamic effects of sea ice mitigate this when the Arctic Ocean is ice covered during a sufficiently large fraction of the year. These results suggest that critical threshold behaviour is unlikely during the approach from current perennial sea-ice conditions to seasonally ice-free conditions. In a further warmed climate, however, Eisenman and Wettlaufer (2009) find that there may be a critical threshold associated with a sudden loss of the remaining wintertime-only sea-ice cover.

2.4.2. Southern Hemisphere

The annual mean extent of Antarctic sea ice has increased at a statistically significant rate of 1 % per decade since the late-1970s. The largest increase has been in autumn when there has been a dipole of significant positive and negative trends in the Ross and Amundsen-Bellinghshausen Seas, respectively. Turner et al. (2009) found that the autumn increase in the Ross Sea sector is primarily a result of stronger cyclonic atmospheric flow over the Amundsen Sea. Model experiments suggest that the trend towards stronger cyclonic circulation is mainly a result of stratospheric ozone depletion, which has strengthened autumn wind speeds around the continent, deepening the Amundsen Sea Low through flow separation around the high coastal orography. However, results from a climate model suggest that the observed sea ice increase might still be within the range of natural climate variability.

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3. Climate changes due to solar variability

Summary

Recent studies either confirm or do not convincingly reject the conclusion (as stated in the IPCC AR4 report) that for the period from 1950 to 2005, it is very unlikely that solar radiation has had a significant warming effect.

Claims that heat radiated from the sun could account for more than 50% of the increase in the Earth's average temperatures, since 1950, are at odds with observations. The 11-year solar activity cycle is poorly correlated with temperatures, and plays a minor role, of the order of a few hundredth of a degree.

There are many ambiguities still to be resolved regarding the mechanism of cosmic rays - cloud cover variations due to solar (magnetic) activity.

Firstly, observations do not show clear connections between changes in cosmic rays and cloud amounts. For instance, there is no 11-year cycle in cloud amount, as seen in satellite measurements. Several studies focused on observations of clouds after sudden decreases in cosmic rays. Conclusions from these studies are quite diverse, ranging from no significant effects, to a 7% change in cloud liquid water about a week after the strongest events. Secondly, the transition between the ultra fine particles provided by cosmic rays and actual cloud condensation nuclei (CCN), is still a missing link. From a recent model perspective, the conclusion is that the hypothesised cosmic ray effect on CCN is too small to play a significant role in current climate change. Also, there is scientific debate about the altitude at which cosmic rays are potentially most effective in producing CCN. This is important with respect to the effects on the radiation balance: increasing low cloud formation acts as a cooling factor of climate, whereas increasing high cloud formation has the opposite effect. In some recent studies, model simulations of aerosol nucleation under various atmospheric conditions point to the potential role of electrical charge for aerosol nucleation. The upper tropical troposphere is found to be a favored region for the production of CCN. This would make the total solar effect over the 11-year sunspot cycle weaker, or even at odds with the observations, in case of a very strong effect on high clouds. A few studies hypothesised the aa-index, a measure of the Earth's magnetic field, or derivatives to be potentially important for variations in global average temperature, but the physical explanations for this influence remain unclear. Moreover, the correlations found in the studies considered are either questionable, or based on erroneous assumptions.

The present solar minimum is characterised as very deep. In case of a quiet sun for decades (such as the Maunder Minimum or Dalton Minimum) this might be important to the temperature projections for the coming decades, that is, a solar induced cooling of the order of 0.2 °C, within two to three decades.

The total solar irradiance (TSI) value of 1365.3 Wm⁻² in the recent solar minimum (2008) has dropped 0.25 Wm⁻² below the solar minimum of 1996, and 0.3 Wm⁻² below the minimum of 1986 (i.e. both of the order of 0.02%), indicating that we are experiencing a very deep solar minimum (see Figure 3). This decrease in TSI is four times smaller than the change between solar maximum and solar minimum conditions. However, a long lasting period of solar inactivity, comparable to the Maunder Minimum, may result in a tempering of the projected warming, that is, a solar induced cooling of the order of 0.20 °C (and 0.4 °C as a maximum estimate) within two to three decades on top of the warming of 0.20 °C per decade induced by human influence. After recovery of solar activity, the temperature increase will be accelerated, as anthropogenic and solar forcing will both point in the direction of global warming.

Total Solar Irradiance (TSI), 1979-2009

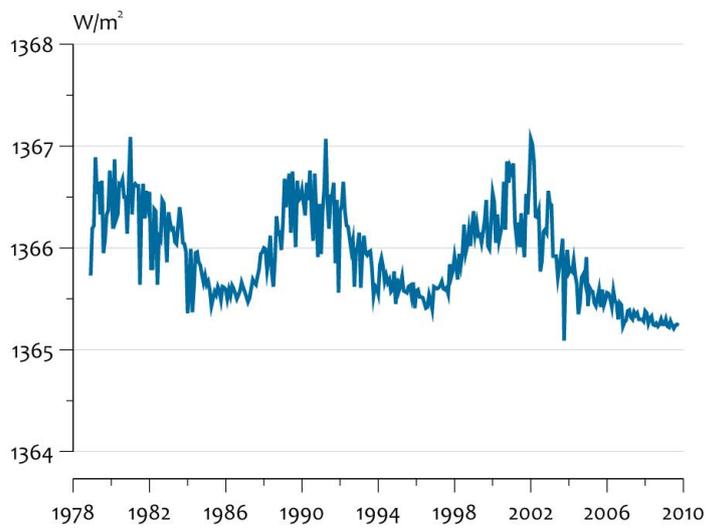


Figure 3. Total Solar Irradiance (TSI) from 1979 to 2009. We are experiencing a very deep solar minimum. This may result in a solar induced cooling of $0.2\text{ }^{\circ}\text{C}$ (max. $0.4\text{ }^{\circ}\text{C}$) within 20 to 30 years.

3.1 Introduction

The extent to which solar activity is a factor in climate change compared with the human influence is a 'hot topic' in climate research, as well as in the public debate. Renewed interest in this subject is due to the claim made by astrophysicists (e.g. De Jager and Duhau, 2009a and 2009b) that solar activity is possibly going to decline for decades. According to NASA, we are experiencing a very deep solar minimum. In 2008, no sunspots were observed on 266 days out of the 366 in a year (73%). To find a year with more blank suns, it is necessary to go all the way back to 1913, when there were 311 spotless days. Such events occur on a centennial timescale, and could be important for temperature projections for the coming decades (see Chapter 1). Over the past four hundred years this has happened twice and may have contributed to global temperature, but by how much is still disputed. This introduces additional uncertainty into forecasts of global temperature change in the near future.

Should we be going down to a quiet sun, comparable to the Maunder Minimum (1645-1715) or the Dalton Minimum (early 19th century), the solar induced temperature drop may be in the order of 0.2 °C below present levels (Van Dorland et al. 2006) and will probably be reached in about two decades. The best estimate of the effect of anthropogenic influence is 0.2 °C per decade (IPCC, 2007). Thus, the most likely scenario in the event of a very deep solar minimum lasting for at least three decades, is that the anthropogenic temperature increase will then be tempered by the solar effect until solar activity recovers. Following recovery the temperature increase will accelerate as anthropogenic and solar forcing will both point in the direction of global warming. Under 'normal' solar activity, the global temperature rise will be influenced by an approximately 11-year cycle of fluctuation, in the order of 0.05°C, rising over the next five years (Lean and Rind 2009, see also Chapter 1).

Since the 4th Assessment Report of the Intergovernmental Panel on Climate Change (IPCC, 2007), a great deal has been published on this subject. On the basis of the new literature assessed, our conclusions do not depart significantly from those of IPCC (2007), also not with regard to the uncertainty surrounding solar-induced climate change. Most problematic is the fact that a vast number of hypotheses are based on correlations between (derived) solar and climate parameters. However, an understanding of the principles of climate change is crucial to be able to make statements about cause and effect relationships.

The main sun-climate relationship is the absorption of solar radiation by the atmosphere and by the Earth's surface. The atmospheric flow is also largely influenced by the distribution of the absorbed solar radiation, as well as the outgoing infrared radiation in the climate system. Even with a constant level of solar activity, climate processes are influenced, if not driven, by solar radiation. Clear examples of this are the diurnal and seasonal cycles. Over a very long timescale, it is the interaction between the Earth's cryosphere and the periodic changes in the Earth's orbital parameters, modifying the integrated seasonal solar insolation at high latitudes, which results in the occurrence of ice ages.

Variations in solar activity may expose the climate system to an additional forcing which induces either climate changes on a global scale or a pattern of responses on a regional scale, depending on the physical mechanism involved. Besides the direct mechanism of variations in the total solar irradiance changing the radiative balance of the climate system, the search is for indirect mechanisms which enhance the effects of variations in solar activity on climate. Examples of these are cosmic ray–cloud correlations and UV - ozone - atmospheric circulation changes. Furthermore, direct and indirect forcings could interact with internal climate system variability (e.g. El Niño – Southern Oscillation or ENSO, Quasi-biennial oscillation or QBO and the North Atlantic Oscillation or NAO), triggering, amplifying or shifting these modes.

Comparing reconstructions of solar activity parameters and climate records may reveal solar-terrestrial relationships but, in general, there is no solid evidence showing to what extent these correlations are caused by changes in solar activity. The origin of such controversies lies in the fact that possible solar signals cannot easily be disentangled from other sources of climate variability, such as volcanic forcing, ENSO and long-term internal variability. Although correlations between solar activity and

climate parameters do not establish cause-effect relationships, they may give indications of the underlying mechanisms of climate change due to solar activity.

This chapter discusses the literature published since the IPCC AR4 report on solar parameters which may potentially influence climate. Section 3.2 deals with solar irradiance changes with respect to recent satellite measurements and estimates of the changes on a centennial timescale. The focus in section 3.3 is on the magnetic activity of the sun and its possible impact on climate, in particular the cosmic ray–cloud hypothesis. In section 3.4 we discuss other solar parameters with a potential impact on the global average temperature, such as solar flares, sunspot numbers and the aa magnetic index. In the last section, section 3.5, an overview of recent solar changes is presented, based on both observations and simulations with a suite of climate models.

3.2 Solar irradiance

3.2.1 Satellite measurements

Since 1979, radiometers on board of satellites directly measure the total solar irradiance without interruption, albeit from a variety of different instruments with various qualities. There is ample evidence that solar irradiance is not constant. For the last almost three solar cycles observations show a periodic variation in total solar irradiance (TSI), with maxima around 1980, 1990 and 2001 and minima around 1986, 1996 and 2008. The difference in TSI between maxima and minima was about 1 Wm^{-2} , which is less than 0.1%. At the top of the atmosphere this translates into an average difference of 0.18 Wm^{-2} , resulting in a cyclic global mean temperature change of a few hundredths of a degree (North et al. 2004, Van Ulden and Van Dorland, 2000)^{2,3}.

Fröhlich and Lean (1998, 2004) made a composite of TSI, dubbed PMOD (Physikalisch-Meteorologisches Observatorium Davos), from various space-born instruments by accounting for inter-satellite differences and the degradation of the individual instruments, in which total irradiance between the successive solar minima of 1986 and 1996 is constant with an accuracy of less than 0.01%. In a recent analysis, Duffy et al. (2009) also found no increasing trends in the successive solar minima of 1986 and 1996, using measurements from ERBS, NOAA-9, NOAA-10, SMM, Nimbus-7, ACRIM2 and ACRIM3. The TSI value of 1365.3 Wm^{-2} in the recent solar minimum was 0.25 Wm^{-2} below the solar minimum of 1996.

However, from their re-evaluation of measurements by the ACRIM1 and ACRIM2 satellites combined with the modelled TSI of Krivova et al. (2007), Scafetta and Willson (2009) concluded that the TSI had increased by about 0.033% per decade between the solar activity minima of 1986 and 1996. The use of the model was needed because of the ACRIM gap (in measurements) in the period from June 1989 to October 1991. In the PMOD composite Fröhlich and Lean (2004) filled the ACRIM gap with data from the Nimbus7/ERB satellites. These corrections are clearly not supported by Scafetta and Willson (2009) and earlier by Willson (1997) as they made the contention that Nimbus7/ERB data are

² The Earth receives only a fraction of the TSI at the top of the atmosphere: the ratio between the surface of a circle and a globe having the same diameter, which is 0.25. Correcting further for the planetary albedo results in a ratio of radiative forcing and TSI changes of 0.18.

³ There is a delay between the equilibrium temperature response and the original radiative forcing, mainly due to the large heat capacity of the oceans. A cyclic forcing of 11 years due to changes in TSI, results in a temperature response which is approximately 25% of the equilibrium response. Comparison of the temperature responses due to forcings of a different signature, such as the non-cyclic total anthropogenic forcing (about 1.6 Wm^{-2} in 2005 relative to 1750), should therefore be treated with caution.

flawed due to uncorrected degradation during the ACRIM gap. Scafetta and Willson (2009) concluded from their composite that increasing TSI between 1980 and 2000 could have contributed significantly to global warming during the last three decades. In an earlier publication, Scafetta and West (2008) estimate that the Sun could account for as much as 69% of the increase in the Earth's average temperature since 1950. This estimate is based not only on the solar minimum trend, but also on their reconstructions of the solar signature in the Northern Hemisphere surface temperature records since 1600 (see section 3.2.2).

However, the warming trend due to solar variability (Scafetta and West, 2008) is at odds with observations (Duffy et al. 2009): firstly, a dominant role for the sun would not explain the observed stratospheric cooling, whereas the enhanced greenhouse effect does. Secondly, the forcing histories of solar and volcanic forcing over the period 1960 to 1998 are partly competitive (Van Ulden and Van Dorland, 2000). TSI variations between maxima and minima in the 11-year solar cycle are in the order of 0.1%, which would influence the surface temperature by only a few hundredths of a degree (North et al. 2004). Thirdly, there might be the possibility of larger indirect effects, but a 69% solar warming trend is unlikely as the hypothesis of Scafetta and Willson (2009) predicts greater than observed temperature variations during prehistoric times. Finally, a TSI trend of 0.033% per decade, as suggested by Scafetta and West (2008), would result in an approximate temperature trend of 0.06°C per decade using the best estimate of climate sensitivity (IPCC, 2007). This would be about one third of the observed trend over the last 30 years.

3.2.2. Contribution of total solar irradiance to global temperatures over the last few centuries

Direct measurements of TSI are only available for the past thirty years. Therefore, long-term solar irradiance reconstructions are assessed by using some proxy for changes in solar output, such as sunspot numbers measured since 1610, in conjunction with recent observations in TSI. The existence of long-term irradiance changes is, in fact, based on three observables. Firstly, changes in the aa index as a measure of the magnetic activity of the sun, which influences the Earth's magnetic field (see section 3.4.1.). The aa index is derived from measurements taken since 1868. Secondly, reconstructions of cosmogenic isotopes, such as ^{14}C , pointing to cosmic ray fluctuations also as a result of the sun's magnetic activity. Cosmogenic radionuclides are formed by high-energy particles entering the atmosphere where they collide with the atoms in the air and produce rare and unstable isotopes. The solar wind shields the atmosphere from these particles so that cosmogenic radionuclide production rates decrease as solar activity increases. The influence of the strength and shape of the magnetic field on the cosmic ray dose on the atmosphere further modulates radionuclide production such that production increases with decreasing field strength and increases towards the magnetic poles. These isotopes are found in sediments and constitute the major source of information on solar activity for the pre-instrumental era. Thirdly, comparison of the sun's activity with the range of variability in Sun-like stars.

Studies (Lean et al. 2002, Foster 2004, Foukal et al. 2004, Wang et al. 2005) published at the time of the IPCC report (IPCC, 2007) suggested that long-term irradiance changes were notably less over the past four hundred years than estimated in the previous IPCC report (IPCC, 2001). These studies questioned each of the three assumptions of changes in the aa index, cosmogenic isotopes and comparison with sun-like stars and pointed to long-term total solar irradiance variations of a factor of about 3 less than those in previous reconstructions (IPCC, 2007).

In terms of current physical understanding, the most likely long-term total irradiance increase from the Maunder Minimum (i.e. the period 1645-1715) to the current cycle minima, including the effects of the 11-year solar cycle, is 1.1 Wm^{-2} , but may be as much as 2.2 Wm^{-2} . As indicated in section 3.2.1, this translates into a probable radiative forcing between the Maunder Minimum and the present average sun of 0.19 Wm^{-2} , and in the upper limit 0.39 Wm^{-2} . Converted into a temperature response using the high climate sensitivity of 4.5 degrees for a doubling of CO_2 (see chapter 5), the upper limit (equilibrium) response would be 0.4°C (Van Dorland et al. 2006).

Scafetta and West (2007) estimated the relative contribution of a solar-induced versus anthropogenic-added climate forcing during the industrial era using different pre-industrial temperature and solar data reconstructions since 1610. These reconstructions are characterized by either more or less variability in temperature (low in Mann and Jones, 2003, and high in Moberg et al. 2005) and in estimated solar forcing (low in Wang et al. 2005, and high in Lean 2000; IPCC, 2007). They concluded that climate is relatively insensitive to solar changes when a temperature reconstruction showing little pre-industrial variability is adopted. In this case most of the global warming since 1900 has to be interpreted as anthropogenically induced. On the other hand, if the high variability temperature reconstruction of Moberg et al. (2005) is adopted, the climate is found to be very sensitive to solar changes and a significant fraction of the global warming that occurred during last century would have been solar induced. Using the high solar forcing composite of Wang et al. (2005), Scafetta and West (2007) estimate that the sun may be attributed to have contributed up to approximately 50% of the observed global warming since 1900.

This conclusion however contradicts IPCC (2007) for which all datasets used by Scafetta and West (2007) were available and discussed. Unlike Scafetta and West, the earlier attribution studies (e.g. Hegerl et al. 2006) took all forcings into account, i.e. solar, volcanic and anthropogenic. It was concluded that the uncertainty associated with the proxy-based temperature reconstructions and climate sensitivity of the models is too large to establish on the basis of these simulations which of the two solar irradiance histories is the most likely. However, in the simulations that do not include anthropogenic forcing, NH temperatures reach a peak in the middle of the 20th century and decrease thereafter, for both the strong and the weak solar irradiance situations. This suggests that the contribution of natural forcing to the observed 20th century warming is small, and that solar and volcanic forcings are not responsible for the degree of warmth that occurred in the latter half of the 20th century (IPCC, 2007), which is consistent with the evidence of earlier work based on simple and more complex climate models (Crowley and Lowery 2000, Bertrand et al. 2002, Gerber et al. 2003, Hegerl et al. 2006, Tett et al. 2007).

Benestad and Schmidt (2009) repeated the analyses of Scafetta and West, together with a series of sensitivity tests to some of their arbitrary choices. These tests clearly showed that the published uncertainty in their estimates was greatly underestimated. In particular, their arbitrary assumption of equilibrium sensitivity has a dramatic impact on their attribution of 20th century changes to solar forcing. It was concluded that the Scafetta and West methodology is highly sensitive to the internal variability of the climate system and the presence of colinear trends in different forcings. Given the concomitant increases in greenhouse gas forcings over the 20th century, this implies that their published attributions greatly exaggerate the role of solar variations in global mean temperature trends. Moreover, the error bars are likely to be significantly larger than reported by Scafetta and West.

Camp and Tung (2007) obtained a global temperature signal of almost 0.2°C attributable to the 11-year solar cycle, using a linear regression between the TSI forcing and surface temperature data for the period 1959–2004. They found a correlation coefficient of 0.47. However, there are several drawbacks to the method they used: their spatial filter may not remove the volcanic-aerosol cooling, which probably has a similar global distribution, and other factors of temperature variability such as the El Niño – Southern Oscillation (ENSO). Van Ulden and Van Dorland (2000) analysed the separate contributions of solar irradiance changes, volcanic eruptions and ENSO to global mean temperature variations from 1882 to 1999, using an energy balance upwelling diffusion climate model. It appeared that fast temperature variations (2-20 years) are primarily due to ENSO and volcanic forcings. The 11-year solar activity cycle is poorly correlated with temperature and plays a minor role, in the order of a few hundredths of a degree (e.g. North et al. 2004). A second point of concern in the analysis of Camp and Tung (2007) is that in many cases the temperature change precedes the change in TSI, pointing to a problem in the cause-effect relationship. However, a time-lag of about two years between TSI (cause) and temperature (effect) would be expected. Thirdly, extending the period would show a much lower correlation, because TSI based on sunspot numbers and global mean temperature fluctuations

are in anti phase (Van Dorland et al. 2006). Therefore, the strong effect of the 11-year solar cycle found by Camp and Tung is questionable.

3.2.3. *Effects of ultra violet changes*

The physics behind the influence of ultra violet (UV) radiation on the composition and state of the atmosphere is well established. The changes in TSI, i.e. as observed over the sunspot cycle, are concentrated in the UV spectral region, implying a considerable influence in the stratosphere and mesosphere as only a small amount of UV radiation enters the troposphere (Gray et al. 2009). Changes in the UV have the potential to alter the chemical composition of the higher atmosphere due to photolysis. Ozone concentrations especially, are affected by changes in UV radiation.

The changes in stratospheric ozone due to UV variations, being in the order of a few percent, are dependent on latitude, longitude and season. They create a differential heating pattern in the stratosphere due to the solar radiation absorption properties of ozone. Moreover, changes in stratospheric ozone interact with the longwave radiation from the surface, greenhouse gases and clouds by absorbing and emitting radiation. Therefore, a complex pattern of temperature changes occurs which, in turn, modifies circulation patterns within the stratosphere. These changes are likely to induce a chain of dynamic interactions with the troposphere (e.g. Kodera et al. 2007).

The effects of UV variations have been extensively studied using fully interactive 3-D coupled chemistry-general circulation models. Although the stratospheric ozone changes as a response to a realistic solar cycle enhancement of UV radiation compares reasonably well with observations, the climate effects are found to be relatively small compared to climate variability. Therefore, model results are statistically significant only for some regions. Although progress has been made in this field, dependency on the model's state hampers reaching any unequivocal conclusions about the relative importance of the various dynamic mechanisms which result in a tropospheric response.

Meehl et al. (2009) approached the UV-climate link from an observational point of view. In the observed global sea surface temperature (SST), they found a response of about 0.1°C . This would require a radiative forcing of more than 0.5 Wm^{-2} , while the forcing due to TSI variations is about 0.2 Wm^{-2} . Meehl et al. studied two mechanisms, the top-down stratospheric response of ozone to fluctuations in UV solar forcing and the bottom-up coupled ocean-atmosphere surface response due to TSI variations, using three global climate models, with each mechanism acting either alone or both together. They showed that the two mechanisms act together to enhance the climatological off-equatorial tropical precipitation maxima in the Pacific, lower eastern equatorial Pacific sea surface temperatures during peaks in the 11-year solar cycle, and reduce low-latitude clouds to amplify solar forcing at the surface. However, it is not clear whether the observed 0.1°C fluctuation can be ascribed to the solar forcing only. The response found by Meehl et al. also cannot be used to explain recent global warming because the 11-year solar cycle has not shown a measurable trend over the past 30 years.

3.3 Magnetic activity

3.3.1. *The cosmic ray – cloud hypothesis*

An indicator of the sun's magnetic activity is the amount of galactic cosmic rays (GCR) bombarding the Earth. This flux is greater during solar (sunspot) minima than during maxima due to the shielding effect of the sun's magnetic field. However, GCRs show some differences from the sunspot record, e.g. GCR decreases tend to occur 1 to 2 years after the corresponding sunspot peaks and the magnitude of GCRs does not uniquely correspond to the magnitude of the sunspot maxima. Direct measurements using ground-based ionization chambers go back to 1934.

From an energetic point of view, the approximate 15% modulation in cosmic ray flux due to solar activity produces an energy change of less than one millionth of the energy change in the 0.1% total solar irradiance cycle. Nevertheless, various scenarios have been proposed whereby galactic cosmic rays might influence climate, for example, by altering the tropospheric electric field and cloud cover. Since the plasma produced by cosmic ray ionization in the troposphere is part of an electric circuit that extends from the Earth's surface to the ionosphere⁴, cosmic rays may also affect thunderstorm electrification (Carslaw et al. 2002). The galactic cosmic rays that reach the troposphere are supposed to influence several microphysical mechanisms (Harrison and Carslaw, 2003), such as providing nuclei for cloud condensation analogous to tropospheric aerosols and electrically charged particles, which serve as ice nuclei, thereby enhancing the glaciation of supercooled cloud droplets. (Carslaw et al. 2002). When solar activity is *high*, the more complex magnetic configuration of the heliosphere *reduces* the cosmic ray flux. So, if the activity of the sun is *low*, more cosmic rays may reach the troposphere.

There are many ambiguities still to be resolved regarding the mechanism of cloud cover variations due to solar activity. These concern, firstly, the absence of a trend in cosmic rays, while the temperature has risen in the last 30 years; secondly, the size of the nuclei provided by cosmic rays and the subsequent coagulation process in the formation of larger aerosols, which may serve as cloud condensation nuclei (CCN); thirdly, whether the change in CCN is significant compared to the variations in background concentrations and, last but not least, the altitude dependence of CCN production, which is important in terms of the effect on the radiation balance: with increasing low cloud formation acting as a climate cooling factor, while increasing high clouds have the opposite effect.

Svensmark et al. (2007) found through laboratory experiments that ions help generate small thermodynamically stable clusters which play a role in CCN production. The nature and extent of this role, however, is more uncertain, and there is still a missing link in the GCR-CCN hypothesis in relation to the transition from the ultrafine particles to actual CCN. Most CCN in the atmosphere are about 100 nm in radius. Yu and Turco (2001) did not find an enhancement of CCN concentrations when comparing GCR fluxes corresponding to solar minimum and solar maximum, respectively.

The GCR induced ionization peaks at an altitude of around 13 km in the atmosphere (Neher, 1971), so the largest effect might be expected in the upper troposphere. Consistent with this notion, Eichkorn et al. (2002), who carried out aircraft measurements of aerosols in the upper troposphere, found large cluster ions which were presumably caused by GCR ionization. Yu (2002) and Arnold (2008), on the other hand, suggested that the bottleneck in the formation of upper tropospheric aerosol particles which are large enough to be climate-relevant is not specifically nucleation, but the availability of condensable gases. Yu (2002) therefore suggests that, despite a lower ionization rate, the lower troposphere might be a more favourable region for a GCR-influence on clouds, due to the higher precursor gas concentrations. However, in more recent studies, Kazil et al. (2006) and Yu et al. (2008) have presented model simulations of aerosol nucleation under various atmospheric conditions. Both studies find support for the role of electrical charge in aerosol nucleation. In both studies the upper tropical troposphere is found to be the favoured region for aerosol nucleation. The latter study also finds strong signs of this in the entire mid-latitude troposphere and over Antarctica.

It can be concluded that there is scientific debate about the altitude at which GCR are potentially most effective in producing CCN. If it turns out to be in the upper troposphere and if it is proven that sufficient CCN are produced to change the extent of high clouds, the effect of CGR on the global mean surface temperature would be the opposite to what is hypothesized by Svensmark and Friis-Christensen (1997), namely, a strong reinforcement of the effects of solar irradiance changes. An impact on high clouds would make the total solar effect over the 11-year sunspot cycle weaker or even contrary to the observations where the effect is very strong.

⁴ The **ionosphere** is the uppermost part of the [atmosphere](#), so named because it is [ionized](#) by solar radiation.

3.3.2. Testing the cosmic ray – cloud hypothesis

Present research on the cosmic ray – cloud connection aims to reveal the potential mechanism(s) of this link and to quantify its effect(s) on climate. Because of the many ambiguities, studies are predominantly focused on the search for correlations between measurements of GCR, ionization rates, clouds and surface temperatures. Another approach is to study the mechanisms using models. Clouds are an important part of the claimed mechanism, since they may act as an amplifier for the temperature response through effects on the albedo (reflected solar radiation) as well as on the greenhouse effect (outgoing longwave radiation). The net effect strongly depends on the altitude of the changes in clouds (see section 3.3.1). Given the present altitude distribution, a 0.6% decrease in cloudiness would result in a radiative forcing similar to the measured forcing of TSI changes between solar minimum and maximum conditions (about 0.2 Wm^{-2}).

Erlykin et al. (2009) studied the variation over time in the period 1956 to 2002, of the globally averaged rate of ionization produced by galactic cosmic rays (GCR) in the atmosphere. The computations of the total ionization rates and long term data on the charged particle fluxes which recently became available (Usoskin and Kovaltsov 2006, Bazilevskaya et al. 2008), show an 11-year modulation as well as a cyclic variation with a period that is consistent with the solar magnetic (Hale) cycle, which is roughly twice the 11-year sunspot cycle. Long term variations in the global average surface temperature as a function of time since 1956 are found to have a similar cyclical component. Cyclical variations are also observed in the solar irradiance and in the mean daily sun spot number. The cyclical variation in the global temperature, showing an amplitude of about 0.07°C , is found to be in phase with the solar cycle as measured from sunspot numbers and the solar irradiance, and out of phase with the cosmic ray variation: the cyclical variation of the GCR cycle is delayed by 2–4 years. This indicates that if this correlation is caused by variations in solar activity, it is most likely to be due to TSI changes rather than due to GCR. Whether there is a causal link between TSI changes and the deduced cyclical changes in global mean temperature is also questionable, since a 22-year cyclical forcing causes a delay in the response in the order of a few years. The results of Erlykin et al. (2009) do not show any such time lag.

The long term variations in both the cosmic ray rate and the solar irradiance are observed to be less than their cyclical variations. Therefore, assuming that there is a causal link between either of them and the mean global surface temperature, the long term variation in the temperature must be less than the amplitude of its cyclical variation of 0.07°C . Hence under the assumptions of Erlykin et al. (2009) the effect of varying solar activity, either by direct solar irradiance or by varying cosmic ray rates, must be less than 0.07°C since 1956, i.e. less than 14% of the observed global warming.

Goode and Palle (2007) investigated albedo changes using earthshine measurements and ISCCP, in order to illustrate a possible solar-albedo link. The earthshine, or ashen light, is sunlight reflected from the Earth and retroreflected from the Moon back to the night-time Earth. Global scale albedo can be determined at any moment by measuring the intensity of the earthshine. Uninterrupted earthshine data from the Big Bear Solar Observatory (BBSO) span the period from November 1998 to the present, with some more sporadic measurements taken during 1994 and 1995. In Palle et al. (2004; 2006), earthshine measurements of the Earth's reflectance from 1999 to mid-2001 were correlated with satellite observations of global cloud properties. These observations were then used to construct a proxy measure of the Earth's global sunshine reflectance. Cloud data were taken from ISCCP.

A major decrease in albedo was found using earthshine measurements from $31.0\pm 0.4\%$ in the period 1994/1995 (Palle et al. 2003) to $29.5\pm 0.2\%$ in the period 1999/2001. Goode and Palle (2007) assumed that the temporal variations in the (reconstructed) albedo are closely associated with changes in cloud cover and calculated an albedo decrease of roughly 2% between the late-1980s and the late-1990s, corresponding to the peak and dip in total cloud coverage, respectively. Such a change in albedo would imply a shortwave radiative forcing of about 6.8 Wm^{-2} . However, direct measurements (ERBS, ISCCP) of the reflected solar radiation show a radiative forcing of approximately half this value (Zhang et al. 2004). The longwave forcing of clouds, which is also substantial, generally

counteracts the shortwave forcing, thus making the net forcing due to clouds much smaller than calculated by Goode and Palle (2007).

For the tropical region (20 S to 20 N) Zhang et al. (2004) found no significant trend in the net outgoing radiation between the late-1980s and the late-1990s. Moreover, the determination of trends is strongly dependent on the period taken, since there are large year-to-year variations. It is likely that apart from the effects of stratospheric aerosols (e.g. the eruption of the Pinatubo in 1991) and of temperature (e.g. El Niño and La Niña events), cloud changes play a role in the net outgoing radiation.

It is certainly not clear from the study by Goode and Palle (2007) whether the changes detected are related to solar activity. There is no 11-year cycle as would be expected if the changes in cloud cover extent are related to cosmic rays. It could be argued that there is possibly a 22-year cycle – which is weaker than the 11-year cycle for cosmic ray variation – but the two decades investigated cover one cycle at most. Hence, it is difficult to be conclusive on the basis of the data.

Sloan and Wolfendale (2008) examined various incidences of ionizing radiation changes in the atmosphere due to cosmic rays, to look for consequential changes in low cloud cover which would result if there is a causal connection. Cases where changes in the ionization rate occurred were examined to see if there was a corresponding change in cloud cover, as might be expected from the causal connection hypothesized in Palle and Butler (2000) and March and Svensmark (2000). Sloan and Wolfendale (2008) deduced a decrease in globally-averaged low cloud cover in solar cycle 22 from minimum (1986) to maximum activity (1990) of about 1.3% using the same ISCCP (International Satellite Cloud Climatology Project) infrared data (such a change may be real or could be due to an artefact of the satellite instrumentation, as discussed in Evan et al. 2007). They estimated that less than 23%, at the 95% confidence level, of the 11-year cycle change in the globally averaged cloud cover observed in solar cycle 22 was due to the change in the rate of ionization from the solar modulation of cosmic rays. As the cloud radiative forcing is estimated to be 0.6 Wm^{-2} for each percent of low cloud cover change (Hartmann et al. 1992), the maximum forcing due to GCR would have a magnitude of about 0.2 Wm^{-2} . This is comparable to the forcing due to the change in TSI from minimum to maximum activity in the sunspot cycle (see section 3.2.1). Its global mean temperature response is of the order of a few hundredths of a degree.

Close passages of coronal mass ejections from the sun are signalled at the Earth's surface by Forbush decreases (FD) in cosmic ray counts. Kristjansson et al. (2008) investigated the response of clouds to these sudden decreases in the flux of GCR using cloud data from the MODIS (Moderate Resolution Imaging Spectroradiometer) satellite, which has been in operation since 2000. By focusing on pristine Southern Hemisphere ocean regions they examined areas where a cosmic ray signal should be more easily detected than elsewhere. While the studies mentioned above mainly considered cloud cover from the ISCCP, the high spatial and spectral resolution of MODIS allows for a more thorough study of microphysical parameters such as cloud droplet size, cloud water content and cloud optical depth⁵, in addition to cloud cover, the latter being of higher quality than the cloud cover data from ISCCP. In all six areas examined in this study a negative correlation was found between GCR and cloud droplet size. This is in agreement with a cosmic ray – cloud coupling, but in only one of the areas (eastern Atlantic Ocean) was the correlation statistically significant. Conversely, cloud optical depth was mostly negatively correlated with GCR, and in the eastern Atlantic Ocean that correlation was statistically significant. For cloud cover and liquid water path, the correlations with GCR were even weaker, with large variations between the different areas. When only the six Forbush decrease events with the largest amplitude (more than 10% decrease) were studied, the correlations fit the hypothesis slightly better, with 16 out of 24.

The overall conclusion by Kristjansson et al. (2008), built on a series of independent statistical tests, is that no clear cosmic ray signal (associated with Forbush decrease events) is found in highly

⁵ **Optical depth** is a measure of opacity and is defined as the negative [logarithm](#) of the fraction of [radiation](#) (or [light](#)) that is scattered or absorbed on a path.

susceptible marine low clouds over the southern hemisphere oceans. The existence of such a signal cannot be ruled out on the basis of this study, due to the small number of events and because the strongest Forbush decrease events indicate slightly higher correlations than the average events. Even though such strong events are rare, with only six events occurring over 5 years, the amplitude is similar to that during the solar cycle, so from a climate perspective these strong events may deserve some particular attention. Further investigation of a larger number of such events is needed before final conclusions can be drawn on the possible role of galactic cosmic rays in clouds and climate. Future investigations should also explore sensitivity to the selection of geographical regions for study. For instance, the recent studies by Kazil et al. (2006) and Yu et al. (2008) indicate that ion-induced aerosol nucleation may be most effective in the upper tropical troposphere. For the ongoing global warming, however, the role of galactic cosmic rays would be expected to be negligible, considering the fact that – apart from the 11-year cycle – the cosmic ray flux has not changed over the last few decades (Lockwood and Fröhlich, 2007).

Svensmark et al. (2009) found that low clouds contain less liquid water following Forbush decreases, and for the most influential events, the liquid water in the oceanic atmosphere can diminish by as much as 7%. Cloud water content as gauged by the Special Sensor Microwave/Imager (SSM/I) reaches a minimum about 7 days after the Forbush minimum in cosmic rays, as does the fraction in low clouds seen by MODIS and in ISCCP. Parallel observations by the aerosol robotic network AERONET reveal falls in the relative abundance of fine aerosol particles which, under normal circumstances, could have evolved into cloud condensation nuclei. Svensmark et al. disagree with the conclusions of Sloan and Wolfendale (2008) and Kristjansson et al. (2008) that the role of galactic cosmic rays is probably negligible. Sloan and Wolfendale (2008) used the ISCCP data, for which the error bars are particularly large. The meteorological noise can easily mask the signal of the cloud response to FD events. The error bars for MODIS are smaller, but while Svensmark et al. selected only 13 FDs in the period 2000–2007, Kristjansson et al. (2008) used about 22 FDs. As a result their data were dominated by weak FDs. Svensmark et al. thus concluded that there appears to be a link between the sun, cosmic rays, aerosols, and liquid-water clouds on a global scale.

Pierce and Adams (2009) performed calculations to determine the magnitude of the ion-aerosol clear-air mechanism using a general circulation model with online aerosol microphysics. The ‘ion-aerosol clear-air’ hypothesis suggests that increased GCR causes increases in new-particle formation, cloud condensation nuclei concentrations (CCN), and cloud cover. Years with high (solar min) and low (solar max) cosmic ray fluxes approximately corresponding to the years 1986 and 1990, respectively, were simulated using the ion-formation rates from cosmic rays based on the method of Usoskin and Kovaltsov (2006). The CCN concentrations in these tests were found to be quite insensitive to the changes in nucleation rates that occur between the solar minimum and maximum. Pierce and Adams (2009) found that at faster nucleation rates, more nm-sized particles are competing for condensable gases that can grow to CCN sizes. Each particle thus grows more slowly, keeping the particles at the smallest sizes where they have fast coagulation scavenging timescales (Pierce and Adams, 2007). Although more particles are formed giving the potential for more CCN to form, the probability of any particle surviving to CCN size is much lower. This strong negative feedback strongly dampens the sensitivity of CCN to changes in the new-particle formation rate (dampening effect). Finally, emissions of primary particles contribute a significant fraction of tropospheric CCN and do not vary with solar activity (primary emissions effect).

Taking into account the first and second indirect aerosol effect (increase in cloud reflectivity and changes in precipitation, cloud lifetime and cloud depth, respectively), the changes in cloud cover and thickness from the changes in CCN induced by clear-sky ion-induced nucleation changes during both the solar-cycle and centennial-scale changes in cosmic rays, should be in order 0.02 Wm^{-2} . This is two orders of magnitude too small to explain the changes seen in cloud cover. Based on the results of these simulations, Pierce and Adams (2009) concluded that – barring any strong biases due to model uncertainties – the ion-aerosol clear-air mechanism is too weak to explain putative correlations between cloud cover and the solar cycle. This does not rule out the potential legitimacy of a connection between cosmic rays and clouds by other physical mechanisms.

3.4 Other indicators for solar driven climate influence

3.4.1 aa index and its potential for climate change

The Earth's magnetic field shows variability on all time-scales, ranging from nanoseconds to millions of years. Most transient variations are of external origin, reflecting interactions between solar activity and the Earth's atmosphere. Longer-term changes, called secular variations, are generally attributed to changes in fluid motions within the outer core, hence of internal origin. This implies that the intensity of Galactic Cosmic Rays is influenced not only by changes in solar activity, but also by changes in the Earth's magnetic field, although differently for the various time scales. The average annual field has been closely defined by satellite observations for 1980. Previous annual patterns were based on geomagnetic observatory data, supplemented by field surveys and corrected to some particular time on the basis of observed secular variation in the preceding years. One of the geomagnetic measures used is the aa index which is calculated from magnetic field measurements of the Earth taken by two nearly antipodal magnetic observatories in England and Australia. Although variations in GCR and in the aa index both show a strong 11-year cycle, there are discrepancies in strength and phasing.

Courtillot et al. (2007) use empirical correlations between geomagnetic records and climatic proxies to propose a connection between the Earth's magnetic field, solar activity and climate. They conclude that the correlation between magnetic and geomagnetic variation, and global temperature suggests that solar influence on climate may be more significant than previously realized. However, Bard and Delaygue (2007) found many oddities in the study of Courtillot et al. (2007). Firstly, the temperature curve used looks different from the global temperature curve published by Jones et al. (1999) and updated by Brohan et al. (2006). Secondly, the reconstruction of TSI used shows much larger variations than the reconstruction by Solanki (2002) and updated by Krivova et al. (2007), resulting in an exaggeration of the decrease in TSI around 1970. Thirdly, the geomagnetic curves exhibit a large discrepancy in amplitude around 1970 compared to the aa index. The use of all these curves gives a false impression that there is a good correspondence between solar activity and climate.

Finally, as argued by Bard and Delaygue (2007), Courtillot et al. (2007) claim that there is a 'significant' correlation between cooling periods and particular increases in geomagnetic intensity (known as 'jerks'). According to these authors, this correlation supports the hypothesis of a direct link between cosmic ray flux and cloud cover (Marsh and Svensmark, 2000). However, if the causal chain proposed by Marsh and Svensmark (2000) is accepted, exactly the opposite correlation would be expected: a high geomagnetic intensity would lead to a decrease in cosmic rays and, hypothetically, a decrease in low cloud cover which, in turn, would decrease the albedo and thus increase the surface temperature. In their discussion, Courtillot et al. (2007) speculate on other hypothetical mechanisms that could reverse this chain on a local scale. Based on interpretations of biblical accounts, Courtillot et al. (2007) invoke a migration of magnetic poles to lower geographic latitudes during these geomagnetic 'jerks'.

In their paper, notably in its conclusion, Courtillot et al. (2007) express their doubts about commonly accepted facts concerning climatic evolution over the last century. This leads them to invoke geomagnetism, through its effect on cloud cover, as an additional climate driver. The climate evolution over the last century can readily be explained by a combination of natural (Sun and volcanoes) and anthropogenic forcings that became significant during the latter half of the century. Instrumental data on cosmic rays and heliomagnetic modulation do not show a long term trend that could contribute to the global warming observed over the last fifty years. So, there is still no reason to invoke the speculative forcing of Courtillot et al. (2007) based on a hypothetical link between geomagnetism, cosmic rays and cloud cover.

3.4.2 Sunspots and aa index

De Jager and Duhau (2009) undertook an empirical investigation of whether and to what degree two parameters related to solar activity may have contributed to terrestrial surface temperatures over the past four centuries (Moberg et al. 2005), using linear regression with the 21-year smoothed averages of these data. The first parameter is the number of known sunspots, or sunspot groups, from 1610 onward. These are related to the toroidal component of the Sun's magnetic field and manifest themselves mainly in the equatorial regions. The second parameter is the poloidal component of the Sun's magnetic field, which shows itself mainly in solar polar areas. Solar polar activity has various manifestations, among which are 'polar faculae' and 'brightpoints'. The maximum of this field precedes that of the sunspot number by about half a Schwabe cycle, which is the 11-year cycle of solar activity. A proxy for its maximum value is aa_{\min} , the geomagnetic aa index during sunspot minimum, which is based on direct observations from 1844 onward (Mayaud 1972, Nevalinna and Kataja 1993). Data from before that year were derived by Nagovitsyn (2006). A fundamental weakness is that these data are partly based on the sunspot numbers for that period, thus introducing some degree of correlation with the number of sunspots.

De Jager and Duhau (2009) applied a linear regression on data of sunspot number, geomagnetic index and northern hemisphere surface temperatures for the period 1844-1960. This empirical relationship was used to compare Northern Hemisphere tropospheric temperatures of the past four centuries with the expected temperatures from the two solar proxies. The residual temperatures, i.e. the difference in observed and expected temperatures from the empirical relationship, show some quasi-regular episode increases and decreases with variations of up to $\sim 0.6\text{C}$ and a periodicity in the order of a century. They conclude that the present period of global warming since 1976 is just such an episode of residual temperature increase due to internal variability in the climate system.

However, the residual temperatures of De Jager and Duhau (2009) show episodes of warming and cooling in anti phase with the smoothed sunspot number and the geomagnetic index. For instance, the dip around 1960 in the residual signal corresponds with the peaks in sunspot number and geomagnetic index, while the peaks in residual temperatures around 1660 and 1810 correspond with dips in the datasets of the solar proxies. This suggests that the fluctuations in the observed temperature can hardly be explained by the variations in the sunspot number nor by the variations in the geomagnetic index. In De Jager and Duhau's study only the trends correlate, which is true by definition. Furthermore, the amplitude of the variations in the residual temperatures are of a similar magnitude to those in the observed temperatures, making the explained variance in the temperature record of the solar signals quite weak. The conclusion reached by De Jager and Duhau (2009) that the present warming is caused by natural variability, is therefore questionable.

3.4.3 Solar flares

A flare is caused by a sudden release of energy in an active region on the sun. Flares seldom occur on or in a spot, but practically always in the vicinity of the spot. A flare starts with a burst of solar energetic particles, releasing energy and heating the surrounding plasma. This phase generally lasts for a few seconds, up to one minute or, at most, a few minutes. In the second phase, the heated plasma cools down. This lasts longer, a typical duration is some 10 minutes, but there are large variations from one flare to another. Solar flares have been observed to have effects on the Earth's middle and lower atmosphere, including the large-scale destruction of polar stratospheric and tropospheric ozone (Jackman et al. 1993, 2001; Seppala et al. 2004). However, Lockwood and Fröhlich (2007) argued that solar flares play a minor role in the solar terrestrial link for a number of reasons: "(1) the events that cause significant effects are sufficiently rare that detecting a long-term trend in their occurrence is very difficult with the limited data available, (2) observations and models show that the ozone can take up to two months to recover, but that after that there is no apparent longer-lived change induced by the transient, and (3) links between the polar middle atmospheric ozone depletion and the global surface air temperature variation are not clear."

3.5 An overview of recent solar influence

Lockwood and Fröhlich (2007) undertook a thorough review of all known or hypothesized solar parameters which are potential candidates for influencing the climate system. In particular, they studied the measurements from the last 40 years in detail to investigate whether solar variations could have played any role in the observed present-day global warming. The parameters are linked to solar variability with respect to irradiance changes in the ultraviolet (UV) spectral region as well as integrated over the spectrum (TSI), and to changes in the magnetic field as present in the vicinity of planet Earth by changes in the bombardment of cosmic rays. These three proposed mechanisms are linked by a common element, namely the magnetic field that emanates from the Sun. The field threading the solar surface modulates the TSI and UV irradiance and understanding of these effects has advanced dramatically in recent years (Krivova et al. 2003, Solanki & Krivova 2006). Hence, TSI, UV and direct GCR effects vary together and cannot be distinguished by correlative methods alone (Lockwood 2006). However, in contrast to TSI and UV, GCRs are also influenced by changes in the Earth's magnetic field (see section 3.4.1).

In the study by Lockwood and Fröhlich (2007) the monthly means of observations of various solar parameters taken since 1975 were compared with the GISS reconstruction of global mean air surface temperature (see Chapter 1), based primarily on meteorological station measurements (Hansen et al. 1999). Four parameters were investigated:

- 1) The sunspot number, which reveals the decadal-scale solar activity cycle that also dominates the variation in the other solar parameters.
- 2) The open solar flux, derived from the observed radial component of the interplanetary magnetic field (Lockwood et al. 1999).
- 3) The counts of neutrons generated by cosmic rays incident on the Earth's atmosphere, as observed at Climax.
- 4) The total solar irradiance (TSI). The TSI data are from the Physikalisch-Meteorologisches Observatorium (PMOD) TSI data composite (Fröhlich & Lean 2004) based on various space-based radiometers. Fröhlich (2006) claimed that this composite has the most rigorous set of time-dependent intercalibrations between the radiometers to account for both instrument degradations and pointing 'glitches'.

Lockwood and Fröhlich (2007) concluded from these observational data that over the past 20 years, all the trends in the Sun that could have had an influence on the Earth's climate have been in the opposite direction to that required to explain the observed rise in global mean temperatures. Therefore, the observed rise in global mean temperatures seen after 1985 cannot be ascribed to solar variability, whichever of the mechanisms is invoked and no matter how much the solar variation is amplified.

Centennial variations in solar variability are derived from proxies which are assumed to represent the TSI as well as the intensity of GCR, or both. These are sunspot number, Earth magnetic field strength (e.g. aa index) and the abundance of cosmogenic isotopes (e.g. ^{14}C and ^{10}Be). Lockwood and Fröhlich (2007) concluded from the investigated proxies since 1890 that the Sun may well have been a factor in climate change in the first half of the last century. This finding is in line with many other attribution studies using global climate models fed with reconstructions of solar activity. The extent to which the solar radiative forcing variations are amplified by some mechanism that is, as yet, unknown, is still a matter of debate, but is of minor importance to the discussion of climate change in the past two decades.

Benestad and Schmidt (2009) used a suite of global climate model simulations for the 20th century to assess the contribution of solar forcing to past trends in the global mean temperature. In particular, they examined how robust different published methodologies are in detecting and attributing solar-related climate change in the presence of intrinsic climate variability and multiple forcings. Their analysis shows that the most likely contribution from solar forcing is $7 \pm 1\%$ (or about 0.05°C) of the observed global warming for the 20th century with a negligible contribution to the warming since

1980, but that anthropogenic and solar forcing do not exclude one another. The solar-related trends over the last century are unlikely to have been greater than 0.1 to 0.2°C.

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4. Carbon Cycle

Summary

Since the IPCC AR4, there is a better understanding of the current land and ocean carbon budgets. The regional top down and bottom up estimates of the budgets are showing improved closure.

Contradictory results were often found when comparing estimates of biospheric uptake from measurements in forests and grasslands on a regional scale, with estimates inferred from atmospheric CO₂, the latter being known as the ‘inverse’ approach. The two approaches are closer together, indicating improved ability to verify the real input of GHG into the atmosphere.

Although old growth and tropical forests seem to take up more carbon than was previously thought, there are indications that the airborne fraction of CO₂ emissions is increasing, which could indicate a declining sink strength of the ocean and/or the land.

Tropical ecosystems and old growth forests in both the tropics and the Northern Hemisphere may currently be strong sinks for CO₂, but the net terrestrial uptake in the Northern Hemisphere appears to play a smaller role than was previously thought. A drop in carbon uptake of the Southern Ocean and the Atlantic has also been observed, but this could be due to long-term variability. Human-induced climate change may play a role in the Southern Ocean, where the poleward displacement and intensification of westerly winds, has enhanced the ventilation of carbon-rich waters, normally, isolated from the atmosphere, at least since 1980.

Compared to the 1990s, on a global level, natural land and ocean CO₂ sinks and land-use change emissions have remained more or less stable during the 2000-2007 period (see Table 4.1). However, emissions of fossil fuels have grown strongly and, therefore, the fraction of emissions being sequestered by the biosphere and the ocean is declining. With respect to the future, model results show a cumulative net land carbon uptake over the 21st century, but the uncertainty range remains large.

Table 4.1: The global carbon budget

	<i>1980s</i>	<i>1990s</i>	<i>2000-2005</i>	<i>2000-2007</i>
<i>Atmosphere</i>	3.3 ± 0.1	3.2 ± 0.1	4.1 ± 0.1	4.2 ± 0.04
<i>Fossil Fuels</i>	5.4 ± 0.3	6.4 ± 0.4	7.2 ± 0.3	7.5 ± 0.4
<i>Ocean</i>	-1.8 ± 0.8	-2.2 ± 0.4	-2.2 ± 0.5	-2.3 ± 0.4
<i>Biosphere</i>	-0.3 ± 0.9	-1.0 ± 0.6	-0.9 ± 0.6	-1.1 ± 0.7
<i>Sum</i>	0	0	0	-0.1
<i>Partitioning Biosphere:</i>				
<i>Land Use</i>	1.4 (0.4 to 2.3)	1.6 (0.5 to 2.7)	N/A	1.5 ± 0.5
<i>Land Sink</i>	-1.7 (-3.4 to 0.2)	-2.6 (-4.3 to -0.9)	N/A	-2.6 ± 0.7
<i>Uptake</i>	65%	75%	N/A	65%

Table 4.1. The global C-budget in the IPCC AR4 (Columns 1980s, 1990s and 2000-2005) and according to the Global Carbon Project (last column). Unit is Gt C. The last row indicates the relative uptake by the land and the ocean, compared to fossil fuel emissions. (N/A = not available.)

After a decade of no growth in methane (CH₄) concentration, in 2007 and 2008, the concentration increased again.

This is probably due to increased emissions from Northern Hemispheric sources, such as wetlands. It has been suggested that a large unknown source exists of methane from dry plants, but this has been disproved.

Large-scale fires and droughts, for instance, in Amazonia, but also at northern latitudes, can lead to significant decreases in carbon uptake.

On land, a number of major droughts and fires in middle-latitude regions, between 2002 and 2005, appear to have contributed to a decrease in the terrestrial carbon sinks in these regions, over that period. The strong non-linear relation between droughts and fires and carbon emissions and deforestation, highlights a climate-carbon feedback that may lead to higher CO₂ concentrations if droughts become more frequent in the future.

There is growing evidence that increased nitrogen availability may be responsible for part of the current terrestrial sink.

Since the industrialisation, N deposition has increased, particularly in the eastern United States, Europe, and in South and East Asia. There is growing evidence that this increase may also be responsible for maintaining at least part of the current terrestrial sink. However, recent modelling studies of coupled C and N cycles suggest that the likelihood of greatly enhanced global CO₂ sequestration resulting from future changes in N deposition, is low. Even if N emissions were to follow the more pessimistic IPCC emission scenarios, these models show a decline in the terrestrial sink, compared to models that have no N cycle.

The amount of carbon stored in permafrost areas might be up to two times larger than previously thought.

A recent study shows that accounting for C stored deep in the permafrost, more than doubles previous high-latitude inventory estimates of carbon stocks, and puts the permafrost carbon stock at an equivalent of twice the atmospheric C pool. It is also shown that the losses over several decades negate increased plant carbon uptake at rates that could make permafrost a large biospheric carbon source in a warmer world, at an amount of around 1 Pg C/year.

Preservation of existing marine ecosystems could require a CO₂ stabilisation of as low as 450 ppmv.

Recent studies review available databases on regions, ecosystems, groups of organisms, and physiological processes believed to be most vulnerable to ocean acidification. They have concluded that ocean acidification and the synergistic impacts of other anthropogenic stressors provide great potential for widespread changes to marine ecosystems, and preservation of existing marine ecosystems could require a CO₂ stabilisation level that is lower than what might be chosen based on climate considerations alone.

4.1 Introduction

There is no doubt that understanding the future behaviour of the carbon cycle is critical in making realistic mitigation plans for the near and far future (Meinshausen et al. 2009). Since the publication of the IPCC Fourth Assessment Report (Denman et al. 2007), the basic message still holds that 40 to 50% of global fossil fuel emissions is sequestered by the terrestrial biosphere and the oceans but, at the same time, the uncertainty that still exists in our understanding of key processes is worrying. How long this huge uptake will continue to operate is therefore uncertain, with some studies already suggesting that it is declining. Overall, the sparseness of the CO₂ concentration network is still the main stumbling block to improving inversion based estimates (Marquis and Tans, 2008). With the recent loss of the Orbiting Carbon Observatory this situation has not improved, although the Japanese GOSAT mission will still provide useful data. However, to be able to study interannual variability and make comparisons with inventory data, long time series of dense observations are needed.

4.2 Observations

4.2.1 CO₂ emissions and atmospheric concentration

The emissions increase as seen in the IPCC AR4 report has continued (Canadell et al. 2007). The average growth rate in the atmospheric CO₂ concentration from 2001 to 2007 was observed at 1.98 ppm/year (4.2 PgC/year), compared to 1.58 and 1.49 ppm/year in the 1980s and 1990s. Growth rates in 2007 and 2008 were even higher: 2.16 and 2.05 ppm, respectively⁶. The growth rate in the period 2000-2007 was 33% higher than that in the period 1980-2000. The average proportional growth rate of fossil fuel emissions increased from 1.3% per year for 1990–1999 to 3.3% for 2000–2006. The increase in emissions has put the current emission trajectory at the high end of the global emission scenarios developed by IPCC (2000).

Canadell et al. (2007) suggest the increase is caused by recent growth in the world economy up until the financial crisis, combined with an increase in global carbon intensity, i.e. the CO₂ emission per unit of economic activity. The latter is related to the increasing contribution to overall emissions by the strongest growing economies, India and China, both having above average carbon intensity levels (Raupach et al. 2007). Van Vuuren and Riahi (2008) suggest that there may be reasons to assume that the rise in emissions in the longer term may be less rapid than in recent years. These relate to the quality of emissions and other data for the period of 2000-2006, and in particular to the major uncertainties with respect to the underlying driving forces, including the future development of China, global oil prices, and the long-term economic prospects. For instance, it is likely that the financial crisis might result in a temporarily lower increase in CO₂ emissions, not least because the impact of the crisis on heavy industry has been unevenly strong. They also indicate that scenarios will always have a limited ‘lifetime’. However the increase in growth rate has several implications for policy negotiations, but these primarily relate to the very uneven regional development of the emissions (e.g. Raupach et al. 2007) and the difficulty of negotiating a package which is also fair to fast-growing economies.

Compared to the 1990s, at global level, natural land and ocean CO₂ sinks and land use change emissions have remained stable during the period 2000-2007 (see Table 4.1). However, emissions of fossil fuels have grown strongly and therefore the fraction of emissions being sequestered by the biosphere and the ocean is declining. With respect to the future, model results show a cumulative net land carbon uptake over the 21st century, but the uncertainty range remains large.

⁶ See <http://www.esrl.noaa.gov/gmd/ccgg/trends/>

4.2.2 Methane emissions and atmospheric concentrations

Rigby et al. (2008) show that after a decade of no growth in atmospheric methane (CH_4) concentration, in 2007 and 2008 the concentration increased again. They attribute this increase to increased emissions from Northern Hemisphere sources, primarily wetlands. The idea that there exists a large, hitherto unknown, source of methane from dry plant origin (Keppler et al. 2006) was disproved by Duek et al. (2007). They showed conclusively that there is no evidence for substantial aerobic methane emission by terrestrial plants. They put this at maximally 0.3% of the previously published values. Additional evidence for a strong reduction in the potential source from tropical plants as originally suggested, comes from Frankenberg et al. (2008). They use inversions based on SCIAMACHY instrument retrievals and reduce the total (not only the dry part) annual tropical emission from 260 to about 201 Tg CH_4 . The active contribution of terrestrial plants to global methane emission is thus very small at best, making it unnecessary to take this into account in Kyoto or post-Kyoto land use emissions.

Paleoclimatology studies (Petrenko et al. 2008, Fisher et al. 2008) suggest that the last glacial termination was likely to have been caused more by increasing methane emission from boreal wetlands than emissions stemming from methane hydrate instability. There are also, as yet unpublished, observations of a substantial increase in methane emissions from beneath the permafrost of the east Siberian shelf. How much these could contribute to the observed rise in Northern Hemispheric emissions is unknown. Westbrook et al. (2007) suggest that increasing sea temperatures along continental margins have increased the release of CH_4 . How much of this released methane actually enters the atmosphere remains uncertain.

4.2.3 Airborne fraction of CO_2 emissions

While both land and ocean continue to accumulate carbon at an average rate of $5 (\pm 0.6)$ PgC/year, Canadell et al. (2007) suggest evidence for a long-term (50year) increase in the airborne fraction (AF) of CO_2 emissions. This would imply a decline in the efficiency of the CO_2 land and ocean sinks in absorbing anthropogenic emissions. If confirmed, the increase in the AF implies carbon emissions have grown faster than the CO_2 sinks. Because the land and oceans are both mosaics of regions that are gaining and regions that are losing carbon, this trend could arise from sink regions having weakened, either absolutely or relative to growing emissions; source regions could have intensified or lastly, sink regions could even have changed into source regions. On the other hand, since the atmospheric fraction also shows strong interannual variability, caused by the effects of El Niños and volcanic eruptions, for example, the suggested increase is not (or not yet) statistically significant. Furthermore, the trend in AF depends on the deforestation flux, which is highly uncertain (van der Werf et al. 2009) and may need downward adjustment to values as low as 1.2 Gton C/year over the period 1997–2006, with a quarter on top of that (0.3 Gton C/year) coming from peat land degradation and peat fires in South East Asia. Satellite-based estimates of this flux are generally lower than estimates based on bookkeeping methods, and using the lower estimates results in a more constant AF over time.

4.2.4 Current CO_2 sink strength of the biosphere

Contradictory results were often found when comparing estimates of biospheric uptake from measurements in forests and grasslands the regional scale with estimates inferred from atmospheric CO_2 , the latter known as the ‘inverse’ approach⁷. To get these different approaches to converge, care must be taken to account for all the components of carbon exchange between the land and the

⁷ Our ability to diagnose the fate of anthropogenic carbon emissions depends critically on interpreting spatial and temporal gradients of atmospheric CO_2 . The state-of-the-art methodology to achieve this is to back-calculate from observed concentrations using a transport model, the areas where the sources or sinks are located. This technique is called inverse modelling.

atmosphere, and the movement of carbon in and out of the region in question (e.g. Dolman et al. 2008). This requires exhaustively cataloguing data from all of the major vegetation and soil categories in a region through the use of inventories, field measurements, surveys and remote sensing. In addition to this 'bottom-up' estimate, process-based ecosystem models and ensembles of atmospheric inversions are needed to arrive at a convergent estimate. Piao et al. (2009) show how this methodology is now mature enough to confidently estimate the carbon balance of large regions like China. They estimate a net carbon sink in the range of 0.19–0.26 PgC/year, smaller than that in the conterminous United States but comparable to that in geographic Europe. This was sufficient to offset 28–37% of China's fossil fuel emissions in the 1980s and 1990s.

Recent estimates of European carbon dioxide, methane and nitrous oxide fluxes between 2000 and 2005 using both top-down estimates, based on atmospheric observations and bottom-up estimates derived from ground-based measurements, are given by Schulze et al. (2009). They show that both methods yield similar fluxes of greenhouse gases, suggesting that methane emissions from feedstock and nitrous oxide emissions from arable agriculture are fully compensated for by the carbon dioxide sink provided by forests and grasslands. As a result, the balance for all greenhouse gases across Europe's terrestrial biosphere is near neutral, despite carbon sequestration in forests and grasslands. The trend towards more intensive agriculture and logging may make Europe's land surface a significant source of greenhouse gases (Schulze et al. 2009).

Luysaert et al. (2008) report that old-growth forests can continue to accumulate carbon, in contrast to the long-standing view that they are carbon neutral. This would apply to over 30% of the global forest area that is still unmanaged primary forest. Half of the primary forests (600 million ha.) are located in the boreal and temperate regions of the Northern Hemisphere. These forests alone sequester about 1.3 ± 0.5 PgC/year. These findings suggest that 15% of the global forest area, which is currently not considered when offsetting increasing atmospheric CO₂ concentrations, provides at least 10% of the global net ecosystem productivity. In Europe conservative forest management, with a steady harvest rate, has not kept pace with increasing woody biomass production, thus creating a net sink (Ciais, et al. 2008).

Over recent decades, long-term monitoring plots across Amazonia (Phillips et al. 2008) show that remaining old-growth forest trees are a sink of 0.62 ± 0.23 MgC/ha/year, amounting to 0.5–0.8 PgC/year for the whole Amazonian region. Stephens et al. (2007) reveal annual-mean vertical CO₂ gradients, using global measurements of vertical atmospheric CO₂ distributions, that are inconsistent with a set of atmospheric inverse models showing a large transfer of terrestrial carbon from tropical to northern latitudes. This suggests these models may have overestimated the contribution of northern latitude sinks compared to tropical ones. The three inverse models that most closely reproduce the observed annual-mean vertical CO₂ gradients estimate weaker northern uptake of 1.5 PgC/year and weaker net tropical emissions (i.e. uptake minus deforestation loss) of 0.1 PgC/year compared with previous estimates of an uptake of 2.4 and net emissions of 1.8 PgC/year, respectively. This suggests northern terrestrial uptake of CO₂ plays a smaller role than previously thought and that, after subtracting land-use emissions, tropical ecosystems may currently be strong sinks for CO₂. It would appear that bottom-up results from biomass inventories and modelling support this view.

Data from a ten-country network of long-term monitoring plots in African tropical forests (Lewis et al. 2009) suggests above-ground carbon storage in live trees increased by about 0.6 MgC/ha/year between 1968 and 2007. Extrapolation to unmeasured forest components (live roots, small trees, necromass) and scaling to the continent, implies a total increase in carbon storage in African tropical forest trees of 0.34 PgC/year. These reported changes in carbon storage are similar to those reported for Amazonian forests per unit area, providing evidence that increasing carbon storage in old-growth forests is a pan-tropical phenomenon. Indeed, combining all standardized inventory data from this study and from tropical America and Asia together yields a comparable figure of 0.49 MgC/ha/year over the period 1987–1997. This indicates a carbon sink of 1.3 PgC/year (0.8–1.6) across all tropical forests during recent decades. This corroborates the estimate of Stephens et al. (2007) discussed above. Changes in resource availability, such as increasing atmospheric carbon dioxide concentrations, may be the cause

of this increase in carbon stock, which is in line with modelling studies (Sitch et al. 2008). Table 4.2 summarizes the C-uptake estimates mentioned in this paragraph.

Mercado et al. (2009) present modelling evidence that a late-20th century increase in diffuse radiation may have enhanced the terrestrial C sink by about 25%. Reducing future anthropogenic pollution, largely aerosols, may thus create a positive feedback to global warming, because of the decrease in associated diffuse radiation.

Piao et al. (2009) show that temperature changes before the 1970s had a limited influence on the current pattern of net carbon sequestration in the boreal zones, and that the impact of temperature changes within the last decade are not strong enough to be observable. Changes in the seasonal patterns of temperature, however, are one of the important drivers of today's forest carbon balance in the Northern Hemisphere. Over long time scales, our ability to accurately model carbon fluxes is thus partly constrained by our limited understanding of how seasonal temperature may vary over time.

Region	Estimated uptake (PgC/Year)	Technique
<i>Africa</i>	<i>0.34</i>	<i>Biomass Inventory</i>
<i>Amazon</i>	<i>0.5-0.8</i>	<i>Biomass Inventory</i>
<i>Pan tropical (incl. Asia)</i>	<i>0.8-1.6</i>	<i>Biomass Inventory</i>
<i>Pan tropical</i>	<i>0.1 (forest uptake – deforestation)</i>	<i>Inversions</i>
<i>Northern Hemisphere old-growth forests</i>	<i>1.3</i>	<i>Biomass Inventory</i>
<i>European forests</i>	<i>0.2</i>	<i>Biomass Inventory</i>

Table 4.2 Recent estimates of C-uptake in the tropics and Northern Hemisphere. Literature cited in text.

4.2.5 Current CO₂ sink strength of the ocean

Gruber et al. (2009) synthesize estimates of the contemporary net air-sea CO₂ flux on the basis of an inversion of interior ocean carbon observations using a suite of ten ocean general circulation models (Mikaloff Fletcher et al. 2007) and compare them to estimates based on a new climatology of the air-sea difference of the partial pressure of CO₂ (pCO₂) (Takahashi et al. 2008). These two independent flux estimates yield a consistent description of the regional distribution of annual mean oceanic sources and sinks of atmospheric CO₂ for the decade spanning the 1990s and early 2000s with differences at the regional level of generally less than 0.1 PgC/year. This mirrors similar progress made in narrowing down uncertainties between models and observations on the land.

The ocean distribution of fluxes obtained by inverse models is characterized by outgassing in the tropics, uptake in mid-latitudes, and comparatively small fluxes in the high latitudes. Both estimates point towards a relatively small (0.3 PgC/year) contemporary CO₂ sink in the Southern Ocean (south of 44°S), a result of the near cancellation between a substantial outgassing of natural CO₂ and a strong uptake of anthropogenic CO₂. In the Southern Ocean the ocean inversion suggests a relatively uniform uptake, while the pCO₂-based estimate suggests strong uptake in the region between 58°S and 44°S, and a source in the region south of 58°S. Globally, and for 1995 to 2000, the contemporary net air-sea flux of CO₂ is estimated at 1.7 ± 0.4 PgC/year for the inversion and at 1.4 ± 0.7 PgC/year based on the pCO₂-climatology, respectively. This is composed of an outgassing flux of river-derived carbon of 0.5 PgC/year, and an uptake flux of anthropogenic carbon of 2.2 ± 0.3 PgC/year (inversion) and 1.9 ± 0.7 PgC/year (pCO₂-climatology). While the convergence of these two independent estimates is encouraging and suggests that it is now possible, as on land, to provide relatively tight constraints for net air-sea CO₂ fluxes on a regional basis, both studies are limited by their lack of consideration of

long-term changes in the ocean carbon cycle, such as the recent possible stalling in the expected growth of the Southern Ocean and Atlantic Ocean carbon sink.

In the Southern Ocean, the poleward displacement and intensification of westerly winds, probably caused by a combination of stratospheric ozone losses, greenhouse gas increases, and natural variability (see chapter 1), has enhanced the ventilation of carbon-rich waters normally isolated from the atmosphere, at least since 1980 (le Quere et al. 2007, le Quere et al. 2008), but it remains probably too early to be able to robustly attribute this to a pertinent change in the circulation or to multi-decadal or interannual variability (Zickfeld et al. 2008).

In the North Atlantic, Schuster and Watson (2007) show that the sink for atmospheric CO₂ exhibits significant interannual variability. They also observe an interdecadal decline, evident throughout the North Atlantic, especially significant in the northeast of the area but excluding the western subtropical areas. The overall sink reduced by more than 50% from the mid-1990s to the period 2002–2005. More precisely, they estimated that the uptake of the region between 20°N and 65°N declined by 0.24 PgC/year from 1994–1995 to 2002–2005. A key question is whether this decline is due to a re-arrangement of the ocean part of the global carbon cycle in response to climate change or part of ‘natural’ fluctuations in the basin caused by the North Atlantic Oscillation (e.g. Gruber 2009).

4.3 Vulnerability: factors influencing the carbon exchange with the biosphere

4.3.1 Droughts and fires

On land, a number of major droughts in mid-latitude regions in 2002–2005 appear to have contributed to a decrease in the terrestrial carbon sinks in these regions over that period. Peters et al. (2007), using the NOAA advanced Carbon Tracker assimilation scheme, find that terrestrial uptake in the US fell to 0.32 PgC/year during the large-scale drought of 2002, compared to an average uptake of 0.62 PgC/year over the period 2001–2005. This was also found by Ciais et al. (2005) who found a reduction in European uptake in the hot, dry summer of 2003 that offset four previous years of uptake. The sensitivity to climate is highly non-linear and there is large potential for extreme climate events to significantly disturb annual long-term average sink behaviour.

Large uncertainties are associated with the response of tropical vegetation to drought, and boreal ecosystems to elevated temperatures and changing soil moisture status. Phillips et al. (2009) used records from multiple long-term monitoring plots across Amazonia to assess forest responses to the intense 2005 Amazon drought. Affected forest lost biomass, and this reversed a large long-term carbon sink, with the greatest impacts observed where the dry season was unusually intense. The drought had a total biomass carbon impact of 1.2 to 1.6 PgC. Amazon forests therefore appear to be particularly vulnerable to increasing moisture stress, with the potential for large carbon losses to exert feedback on climate change.

The fastest way drought can lead to carbon losses (or, to a reduction in the sink strength) is through fire. Fires were recently identified as one of the main features lacking in global models, even though (deforestation) fires not only influence CO₂ concentrations but also seven other radiative forcing components (Bowman et al. 2009). According to Nepstad et al. (2008) the greater vulnerability of Amazon forests due to drought may expose more than 50% of the Amazon to fires and logging. In Indonesia where carbon-rich peat soils are burned during the dry season, fires have been known to be a large emission source. Only recently reliable estimates of emissions have been made, turning out to be comparable to Indonesian fossil fuel emissions (van der Werf et al. 2008). Moreover, these fires have been amplified by human interference and are strongly dependent on seasonal droughts governed by El Niño or positive phases of the Indian Ocean Dipole (Field et al. 2009). The strong non-linear relationship between droughts and fires versus carbon emissions and deforestation, highlights a climate-carbon feedback that may lead to higher CO₂ concentrations if droughts become more frequent

in the future (Li et al. 2007).

4.3.2 Interaction with the nitrogen cycle

Since industrialization, N-deposition has increased, particularly in the Eastern US, Europe, and in South and East Asia. There is growing evidence that this increase may also be responsible (Magnani, et al. 2007, de Vries et al. 2008, Reay, et al. 2008, Janssen and Luysaert, 2009) for maintaining at least part of the current terrestrial sink. The estimates of C sink strength per deposited kg of nitrogen vary from 70 to 400 (Magnani, et al. 2007) kg C per kg N deposition, depending on the estimate of the total N deposition. Due to an increase in fossil fuel combustion for energy, industry and transport, and of fertiliser and manure use for food and biomass production, the deposition rates are likely to become even larger and more widely spread (Galloway et al. 2008). The acceleration in the natural N-cycle as a consequence of global warming is another mechanism by which N availability is increased to the vegetation. Several authors have evaluated the effect of including C-N coupling in carbon and/or climate models (Sokolov et al. 2008, Xu-Ri and Prentice, 2008). In response to a 2°C warming, simulated Net Primary Production (NPP)⁸ decreased when the C and N cycles were uncoupled, due to higher plant respiration. But simulated NPP increased when they were coupled, because the higher decomposition rates and associated N availability offset the cost of plant respiration (Sokolov et al. 2008).

The limited number of studies that have been carried out with full CX-N coupling suggests that global integrated responses of net land carbon exchange to variation in temperature and precipitation are significantly damped by the carbon-nitrogen cycle coupling (Thornton et al. 2007). These modelling studies of coupled C and N cycles suggest that the likelihood of greatly enhanced global CO₂ sequestration resulting from future changes in N deposition is small. Even if N emissions were to follow the more pessimistic IPCC emissions scenarios, these models show a decline in the terrestrial sink compared to models that have no N-cycle. In general, several processes are not yet fully integrated into the land surface and vegetation models, including the effect of climate change on the C-N coupling, influence of anthropogenic N deposition, biological nitrogen fixation, re-allocation of nitrogen between vegetation and soil and between labile and recalcitrant pools, and the redistribution of plant species in response to disturbance or climate change (Bonan 2008).

A recent review indicates that the likelihood of greatly enhanced global CO₂ sequestration resulting from future changes in N deposition is small (Reay et al. 2008). Even if N emissions were to follow the more pessimistic IPCC emissions scenarios and large increases in the strength of the terrestrial and oceanic carbon sinks were achieved, this may be offset by any simultaneous enhancement of N₂O emissions. A doubling of the year 2000 N emissions by 2030 may achieve 3 Pg of additional CO₂ sequestration in northern and tropical forests per year, but it would also induce global annual emissions of between 0.54 and 2.7 Pg of CO₂ equivalent, in the form of N₂O via increased nitrification and denitrification on land and in the oceans. Depending on the precise value of the N₂O emission rates, some pollution swapping could greatly offset any net climate change mitigation benefits.

4.3.3. Permafrost decomposition

Schuur et al. (2008) show that accounting for C stored deep in the permafrost more than doubles previous high-latitude inventory estimates of carbon stocks. This new estimate puts the permafrost carbon stock to an equivalent of twice the atmospheric C pool. Thawing of permafrost with warming occurs both gradually and catastrophically, exposing organic C to microbial decomposition. Growing season length, plant growth rates and species composition and ecosystem energy exchange may

⁸ NPP is the rate at which plants in an ecosystem produce net useful chemical energy. It is equal to the difference between the rate at which the plants in an ecosystem produce useful chemical energy (GPP) and the rate at which they use some of that energy through cellular respiration.

balance these losses to some extent. However, these processes do not appear to be able to compensate for C release from thawing permafrost. Schuur et al. (2009) find that areas that thawed over the past 15 years had 40% more annual losses of old carbon than less thawed areas, but had an overall net ecosystem carbon uptake as increased plant growth offset these losses. In contrast, areas that thawed decades earlier lost even more old carbon, a 78% increase over the less thawed areas. This old carbon loss contributed to overall net ecosystem carbon release despite increased plant growth. They conclude that over decadal timescales, the losses overwhelm increased plant carbon uptake at rates that could make permafrost a large biospheric carbon source in the order of 1 Pg C/year in a warmer world, comparable to roughly half the current global biospheric uptake. Dorrepaal et al. (2009), using data from long term ecosystem manipulation experiments, find that respiration rates increased through the warming of sub-arctic soil, particularly in the layer from 25-50 cm, by almost 70%. These effects were sustained for eight years. If these results can be extrapolated globally, an additional 38-100 Pg C could be released per degree of warming.

4.3.4 Ocean acidification

Oceanic uptake of anthropogenic carbon dioxide (CO₂) is altering the seawater chemistry of the world's oceans with consequences for marine biota. Elevated partial pressure of CO₂ is causing the calcium carbonate saturation horizon to shoal in many regions, particularly in high latitudes and regions that intersect with pronounced hypoxic zones, as a result of increased nitrogen run-off and deposition. The ability of marine animals, most importantly pteropod molluscs, foraminifera, and some benthic invertebrates, to produce calcareous skeletal structures is directly affected by seawater CO₂ chemistry. CO₂ influences the physiology of marine organisms as well through acid-base imbalance and reduced oxygen transport capacity. The few studies at relevant pCO₂ levels limit our ability to predict future impacts on food web dynamics and other ecosystem processes. Fabry et al. (2008) present new observations and review available databases on regions, ecosystems, taxa, and the physiological processes believed to be most vulnerable to ocean acidification. They conclude that ocean acidification and the synergistic impacts of other anthropogenic stressors have great potential to cause widespread changes in marine ecosystems.

Silverman et al. (2009) show that by the time atmospheric partial pressure of CO₂ reaches 560 ppm, all coral reefs (more than 9000 reef locations) will cease to grow and start to dissolve as result of ocean acidification. Cao and Caldeira (2008), using a coupled climate-carbon model, similarly conclude that ocean chemistry will be significantly disturbed even if atmospheric CO₂ can be stabilized at low to modest levels. At stabilization levels as low as 450 ppm, some parts of the high latitude ocean would become undersaturated with respect to aragonite and experience a drop in pH by more than 0.2 units (i.e. an almost 60% increase in acid particles). They find that at stabilization levels of 550 ppm, there will be no water left in the open ocean with the kind of chemistry (aragonite saturation levels) experienced by more than 98% of shallow-water coral reefs before the advent of the industrial revolution. As also stated in the Monaco Declaration in 2008 that was signed by 155 scientist from 26 nations, Silverman et al. suggest that preservation of existing marine ecosystems could require a CO₂ stabilization level that is lower than what might be chosen based on climate considerations alone.

4.4 Projections of carbon removal

Since the IPCC 4AR report, the carbon cycle feedback has received much attention and modelling groups have been active in incorporating this feedback into their models (e.g. see Boer and Arora, 2009, for an example and analysis of this feedback). However, since the range in carbon sensitivity is large between the models used in the IPCC 2007 report (Denman et al. 2007), it remains important to understand how well our land and atmosphere models capture all relevant processes. Intercomparison studies are critical in this respect.

4.4.1 Terrestrial carbon uptake

In a recent modelling study, probably one of the most comprehensive undertaken so far, (Sitch et al. 2008) five Dynamic Global Vegetation Models (DGVMs)⁹ were compared with global land carbon budgets for the 1980s and 1990s. Their results show that all DGVMs are consistent with these global budgets and in agreement with earlier modelling studies on cumulative land uptake over the last 50 years (e.g. Friedlingstein et al. 2006). However, this study went beyond earlier intercomparisons by including and diagnosing climate-carbon cycle feedbacks, and by spanning a wider range of emission scenarios. The five models were run with observed climatologies and atmospheric CO₂ and with future values derived from four IPCC scenarios (IPCC, 2000) and from a simple climate model and ocean carbon cycle model. The DGVMs are able to simulate the correct sign of the global land carbon response to ENSO but, importantly, still with differing magnitudes of response. Randerson et al. (2009) also develop a consistent strategy for assessing DGVMs compared with metrics derived from a variety of observations. They propose establishing an open source, community-wide platform for model-data intercomparison to speed up model development and to strengthen ties between the modelling and measurement communities.

All models show a cumulative net uptake in the 21st century, but land uptake varies markedly between DGVMs. This indicates large uncertainties in future atmospheric CO₂ concentrations associated with uncertainties in the terrestrial biosphere response to changing climatic conditions. It is likely that part of this uncertainty is caused by poor and divergent representation of the interactions with nitrogen (see section 4.3.2). The DGVMs show more divergence in their response to regional changes in climate than to increases in atmospheric CO₂ content. All the models simulate a release of land carbon in response to climate when the physiological effects of elevated atmospheric CO₂ on plant production are not taken into account, thereby implying a positive terrestrial climate-carbon cycle feedback. All DGVMs simulate a reduction in global Net Primary Production (NPP)³ and a decrease in soil residence time in the tropics and extratropics in response to future climate. When both counteracting effects of climate and atmospheric CO₂ on ecosystem function are considered, all the DGVMs simulate cumulative net land carbon uptake over the 21st century for the four SRES emission scenarios. However, for the most extreme A1FI emissions scenario, three out of the five DGVMs project an annual net source of CO₂ from the land to the atmosphere in the final decades of the 21st century. For this scenario, cumulative land uptake differs between DGVMs by 494 PgC: equivalent to over 50 years of anthropogenic emissions at current levels.

4.4.2. Oceanic carbon uptake

Previously ocean models suggested that the interannual variability in the global sink is relatively small, while new estimates based on atmospheric inversions (see section 4.1.4) indicate a year-to-year variability that is substantial. Projections (Denman et al. 2007) suggest that the sink should increase as atmospheric CO₂ continues to rise, but that under anthropogenically induced climate change, increasing stratification¹⁰ and a slowing overturning circulation, rates of ventilation decrease which tends to slow the uptake.

Thomas et al. (2008) propose that the interpretation of a reduced sink may need to be viewed with great caution. Using a global model of the ocean carbon cycle, forced with observed atmospheric conditions they investigated how variations and trends in the North Atlantic Oscillation (NAO)¹¹ have

⁹ A DGVM simulates changes in potential vegetation and associated bio-geochemical and hydrological cycles in response to climate shifts.

¹⁰ Water stratification occurs when water of high and low salinity ([halocline](#)), as well as cold and warm water ([thermocline](#)), form layers that act as barriers to water mixing.

¹¹ The NAO is a [climatic](#) phenomenon in the North [Atlantic Ocean](#) of fluctuations in the difference of atmospheric [pressure at sea-level](#) between the [Icelandic Low](#) and the [Azores high](#). Through east-west oscillation motions of the [Icelandic Low](#) and the [Azores high](#), the NAO controls the strength and direction of westerly [winds](#) and [storm](#) tracks across the North Atlantic.

affected the North Atlantic carbon sink since 1979. They argue that the observed trends reflect fluctuations on a decadal timescale that are a response to climate variability in the North Atlantic region and find substantial year-to-year changes in the surface-ocean carbon cycle, with an overall tendency of increased uptake in the temperate North Atlantic.

The Southern Ocean stands out with its large uptake of anthropogenic CO₂ and its substantial outgassing of natural CO₂. This region reacts with relatively high sensitivity to climate variations over the last 50 years as well as to future climate change, because of the interaction of sea-ice, upwelling, and convection with changes in heat and freshwater. Given the large exchange fluxes of the two CO₂ components, relatively small changes in the Southern Ocean can lead to large changes in the net atmosphere-ocean balance of CO₂ (Gruber et al. 2009, le Quere et al. 2008, 2009).

Wilson et al. (2009) suggest that marine fish contribute 3 to 15% of total oceanic carbonate production. Importantly, they suggest that fish carbonate production may rise in response to future CO₂, and thus become an important component of the C cycle.

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5. Climate sensitivity and feedbacks

Summary

Recent studies that use information in a relatively complete manner, generally confirm the likely range (66%) of the climate sensitivity to be between 2 and 4.5 °C, as indicated in the IPCC AR4 report.

Since the IPCC AR4 report, the discussion concerning climate sensitivity has focused on how to merge the constraints of various observations and how to deal with the uncertainty in total radiative forcing, mainly due to aerosol and solar forcing. With respect to combining different lines of evidence, the dependency of the methods used is crucial in determining the likely range of climate sensitivity. Climate sensitivity may also be time-dependent or state-dependent due to the fact that some feedbacks do not scale linearly with temperature. For example, in a much warmer world with little snow and ice, the ice surface albedo feedback would be different from that of today.

Climate sensitivity

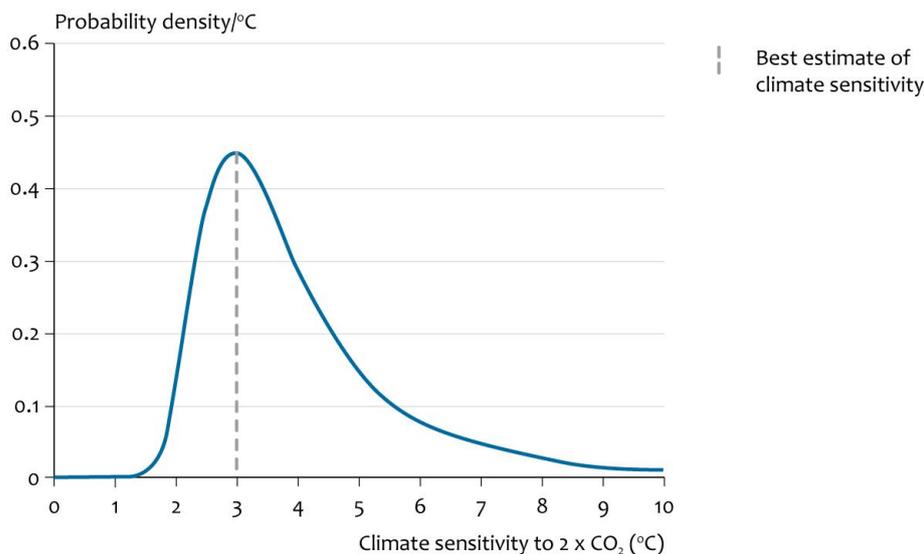


Figure 5.1 Stylised probable distribution of climate sensitivity, based on the analysis of Roe and Baker (*Why is climate sensitivity so unpredictable?* *Science*, 318, 629–632, 2007). The best estimate of climate sensitivity in this analysis is a rise of 3 °C due to a doubling of the CO₂ concentration compared to the pre-industrial level. As the graph shows, the area to the left of the dotted line is smaller than the area to the right. In other words, the probability that the climate sensitivity is underestimated, compared to the best estimate of 3 °C, is greater than that the sensitivity is overestimated, due to the skewed form of the distribution.

While a climate sensitivity below 1.5 °C is physically extremely unlikely, high values exceeding 6 °C cannot be ruled out by the comparison of models with observations.

Due to the essentially skewed distribution, the probability that climate sensitivity is underestimated, compared to the best estimate of 3 °C, is greater than that the sensitivity is overestimated (see Figure 5.1). In general, an upper bound of climate sensitivity is difficult to assess, but studies rarely assign a high probability to values in excess of 10 °C and they generally point to a maximum likelihood value of well within the consensus range of IPCC AR4. Studies that suggest very low climate sensitivity values, use erroneous forcing assumptions, neglect internal climate variability, use overly simplified

assumptions, neglect uncertainties, or have other errors in the analysis or dataset, or a combination of these.

Slow feedbacks may be insufficiently incorporated in climate models. Paleo data suggest that by including these feedbacks, the climate system may be twice as sensitive as our present ‘best estimate’ in the very long term, i.e. the millennium timescale and beyond.

Paleo data (i.e. data covering the past millions of years) suggest that equilibrium sensitivity, including slower feedbacks, such as ice albedo, vegetation migration and the release of greenhouse gases from soils, tundra or ocean sediments, is ~ 6 °C for doubled CO₂ for the range of climate states between glacial conditions and an ice-free Antarctica. However, the determination of climate sensitivity from these data is based on the assumption that all additional changes in temperature are linearly related to CO₂ changes, implying constant feedback factors independent of the climate state (for example, the amount of ice on Earth). It is also hypothesised that the strength of fast feedbacks, such as atmospheric water vapour content, lapse rate (i.e. the temperature decrease with the altitude in the lower part of the atmosphere), and clouds, decreased during the last glacial maximum compared to present conditions. In addition, it is known from ice core data that the dust (aerosol) amount is higher in glacial conditions, creating an additional cooling. Uncertainty in this cooling effect hampers accurate estimates of the climate sensitivity. Either a large aerosol forcing or weaker fast feedbacks would imply a reduction in climate sensitivity, i.e. less than twice the best estimate in the IPCC AR4 report. Understanding the role of aerosols either as a forcing or as a feedback factor is crucial in reducing the uncertainty in climate sensitivity derived from observations.

5.1 Introduction

Radiative forcing¹² and climate sensitivity¹³ are two key factors in understanding the Earth's climate. There is considerable interest in decreasing the present uncertainty in climate sensitivity in order to narrow future projections of global average temperature changes. Climate sensitivity as well as radiative forcing cannot be measured directly, but they can be estimated from comprehensive climate models and calculated using radiative transfer models, respectively.

Given the conservation of energy, the general temperature structure of the atmosphere and the well-known radiative properties of gases in our atmosphere, in theory, we can derive the relationship between temperature response and radiative forcing. In an equilibrium state – still allowing for natural variability and climate fluctuations – this will be an approximately linear relationship. The ratio of the equilibrium global mean surface temperature change to the global mean imposed radiative forcing is called the *climate sensitivity parameter*, and expressed in units of KW^{-1}m^2 . *Climate sensitivity* refers to the eventual global mean surface temperature response due to a doubling of the CO_2 concentration in the atmosphere, expressed in units of K. The radiative forcing due to a doubling of the CO_2 concentration is $3.7 \pm 0.4 \text{ Wm}^{-2}$. Climate sensitivity can be obtained by multiplying the climate sensitivity parameter by this radiative forcing value.

Despite this linear relationship between equilibrium temperature response and imposed radiative forcing, the value of the climate sensitivity will depend on a myriad of feedbacks. The response to perturbations in the radiation balance can be either amplified or damped as the result of temperature-dependent processes in the climate system. These climate feedbacks are predominantly found in the hydrological cycle due to the large amounts of water on our planet combined with its strong impact on the energy balance in all its phases (ice, liquid water and water vapour). The ensemble of climate feedbacks thus determines the climate sensitivity.

There is discussion about the climate feedbacks incorporated in climate simulations for the 21st century. Feedbacks act over a wide variety of timescales. The primary feedbacks involving water vapour and clouds occur on timescales of days. This is much faster than the response timescale of reaching an equilibrium global temperature (within the natural variability) after applying a constant radiative forcing. The time lag of several decades between forcing and response emanates from the large heat capacity of the oceans. Therefore, the primary feedbacks are fully accounted for in the response. However, slower feedbacks, such as ice albedo, ocean circulation changes and biological feedbacks, may be partly incorporated in climate models, since validation against observations is hardly feasible for the feedbacks operating on a millennium timescale.

The delay in response also implies that under the condition of ever changing forcing agents, climate sensitivity cannot be determined by using temperature and forcing information alone, but also involves understanding of the amount of heat uptake by the oceans. This in turn, requires insight into the partition between the heat storage in the oceans and the energy directly rendered into the atmosphere through either longwave radiation¹⁴ or latent¹⁵ and sensible¹⁶ heat flows. This partition is dependent on

¹² **Radiative forcing** is defined as the change in net [irradiance](#) at the top of the atmosphere. 'Net irradiance' is the difference between the incoming radiation energy and the outgoing radiation energy in a given climate system and is thus measured in Watts per square metre.

¹³ **Equilibrium climate sensitivity** refers to the equilibrium change in global mean near-surface air temperature that would result from a sustained doubling of the atmospheric (equivalent) CO_2 concentration compared to pre-industrial times. This value is likely to be in the range of 2 to 4.5°C with a best estimate of about 3°C .

¹⁴ **Outgoing Longwave Radiation** (OLR) is the energy leaving the earth as infrared radiation at low [energy](#).

¹⁵ **Latent heat flux** is the flux of heat from the Earth's surface to the atmosphere that is associated with evaporation or transpiration of water at the surface and subsequent condensation of water vapour in the troposphere.

¹⁶ **Sensible heat flux** is the flux of energy per square metre ($\text{Js}^{-1}\text{m}^{-2}$) in heating a surface without evaporating a liquid from it.

the thermal and salinity structure of the oceans, regulating heat diffusion as well as the upwelling processes in the ocean.

Evaluation of the components of the energy balance is only partly feasible. Satellite observations of incoming and outgoing radiation are only available for a about three decades. Datasets of ocean heat uptake in the top 700 m over the last 50 years have been updated recently (Domingues et al. 2008, Ishii and Kimoto 2009, Levitus et al. 2009, see also section 2.2.3.). However, assumptions for the energy flows towards the deeper ocean layers have to be made for the determination of the total heat uptake (Murphy et al. 2009).

Two approaches may be distinguished to derive the range of possible values of climate sensitivity: firstly, the bottom-up approach using climate models to determine the radiative forcing and the temperature response, from which climate sensitivity can be calculated. Secondly, the top down approach using observations of temperature and estimates of the radiative forcing to derive the range of possible values for climate sensitivity.

Bottom-up approach using models

Climate models include the known individual climate processes allowing for a calculation of the total feedback and thus the climate sensitivity. The quality of the climate models is tested using measurements. The calculated climate patterns are also validated against observations. It should, however, be borne in mind that climate models may produce the right numbers for the wrong reasons. However, there is some confidence in climate models as they reproduce the climate at least on the continental scale and in a statistical sense, given the wide range of climate parameters such as temperature and humidity in the present climate. The advantage of using models is that they can be brought into equilibrium, that a variety of forcings can be imposed separately and that the strength of the feedbacks can be varied (e.g. Murphy et al. 2004).

Top-down approach using observations

Direct and indirect measurements of temperature and forcing factors – converted into estimates of radiative forcing – can be used to estimate climate sensitivity. The value(s) derived in this way usually have a large uncertainty range as the real climate is ever changing. This transient response of the climate system depends not only on radiative forcing but also on the heat uptake of the world's oceans. Heat uptake varies regionally and is difficult to measure: a dense network of instruments is needed to derive global average figures for heat uptake and how it changes over time. Uncertainties in this, as well as in temperature and the net radiative forcing, hamper accurate estimates of climate sensitivity. In addition, feedbacks operate on a wide range of timescales. This implies that the climate sensitivity derived from observations over a limited time span will not include the slow feedbacks. A clear advantage of using observations is that they represent the real world.

Both approaches are discussed in this chapter. Section 5.2 deals with the theoretical and modelling aspects of climate sensitivity and feedbacks. The focus of section 5.3 is on deriving the value of climate sensitivity from observational constraints, using data from various time spans: satellite data from 1979 until now, 20th century observations and paleo data on the chemical composition of the atmosphere and temperature. For both approaches the conclusions from recent literature are merged to provide an estimate of the likely range of climate sensitivity.

5.2 Theoretical and modelling aspects of climate sensitivity and feedbacks

5.2.1 Probability distribution of climate sensitivity

A persistent feature of empirical climate sensitivity estimates is the sizeable probability of high sensitivities: the 'heavy upper tail'. General claims are made that this is either an unavoidable feature of the Earth system or a manifestation of limitations in our observational capabilities. The

observational constraints considered in IPCC AR4 (Meehl et al. 2007) offer little guidance on the matter of whether the actual climate sensitivity is indeed located in the heavy upper tail.

Tomassini et al. (2007) use a formal statistical analysis to show the asymmetric distribution of climate sensitivity estimates. Their statistical analysis identifies ocean heat content, a quantity that is related to the conservation of energy, as an important constraint. However, an estimate of this potential constraint is hampered by the fact that natural variability, such as El Niño, influences the world ocean heat content as well. A comprehensive analysis of the observational error and the reliability of the dataset is also lacking (Gregory et al. 2004).

Roe and Baker (2007) claim that the heavy tail of climate sensitivity is “an inevitable and general consequence of the nature of the climate system in which the net feedbacks are substantially positive”. The logic behind this reasoning is that a combination of many uncertain feedback factors approaches a normally distributed total feedback factor. The non-linear relationship between feedback and climate sensitivity¹⁷, means that this normal distribution for the feedback results in a skewed probability distribution with a heavy upper tail of climate sensitivity estimates. The sum of uncertainties from all the component feedback processes can be interpreted in three ways: uncertainty in understanding physical processes, uncertainty in the observations used to evaluate the feedback factor and, lastly, inherent variability in the strength of the major feedbacks.

Urban and Keller (2009) claim that the heavy upper tail of the probability density function is due to limitations in our observational capabilities. They expect that the inclusion of more observations will reduce uncertainty, in particular both surface temperature and ocean heat uptake, because these quantities are complementary to one another when estimating the uncertainty in climate sensitivity. In other words: a tail of the climate sensitivity estimate which is allowed by one observational constraint may be excluded by the other.

However, the analysis of Urban and Keller neglects several likely important processes and sources of uncertainty, e.g. uncertainties about historic climate forcings (Forest et al. 2002). Since direct and indirect aerosol forcings are significant and their combined uncertainty is also comparable in size to the forcing itself (Alley et al. 2007), treatment of this uncertainty will widen the tails of the climate sensitivity distribution (Allen et al. 2006).

Conversely, Murphy et al. (2009) showed that it is possible to retrieve an upper bound for the (negative) aerosol forcing of -1.9 Wm^{-2} from observations (see section 5.3.1). They used measurements of surface temperature, ocean heat content and satellite observations of radiative fluxes. Published results from radiative transfer calculations have been used to obtain the history of radiative forcing due to greenhouse gases and volcanic aerosols as well as the sun, since 1950. Overall, the constraints on climate sensitivity from this observationally-based analysis are similar to those from model-based analyses. However, the recent climate record provides stronger constraints at the low end of climate sensitivity than at the high end (Foster et al. 2008, Knutti and Hegerl, 2008).

5.2.2 Twentieth century climate model response and climate sensitivity

The application of simple climate models has elucidated the most important factors that, to first order, determine the Earth’s global mean surface temperature: the climate forcing, the climate sensitivity and the efficiency of ocean heat uptake. Given these factors, the evolution of the global mean surface temperature can be calculated. Although there has been much focus on uncertainties in climate sensitivity, the questions that arise are: how important are uncertainties in forcing and what are the implications of these uncertainties in the ability to simulate the climate of the 20th century (Hansen and Sato, 2001)?

¹⁷ **Non-linear relationship between feedback (f) and climate sensitivity (λ): $\lambda = \lambda_0 / (1 - f)$, where λ_0 is the climate sensitivity without feedbacks operating (referred to as non-feedback sensitivity).**

Kiehl (2007) explored the role of the two factors, climate forcing and climate sensitivity, in climate simulations of the 20th century. It was found that the total anthropogenic forcing for a wide range of climate models differs by a factor of two and that the total forcing is inversely correlated to climate sensitivity. Much of the uncertainty in total anthropogenic forcing derives from a threefold range of uncertainty in the aerosol forcing used in the simulations. Therefore, the range in total anthropogenic forcing is slightly over a factor of 2, which is the same order as the uncertainty in climate sensitivity. These model results explain to a large degree why models with such diverse climate sensitivities can all simulate the global anomaly in surface temperature. The magnitude of applied anthropogenic total forcing compensates for the model sensitivity. It is important to note that in spite of the threefold uncertainty in aerosol forcing, all of the models predict a warming of the climate system over the later part of the 20th century. The warming is, in essence, bounded by the fact that climate sensitivity is a positive quantity and the total forcings used by modellers are also positive.

It could also be argued that these results do not invalidate the application of climate models to projecting future climate for at least two reasons (Kiehl, 2007). First, within the range of uncertainty in aerosol forcing, models have been benchmarked against the 20th century as a way of establishing a reasonable initial state for future predictions. The analogy in weather forecasting would be where models assimilate information to constrain the present state for improved prediction purposes. Climate models are forced within a range of uncertainty and yield a reasonable present state, which improves the models' predictive capabilities. Second, many of the emission scenarios for the next 50 to 100 years indicate a substantial increase in greenhouse gases with an associated large increase in radiative forcing. Given that the lifetime of these gases is orders of magnitude larger than that of aerosols, future anthropogenic forcing will be dominated by greenhouse gases. Thus, the relative uncertainty in aerosol forcing may be less important for projecting future climate change.

Knutti (2008) studied the agreement between simulated and observed 20th century warming using the World Climate Research Programme's (WCRP's) Coupled Model Intercomparison Project phase 3 (CMIP3), a set of simulations with different GCMs used in the IPCC Fourth Assessment Report (IPCC, 2007). This agreement is remarkable because the climate models reproduce the observed surface warming better than might be expected given the uncertainties in radiative forcing, climate sensitivity and ocean heat uptake, suggesting that different models show similar warming for different reasons. Models differ because of their underlying assumptions and parameterizations of physical processes, and it is plausible that choices are made based on the model's ability to simulate observed trends.

Knutti (2008) showed that while climate sensitivity and radiative forcing are indeed correlated across the latest ensemble of models, eliminating this correlation would not greatly change the uncertainty range of long-term temperature projections. He argued that, since most models do not incorporate the aerosol indirect effects, model agreement with observations may be partly spurious. The incorporation of more detailed aerosol effects in future models could lead to inconsistencies between simulated and observed past warming, unless the effects are small or offset by additional forcings. Knutti therefore concluded that the most likely and obvious (although not the only) interpretation from the results is that the total aerosol effect is smaller than suggested by most aerosol models.

5.2.3 Feedbacks

Feedback processes in the climate system determine the strength of the climate sensitivity. Many of these processes, at least the fast feedbacks, are parameterized in comprehensive climate models. Soden and Held (2006) studied the climate feedbacks in coupled ocean–atmosphere models (CMIP2: the model intercomparison in the framework of phase 2 of the World Climate Research Programme) using a coordinated set of 21st century climate change experiments. Water vapour was found to provide the largest positive feedback in all models and its strength is consistent with that would be expected from the fact that warmer air contains more water vapour than colder air with the same relative humidity.

The feedbacks from clouds and surface albedo were also found to be positive in all models, while the only stabilizing (negative) feedback comes from the temperature response through the outgoing longwave radiation. Large intermodel differences in the lapse rate¹⁸ feedback were observed and shown to be associated with differing regional patterns of surface warming. Consistent with previous studies, it was found that the changes in temperature and water vapour in the atmosphere are tightly coupled in all models and, importantly, demonstrate that intermodel differences in the *sum* of lapse rate and water vapour feedbacks are small. In contrast, intermodel differences in cloud feedback were found to provide the largest source of uncertainty in current predictions of climate sensitivity.

Dufresne and Bony (2008) showed that it is possible to decompose, and thus to compare, the contributions of the different climate feedbacks as well as of the ocean heat uptake to the global temperature response to a specified forcing. They applied this methodology to an ensemble of models in phase 3 of the Coupled Model Intercomparison Project (CMIP3) of the World Climate Research Programme (WCRP) in support of the IPCC AR4 Report. Besides confirming the main findings of Soden and Held (2006) they found that in transient simulations the multi-model mean contributions to global warming associated with the combined water vapour–lapse-rate feedback, cloud feedback, and ocean heat uptake are comparable. However, intermodel differences in cloud feedbacks constitute by far the most important source of spread in both equilibrium and transient climate responses simulated by GCMs. The spread associated with intermodel differences in cloud feedbacks appears to be roughly 3 times larger than that associated either with the combined water vapour–lapse-rate feedback, the ocean heat uptake, or the radiative forcing.

Using a complex coupled atmosphere-ocean general circulation model (ECHAM5/MPI-OM), Seiffert and von Storch (2008) studied the dependence of the climate sensitivity on the representation of atmospheric small-scale fluctuations. In one of the experiments they added variability (white noise) to the small-scale patterns in the model in order to mimic the possible impact of unresolved processes. However, it does not qualify as a realistic parameterization of the subgrid scale variability. Rather, it aims to isolate the impact of enhanced small-scale fluctuations on the climate sensitivity in the framework of idealized experiments. They found that the climate sensitivity tends to weaken. In a second experiment Seiffert and von Storch reduced the horizontal diffusion by a factor of three in the climate model. This resulted in an increase of 13% in the climate sensitivity.

It is concluded that the climate sensitivity can be changed due to both nonlinear dynamic and thermodynamic processes. Small-scale variability therefore, can have a direct effect not only on the model sensitivity but also indirect effects by influencing the thermodynamic feedback processes. In general, it is difficult to distinguish between the direct and indirect effects of enhanced small-scale fluctuations on the climate sensitivity. Nevertheless, the noise experiments might have been affected more by indirect effects than the runs with reduced horizontal diffusion.

Spencer and Braswell (2008) used a simple model to demonstrate that any non-feedback source of top-of-atmosphere radiative flux variations can cause temperature variability, which then results in a positive bias in diagnosed feedbacks. This effect is demonstrated by daily random flux variations, as might be caused by fluctuations in low cloud cover. These results suggest that current observational diagnoses of cloud feedback—and possibly other feedbacks—could be significantly biased in the positive direction. Correcting for this positive bias in feedbacks would result in a climate sensitivity of only 1.1°C. However, rapid fluctuations – those which are faster than the feedback processes – affect the radiation budget following the climate sensitivity without feedbacks, referred to as the no-feedback sensitivity. When looking at short time spans (faster and comparable with the timescales of the primary feedback processes), Spencer and Braswell hence found mistakenly positive biases in the feedback parameter, since they calculate a climate sensitivity which is biased towards the no-feedback sensitivity.

¹⁸ The **lapse rate** is the actual change of temperature with altitude in the atmosphere (i.e. the vertical [temperature gradient](#)).

Clement et al. (2009) examined changes in low-level clouds (marine boundary layer clouds) over the Northeast Pacific in observations and climate models. The rationale behind this study is that feedbacks involving low-level clouds remain a primary cause of uncertainty in global climate model projections. At present, even the sign of the low-level cloud feedback in climate change is unknown. Decadal fluctuations were identified in multiple, independent cloud data sets, and changes in cloud cover appeared to be linked to changes in both local temperature structure and large-scale circulation. This observational analysis indicated that increased sea surface temperatures and weaker subtropical high pressure systems act to reduce Northeast Pacific cloud cover. They concluded that clouds act as a positive feedback in this region on decadal timescales. Although these cloud changes were only examined in the Northeast Pacific region, Clement et al. claim that changes in this region are part of a dominant mode of global cloud variability. However, Clement et al. do not give an estimate of the effect on the strength of the feedback or on climate sensitivity.

5.3 Observational constraints on climate sensitivity

5.3.1. Twentieth century

Many estimates of the equilibrium climate sensitivity are based on climate change that has been observed over the instrumental period, about the past 150 years. The advantage of exploring the value of climate sensitivity in this period is the consideration of a climate state similar to today's and the use of similar timescales of observations to the projections of interest, thus providing constraints on the overall feedbacks operating today. Several studies subsequently used the transient evolution of surface temperature, upper air temperature, ocean temperature or radiation in the past, or a combination of these, to constrain climate sensitivity. Since IPCC AR4 several studies have included satellite data for the radiation budget to infer climate sensitivity (Chylek et al. 2007, Murphy et al. 2009).

Constraining climate sensitivity using observations and calculated radiative forcing for the last century is difficult, since the net forcing itself is not well constrained due to the possibility of aerosols substantially cancelling out the greenhouse gas forcing. If the net forcing is small, then climate sensitivity would have to be very high to explain the observed warming. A second difficulty is that climate has not reached equilibrium with respect to the forcings over this relatively short period. Therefore, an estimate is needed for the heat uptake by the oceans, introducing additional uncertainty into the climate sensitivity.

Chylek et al. (2007) analysed satellite and surface measurements of aerosol optical depth in the period between 1985 and 2006. They found that the global average of aerosol optical depth¹⁹ (approximately 0.18) has been decreasing recently at the rate of around 0.0014 per year. This produces a positive radiative forcing at the top of the atmosphere that is comparable in size to the positive forcing due to the rate of increasing atmospheric concentration of carbon dioxide and other greenhouse gases over the same period. Consequently, both increasing atmospheric concentration of greenhouse gases and decreasing loading of atmospheric aerosols are major contributors to radiative forcing since at least 1985. This results in a climate sensitivity which is reduced by at least a factor of 2 compared to the case where the radiative forcing is ascribed only to increases in atmospheric concentrations of carbon dioxide. Chylek et al. calculated a range between 1.1 and 1.8 K. However, as stated in section 5.2.2, there is a three-fold uncertainty in the estimates of aerosol forcing, mainly through the possible effects on cloud thickness and cloud cover. Chylek et al. (2007) also used an average ocean heat uptake in the range 0.5 to 1.4 Wm⁻²K⁻¹. This range is probably too narrow due to large year-to-year fluctuations in the ocean heat uptake. Furthermore, it should be realized that the analysed period of about two decades is too short to deduce climate sensitivity from temperature changes and the forcings by greenhouse gases and aerosols, since other factors may have influenced the temperature. Therefore, the low

¹⁹ Aerosol **optical depth** is a measure of [transparency](#), and is defined as the negative [logarithm](#) of the fraction of shortwave [radiation](#) (or [sunlight](#)) that is scattered or absorbed by these aerosols.

climate sensitivities deduced by Chylek et al., as well as the small range in this parameter, are highly questionable.

Schwartz (2007) also calculated a low climate sensitivity of 1.1 ± 0.5 C from a regression of ocean heat content versus global mean surface temperature over the period 1880-2004. His calculations imply a very short relaxation time constant of global mean surface temperature in response to perturbations of only 5 years. According to Knutti et al. (2008b) “this means that global surface temperature is nearly in equilibrium with radiative forcing, and that the sum of all feedbacks (water vapor, lapse rate, clouds, albedo) is close to zero. Schwartz’s results are at odds with most of the current literature on climate sensitivity, the idea of commitment warming, i.e. the response time scale of the order of several decades, and with the magnitudes of climate feedbacks quantified in models and observations.”

Schwartz (2008) published a corrected analysis that resulted in an upward revision of the climate system time constant by roughly 70%, to about 9 years, which is still very low, and a climate sensitivity of 1.9 ± 1.0 K, lower than the central estimate of 3°C as given in IPCC AR4, but consistent within the uncertainties of both estimates. Foster et al. (2008) state in their comment on Schwartz’s second paper that the principal physical mechanism which leads us to believe that not all committed greenhouse gas warming has yet been experienced and a substantial amount still remains ‘in the pipeline,’ is the warming of the deep ocean. Furthermore, basic physical considerations argue strongly against the notion that the global average surface temperature has a single characteristic timescale, or time constant, as proposed by Schwartz.

Murphy et al. 2009 examined the Earth’s energy balance since 1950, identifying results that can be obtained without using global climate models. Important terms that could be constrained using only measurements and radiative transfer models are ocean heat content, radiative forcing by long-lived trace gases, and radiative forcing from volcanic eruptions. They explicitly considered the emission of energy by a warming Earth by using correlations between surface temperature and satellite radiant flux data and showed that this term is already quite significant.

Considering ocean heat content, there are several recent calculations from the surface to a 700 m depth (Domingues et al. 2008, Ishii and Kimoto 2009, Levitus et al. 2009, see also section 2.2.3.) that have been updated since the IPCC AR4 report. In addition to heating of the top 700 m, some heat is transported to the deep ocean. For the study of Murphy et al. the heat content to 3000 m depth was estimated to add about 30% to 40% to recent increases above 700 m, on the basis of limited deep ocean temperature data. For a given year, the deep ocean heat content was scaled to the heat content above 700 m averaged over the preceding 10 years. The actual lag may be longer but the results have been found to be insensitive to the averaging period and longer averages require more assumptions about the heat content before 1950 as measurements are lacking from before that date.

Murphy et al. (2009) showed that about 20% of the integrated positive forcing by greenhouse gases and solar radiation since 1950 has been radiated to space. Only about 10% of the positive forcing (about 1/3 of the net forcing) has gone into heating the Earth, almost all of it into the oceans. About 20% of the positive forcing has been balanced by volcanic aerosols, and the remaining 50% is mainly attributable to aerosols in the lower atmosphere. After accounting for the measured terms, the residual forcing between 1970 and 2000 due to direct and indirect forcing by aerosols and any unknown mechanism was estimated by Murphy et al. as $-1.1 \pm 0.4 \text{ Wm}^{-2}$ (likely range: 66%). This is consistent with the best estimates of IPCC AR4, but rules out very large negative forcings from aerosol indirect effects. Further, the data imply an increase from the 1950s to the 1980s followed by constant or slightly declining aerosol forcing into the 1990s, consistent with estimates of trends in global sulfate emissions.

Many global climate models roughly reproduce the ratio whereby radiative response accounts for about twice as much energy as ocean heat uptake (Gregory and Forster, 2008). Overall, the constraints on climate sensitivity from this observationally-based analysis are similar to those from model-based

analyses: the recent climate record provides stronger constraints at the low end of the climate sensitivity than at the high end (Foster et al. 2008, Knutti and Hegerl, 2008).

5.3.2. *Climate sensitivity from paleo data*

Paleo records based on proxies of temperatures in combination with records or estimates of radiative forcing are used to determine the climate sensitivity. Most problematic for constraining its value is the accuracy of the data concerning radiative forcing (particularly solar and volcanic forcing), especially in periods in which the temperature was not fluctuating much, e.g. in the last millennia. Apart from uncertainties in radiative forcing and the spread in temperature reconstructions, the climate system shows (unforced) climate variability. This hampers the determination of the climate sensitivity. An additional problem regarding the derivation of climate sensitivity from paleo data over the last million years is that the driver of glacial interglacial changes lies in the orbital forcing with no significant global footprint, less than 0.5 Wm^{-2} (Köhler et al. 2009). In fact, the concept of climate sensitivity is not well suited for orbital changes.

During the Last Glacial Maximum (LGM) about 18 thousand years ago, global average temperatures were substantially lower than over the modern pre-industrial state, in the range of 3 to 9 °C. However, temperature estimates are imprecise and rely on interpretation of proxy data. The climate of the LGM may be considered as a quasi-equilibrium response to substantially altered boundary conditions, such as large ice sheets over landmasses of the Northern Hemisphere, lower atmospheric CO_2 and CH_4 levels, different vegetation and a possibly-related higher dust content. The main changes in radiative forcing during the LGM are due to lower levels of greenhouse gases and large ice sheets over the northern hemisphere, which both account for about 3 Wm^{-2} (Taylor et al. 2000) but changes in vegetation and dust also each add a little over 1 Wm^{-2} (Crucifix and Hewitt, 2005, Claquin et al. 2003), giving a net radiative forcing estimated at 6–11 Wm^{-2} . Although the climate sensitivity can be calculated from these temperature data and radiative forcing estimates (see Table 5.1), the substantially different climate state and topography of the LGM, combined with the different seasonal and spatial pattern of forcing, might result in a somewhat different sensitivity to radiative forcing in the LGM compared to the present warmer climate.

In a model study Hargreaves et al. (2007) demonstrate an asymmetry in climate sensitivity between calculations of decreasing and increasing levels of greenhouse gases, with 80% of the ensemble having a weaker cooling than warming. This asymmetry, if confirmed by other studies, would mean that direct estimates of climate sensitivity from the LGM are likely to be underestimated in the order of half a degree. However, this result may be model-dependent. Analysis of the parameters varied in the model suggest the asymmetrical response may be linked to the behaviour of ice in the clouds, implying a weakening of the climate sensitivity during glacial periods relative to the present is in the order of 15%.

5.3.2.1. *Low estimate of climate sensitivity*

Difficulties in the quantification of the radiative forcing by changes in aerosol loading of the atmosphere hamper accurate estimates of the climate sensitivity, not only in the 20th century (see section 5.3.1), but also during the LGM. Chylek and Lohmann (2008) calculated a positive radiative forcing due to decreasing aerosol amounts during the Last Glacial Maximum (LGM) to Holocene transition of $3.3 \pm 0.8 \text{ Wm}^{-2}$. They deduced this forcing and that of greenhouse gases from the Vostok ice core (i.e. air-bubbles in the 3 km ice-layer of Antarctica), including the radiative forcing due to changes in the Earth's albedo due to ice sheet coverage and vegetation changes. To arrive at such a high aerosol forcing, which is about a factor 3 above previous estimates, they made use of a cooling period between about 42 thousand years before the present and LGM and calculated a climate sensitivity of between 1.3 and 2.3 K. As in their previous study, this again supports the lower end of the climate sensitivity range of 2 to 4.5 K suggested by IPCC AR4.

However, Ganopolski and Schneider von Deimling (2008) demonstrated that Chylek and Lohmann strongly underestimate related uncertainties and therefore their method used does not provide meaningful constraints on the range of climate sensitivity and LGM aerosol forcing. As indicated above, to quantify the impact of glacial dust content on global LGM forcing, Chylek and Lohmann focused for the first time on an interval in the ice core record starting at 42 thousand years ago. Comparing the aerosol forcing in both periods requires two additional assumptions. First, that the ratio between global and local temperature change remains the same over different periods of time; and second, that the local (Vostok) dust record is representative of the global aerosol forcing. These two assumptions introduce additional uncertainties, which should be reflected in the uncertainty range of the calculated climate sensitivity, which they do not.

In a reply to Ganopolski and Schneider von Deimling, Chylek and Lohmann (2008b) stated that they consider the aerosol forcing of 1 Wm^{-2} frequently used by other investigators to be a considerable underestimate. IPCC AR4 estimates the mean value of the current radiative forcing due to increasing levels of anthropogenic aerosols to be -1.2 Wm^{-2} . It is difficult to accept that the aerosol radiative forcing during the ice ages when dust concentration in the Antarctic and Greenland ice cores is over 50 times higher than during the Holocene, and when increased dust deposits are observed in marine sediments in Atlantic and Pacific Ocean, could be lower than the current anthropogenic aerosol radiative forcing. A radiative forcing during the LGM of between 1.9 and 3.3 Wm^{-2} has been suggested by Harvey (1988). Other researchers (Harrison et al. 2001; Claquin et al. 2003) pointed out that the aerosol radiative forcing during the LGM, at least in the tropics, could be as strong as the positive radiative forcing due to increased greenhouse gases during the LGM to Holocene transition. The Chylek and Lohmann observational results (and supported by ECHAM5 GCM simulation) confirm these higher values and suggest that the radiative forcing due to aerosols was, indeed, approximately equal to the radiative forcing of greenhouse gases during the LGM to Holocene transition, i.e. both in the order of 3 Wm^{-2} . This would lower the value of climate sensitivity considerably, since the observed temperature during this transition would be caused by a higher radiative forcing.

Köhler et al. (2009) compiled various environmental records (and model-based interpretations of some of them) in order to calculate the direct effect of various processes on the Earth's radiative budget and, thus, on the global annual mean surface temperature over the last 800,000 years. They computed a net forcing of $-12.4 \pm 1.4 \text{ Wm}^{-2}$ during the LGM relative to the present. Excluding the effects due to changes in the slow feedbacks (which are not incorporated in current climate models), i.e. ice sheets and the distribution of vegetation, their estimate of the net forcing is $9.5 \pm 1.2 \text{ Wm}^{-2}$ (66% range). Combined with the estimated range of temperature difference between LGM and the present of $5.8 \pm 1.4 \text{ K}$, this implies a present-day climate sensitivity of between 1.4 and 5.2 K with a most likely value near 2.4 K . This is somewhat smaller than other methods but consistent with the consensus range of 2 – 4.5 K derived from other lines of evidence (Knutti and Hegerl, 2008). Köhler et al. (2009) accounted for a decrease in the fast feedbacks (lapse rate, water vapour, sea ice and snow albedo, and the cloud feedback) of 15% as suggested by Hargreaves et al. (2007) for the LGM sensitivity relative to the present.

5.3.2.2. *High estimate of climate sensitivity*

In order to determine the response to external forcings it is important to distinguish between forcings and feedbacks. This distinction is dependent on the timescale considered. For instance, to determine the transient response in the last few decades to solar, volcanic and anthropogenic forcings only, the fast feedbacks need to be taken into account, such as water vapour and cloud feedbacks (see section 5.2.3). Over timescales of thousands of years the slow feedbacks should be incorporated as well. The effects of these feedbacks could be calculated in terms of radiative forcing – as was actually done in the previous sections – but the notion that in fact an autonomous process or feedback is acting has implications for the response to external forcings.

According to Hansen et al. (2007), paleoclimate data show that the Earth's climate is remarkably sensitive to global forcings. Positive feedbacks predominate. This allows the entire planet to be whipsawed between climate states. One feedback, the 'albedo flip' property of ice/water, provides a powerful trigger mechanism. A climate forcing that 'flips' the albedo of a sufficient portion of an ice sheet could enhance the climate response. The ice sheet and ocean inertia only moderately delays ice sheet disintegration and a burst of added global warming. Vegetation migration and the release of greenhouse gases from soils, tundra or ocean sediments may also be considered as a slow feedback. These slow feedbacks are generally not included in climate models. Hansen et al. (2007, 2008) concludes that our best estimate of climate sensitivity is $\sim 3^{\circ}\text{C}$ for doubled CO_2 , including only fast feedback processes. Equilibrium sensitivity, including slower feedbacks, is $\sim 6^{\circ}\text{C}$ for doubled CO_2 for the range of climate states between glacial conditions and ice free Antarctica.

Hansen et al. (2007) include climate-driven aerosol changes and their cloud effects as a 'fast feedback' because aerosols respond rapidly to climate change. However, part of the aerosol feedback may be classified as slow due to vegetation changes. Ice core data show that aerosols decrease as the climate warms, probably because increased water vapour and rainfall wash out aerosols. Hansen et al. (2007) did not quantify the aerosol forcing, but made a statement about the small net aerosol forcing instead. Therefore, Hansen et al. mainly adopted two 'forcings', namely ice-albedo and greenhouse gases. However, including a large aerosol forcing in the order of 3 Wm^{-2} as a difference between glacial and interglacial conditions as suggested by Chylek and Lohmann (2008) (see section 5.3.2.1) would lower this high climate sensitivity estimate made by Hansen. To what extent actually depends on the timescale of the aerosol feedback, which is still under debate.

5.3.3. Carbon dioxide and temperature in the geological past

If fossil fuel emissions continue unabated, atmospheric CO_2 levels are expected to rise to ~ 1800 ppmv over the next few centuries (Archer 2005, Uchikawa and Zeebe, 2008). Similar CO_2 concentrations prevailed during the Late Paleocene and Early Eocene epochs, between 60 and 50 million years ago with a minimum of 1125 ppmv (Lowenstein and Demicco, 2006) but potentially much higher. Geochemical indicators show that global average surface temperatures were at least 15°C higher than at present (Pearson et al. 2007; Sexton et al. 2006; Sluijs et al. 2008a; Hollis et al. 2009; Greenwood and Wing, 1995; Wing et al. 2005), while deep sea temperatures averaged 10°C (Zachos et al. 2008). The present deep sea temperatures are approximately 2°C . Superimposed on a long-term Late Paleocene - Early Eocene warming trend, strong global warming events occurred during the Paleocene-Eocene thermal maximum (PETM; ~ 55.5 Ma) and the recently discovered Eocene thermal maximum 2 (ETM2; 53.5 Ma). Global warming comprised $5\text{-}8^{\circ}\text{C}$ during the PETM (Sluijs et al. 2006; Thomas et al. 2002; Zachos et al. 2003) and $\sim 3\text{-}4^{\circ}\text{C}$ during ETM2 (Lourens et al. 2005; Stap et al. 2009).

Carbon isotope ratios measured on terrestrial and marine sediments show that large masses of carbon with low ^{13}C concentrations were released into the ocean atmosphere system during these high temperature events (Dickens et al. 1997; Schouten et al. 2007; Sluijs et al. 2007a). This indicates that possible sources include methane from the decomposition of submarine clathrates (Dickens et al. 1995) or of thermogenic origin (Svensen et al. 2004), CO_2 from the oxidation of organic-rich sediments (Kurtz et al. 2003) or a combination of sources released sequentially as positive feedbacks (Sluijs et al. 2007a). This carbon was injected within 1 – 10 kyr and analogous in both mass ($1\text{-}7 \times 10^{18}$ g) (Dickens et al. 1997; Panchuk et al. 2008) and isotopic composition (Böhm et al. 2002; Dickens 2001) to current and projected anthropogenic emissions. Sequestration of carbon occurred on timescales of $10^4\text{-}10^5$ yr through a combination of chemical weathering of rocks and subsequent carbonate burial (Zachos et al. 2005) and the burial of organic carbon (Sluijs et al. 2008a; John et al. 2008). The PETM drove global ecosystem alteration on land and in the ocean (Sluijs et al. 2007b), including algal blooms (Crouch et al. 2001) and severe extinction among deep-marine calcifying protists (Thomas 2007). Despite low abundances of continental ice during the latest Paleocene, the

PETM was associated with global sea level rise, caused at least in part by the thermal expansion of ocean water (Sluijs et al. 2008b).

Tropical and subtropical temperatures between 60 and 50 Ma ago were at least 5 °C higher than at present (Pearson et al. 2007; Sexton et al. 2006; Pearson et al. 2001), consistent with higher CO₂ concentrations and associated feedbacks. High latitudes, particularly, were warm with temperatures of up to 40 °C warmer than at present (Sluijs et al. 2008a; Hollis et al. 2009; Sluijs et al. 2006; Uhl et al. 2007; Ivany et al. 2008; Weijers et al. 2007). In part, this amplified high-latitude warmth can be understood due to the absence of ice-albedo. But current generation IPCC-class fully-coupled global climate models with 1120 ppmv CO₂ and Eocene geography simulate mean annual Arctic temperatures near freezing in the absence of a significant albedo feedback (Shellito et al. 2003). Much higher CO₂ concentrations of ~8000 ppmv are required to match the Arctic proxy-data (Sluijs et al. 2006). Even though constraints on tropical temperatures are under debate and may have been warmer than previously assumed (Huber, 2008), the temperature difference between poles and equator was smaller than can be simulated with the models. Apparently, feedbacks to higher temperatures and/or higher greenhouse gas concentrations currently not incorporated in the models either suppressed tropical temperatures and/or boosted high latitude warmth. Such mechanisms may include tropical cyclone-induced ocean mixing (Korty et al. 2008; Sriviver and Huber, 2007) or high latitude convection and cloud feedbacks (Abbot and Tziperman, 2008a; 2008b; Peters and Sloan, 2000). It is uncertain at which greenhouse gas or temperature threshold such mechanisms may be established.

While both the PETM and ETM2 are clear examples of rapid carbon release and associated global climate warming, uncertainties regarding absolute atmospheric CO₂ concentrations, the mass of added carbon and rate of injection, currently prohibit estimation of climate sensitivity to CO₂. New insights, however, can be expected through modelling ocean chemistry, which should elucidate ocean carbonate concentrations and speciation, and thus atmospheric CO₂ across these extremely warm periods (Zachos et al. 2008; Panchuk et al. 2008; Zeebe and Zachos, 2007) combined with detailed global paleo-temperature estimates.

5.3.4. Merging the constraints: a total picture?

In IPCC (2007) it was stated that there is no well-established formal method of estimating a single probability density function from the individual results, taking into account the different assumptions in each study. Most studies do not account for structural uncertainty and thus probably tend to underestimate the uncertainty. However, as several largely independent lines of evidence indicate similar most likely values and ranges, climate sensitivity values are likely to be better constrained than those found by methods based on single data sets (Annan and Hargreaves, 2006; Hegerl et al. 2006). IPCC (2007) concluded on the basis of observations and the strength of known feedbacks simulated in climate models that the (equilibrium) climate sensitivity, is likely (66% probability) to lie in the range 2°C to 4.5°C, with a most likely value of about 3°C.

Annan and Hargreaves (2006) demonstrated how multiple independent observationally-based estimates of climate sensitivity can be used to generate a substantially tighter bound than any previously presented (see Table 5.1). Their analysis results in a very likely range (95% as defined in their study) of 1.7 to 4.9 °C. With their estimate Annan and Hargreaves (2006) cannot assign a significant probability to climate sensitivity exceeding 6 °C without making what appear to be wholly unreasonable assumptions to discard data and/or hugely inflate the uncertainties attached to a range of observational evidence. Even with generous uncertainty estimates, a value greater than 4.5 °C seems very unlikely (less than 5% chance). However, this range is smaller than can be deduced from IPCC (2007), showing a substantial difference with the probability of high sensitivities. This can be traced back to their assumption of independent probability density functions from each of the lines of evidence, implying a multiplication of these functions. The more (independent) lines there are, the narrower the resulting probability density function is. Therefore, combining multiple lines of evidence should be treated with caution.

Observational constraints	Best estimate (C)	Very likely range (95%)
20 th century	3	1 – 10
Volcanic eruptions	3	1.5 – 6
LGM cooling	2.7	0.6 – 6.1
Combined constraints	2.9	1.7 – 4.9

Table 5.1 Likelihood estimates for climate sensitivity based on various observational constraints. Best estimates (peak in probability density function) and 95% ranges. In the last row the combination of the three constraints is shown, assuming independence of the constraints (values from Annan and Hargreaves, 2006).

Knutti and Hegerl (2008) made an assessment of the ranges for climate sensitivity from different lines of evidence. The similarity of the ranges from these various lines substantially increases confidence in an overall estimate. However, the difficulty in formally combining lines of evidence lies in the fact that every single line of evidence needs to be entirely independent of the others, unless dependence is explicitly taken into account. Additionally, if several climate properties are estimated simultaneously that are not independent, such as climate sensitivity and ocean heat uptake, then combining evidence requires combining joint probabilities rather than multiplying the probability density functions afterwards. The combination of the various lines of evidence independently results in a likely range (66%) of 2 to 4.5°C.

Due to the essentially skewed distribution, the probability that climate sensitivity is underestimated, compared to the best estimate of 3°C, is greater than that it is overestimated. In general, an upper bound of climate sensitivity is difficult to assess, but studies rarely assign a high probability to values in excess of 10°C, and they generally point to a maximum likelihood value of well within the consensus range of IPCC AR4. However, high values exceeding 6°C cannot be ruled out by the comparison of models with observations (Knutti and Hegerl, 2008).

The recent warming does not provide a useful constraint for the upper limit of climate sensitivity for the fundamental reason that the net forcing itself is not well constrained. This is caused by the uncertainty in aerosol forcing, which may substantially cancel out the greenhouse gas forcing (Kiehl 2007; Knutti 2008; Knutti and Hegerl, 2008). If the net forcing is small, then climate sensitivity would have to be very high to explain the observed warming (see section 5.2.2.). However, Murphy et al. (2009) succeeded in bounding the negative aerosol forcing to -1.9 Wm^{-2} , using observations of radiative fluxes and ocean heat uptake. This would still imply an upper bound of climate sensitivity of 10°C.

Another factor which is problematic in constraining the value of climate sensitivity is the accuracy of the data in the last millennium concerning radiative forcing (particularly solar and volcanic forcing), while the temperature was not fluctuating much. Apart from uncertainties in radiative forcing and the spread in temperature reconstructions, the climate system shows (unforced) climate variability. This hampers the determination of the climate sensitivity. An additional problem regarding the derivation of climate sensitivity from paleo data over the last million years is that the driver of glacial interglacial changes lies in the orbital forcing with no significant global footprint, less than 0.5 Wm^{-2} (Köhler et al. 2009). This is important in the question to what extent slow feedbacks will play a role in future climate change. In fact, the concept of climate sensitivity is not well suited for orbital changes. Moreover, the strength of the fast feedbacks such as water vapour, lapse rate, clouds and snow and sea ice cover may be state and time-dependent, hence altering the climate sensitivity during ice ages.

Finally, recent studies (Chylek and Lohmann 2008; Hansen et al. 2008; Knutti and Hegerl 2008; Köhler et al. 2009) show a large spread in estimates of radiative forcing from glacial to interglacials with respect to changes in surface albedo and dust aerosols.

At the low end of climate sensitivity, values below 1.5°C are physically extremely unlikely. Knutti and Hegerl (2008) claim that studies that suggest lower climate sensitivity values use erroneous forcing assumptions (for example, hypothesized external processes such as cosmic rays driving climate, in Shaviv, 2005), neglect internal climate variability (Chylek et al. 2007), use overly simplified assumptions, neglect uncertainties, or have other errors in the analysis or dataset, or a combination of these (Lindzen and Giannitsis 2002; Schwartz 2007; Douglass and Knox 2005; Chylek and Lohmann 2008). These results are typically inconsistent with comprehensive models. In some cases they could be refuted by testing the estimation method with a climate model with known sensitivity (Wigley et al. 2005a; 2005b; Knutti et al. 2008; Foster et al. 2008).

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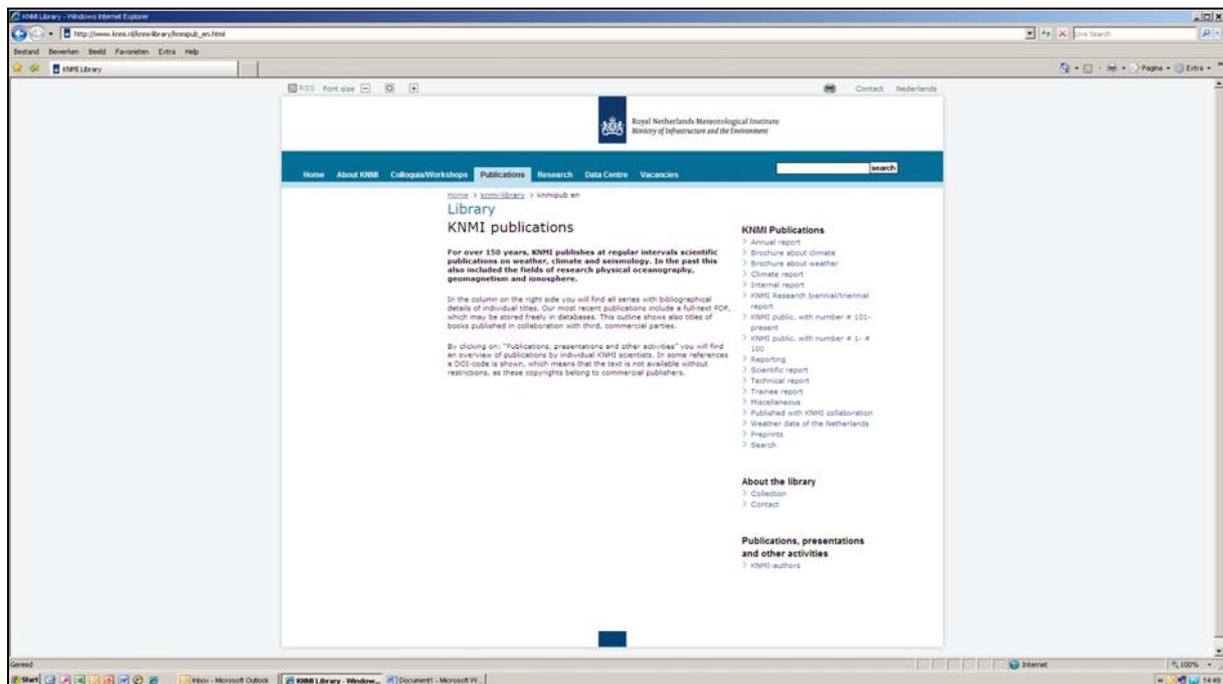
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