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MODELING KEY PROCESSES CAUSING CLIMATE CHANGE AND VARIABILITY

Svante Henriksson

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Title
Modeling key processes causing climate change and variability

Abstract

Greenhouse gas warming, internal climate variability and aerosol climate effects are studied and the importance to understand these key processes and being able to separate their influence on the climate is discussed. Aerosol-climate model ECHAM5-HAM and the COSMOS millennium model consisting of atmospheric, ocean and carbon cycle and land-use models are applied and results compared to measurements. Topics at focus are climate sensitivity, quasiperiodic variability with a period of 50-80 years and variability at other timescales, climate effects due to aerosols over India and climate effects of northern hemisphere mid- and high-latitude volcanic eruptions.

The main findings of this work are 1) pointing out the remaining challenges in reducing climate sensitivity uncertainty from observational evidence, 2) estimates for the amplitude of a 50-80 year quasiperiodic oscillation in global mean temperature ranging from 0.03 K to 0.17 K and for its phase progression as well as the synchronising effect of external forcing, 3) identifying a power law shape $S(f) \propto f^{-\alpha}$ for the spectrum of global mean temperature with $\alpha \sim 0.8$ between multidecadal and El Nino timescales with a smaller exponent in modelled climate without external forcing, 4) separating aerosol properties and climate effects in India by season and location 5) the more efficient dispersion of secondary sulfate aerosols than primary carbonaceous aerosols in the simulations, 6) an increase in monsoon rainfall in northern India due to aerosol light absorption and a probably larger decrease due to aerosol dimming effects and 7) an estimate of mean maximum cooling of 0.19 K due to larger northern hemisphere mid- and high-latitude volcanic eruptions.

The results could be applied or useful in better isolating the human-caused climate change signal, in studying the processes further and in more detail, in decadal climate prediction, in model evaluation and in emission policy design in India and other Asian countries.

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Nimeke

Tärkeiden ilmastonmuutosta ja -vaihteluita aiheuttavien prosessien mallintaminen

Tiivistelmä

Väitöstyössä tutkitaan kasvihuonekaasujen aiheuttamaa ilmaston lämpenemistä, ilmaston sisäistä vaihtelua ja aerosolien ilmastovaikutuksia ja keskustellaan näiden prosessien ymmärtämisen ja voimakkuuden määrittämisen tärkeydestä. Työkaluina sovelletaan aerosoli-ilmastomallia ECHAM5-HAM ja ilmakehä-, valtameri- ja maankäyttö-hiilenkiertomallista koostuvaa COSMOS-millenniummallia, joiden tuloksia verrataan mittauksiin. Aiheista erityisen mielenkiinnon kohteena ovat ilmaston herkkyys, kvasijaksollinen vaihtelu jaksona 50-80 vuotta sekä vaihtelu muilla aikaväleillä, Intian aerosolien ominaisuudet ja ilmastovaikutukset ja pohjoisen pallonpuoliskon keskileveys- ja korkeiden leveysasteiden tulivuorenpurkausten ilmastovaikutukset.

Työn tärkeimmät löydökset ovat 1) jäljellä olevat haasteet havainnoista arvioitavan ilmaston herkkyyden epävarmuuden pienentämisessä 2) arviot kvasijaksollisen maapallon keskilämpötilan 50-80 vuoden heilahtelun amplitudille 0,03 K:n 0,17:n välillä ja heilahtelun vaiheen etenemiselle sekä ulkoisen pakotteen synkronoivalle vaikutukselle, 3) potenssilakimuotoisen spektrin $S(f) \propto f^{-\alpha}$ tunnistaminen maapallon keskilämpötilassa monivuosikymmentasoisten ja El Nino -taajuuksien välillä, jossa $\alpha \sim 0.8$, ja pienempi jos simulaatiossa ei käytetä ulkoisia pakotteita 4) Intian aerosolien ominaisuuksien ja ilmastovaikutusten erottelu vuodenajan ja paikan suhteen, 5) ilmakehässä syntyneiden sulfaattihiukkasten leviäminen primäärisiä hiiliaerosoleja tehokkaammin simulaatioissa, 6) monsuunisateiden lisääntyminen pohjois-Intiassa aerosolien valon absorptiosta johtuen sekä luultavasti sitä suurempi sateiden väheneminen aerosolien valoa himmentävästä vaikutuksesta johtuen ja 7) arvio 0,19 K keskimääräiselle maksimiviilenemiselle suurempien pohjoisen pallonpuoliskon tulivuorenpurkausten jälkeen.

Tulokset saattavat olla hyödyllisiä ihmisen aiheuttaman ilmastonmuutossignaalin eristämisessä, prosessien tarkemmassa ja yksityiskohtaisemmassa tutkimuksessa, vuosikymmentason ilmastoennusteissa, mallien arvioinnissa ja hiukkaspäästöihin liittyvän politiikan suunnittelussa Intiassa ja muissa maissa.

Julkaisijayksikkö
Ilmastonmuutos

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Asiasanat
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Titel

Modellering av viktiga processer som orsakar klimatförändring och variabilitet av klimatet

Sammandrag

Uppvärmning av klimatet p.g.a. växthusgaser, intern variabilitet av klimatet och klimateffekter av aerosoler studeras och vikten av att förstå och kunna kvantifiera dessa processer diskuteras. Aerosolklimatmodellen ECHAM5-HAM och COSMOS-milleniummodellen, som består av atmosfär-, ocean- och landsanvändnings- och kolkretsloppsmodeller, tillämpas och resultat jämförs med observationer. Ämnen som avhandlingen koncentrerar sig på är klimatkänsligheten, kvasiperiodisk variabilitet med en period på 50-80 år och variabilitet med andra tidsskalor, klimateffekter och egenskaper av aerosoler i Indien och klimateffekter av vulkanutbrott vid höga och medelhöga breddgrader av det norra halvklotet.

De viktigaste upptäckterna i denna avhandling är 1) påvisandet av de återstående utmaningarna i att minska osäkerheten av klimatkänsligheten uppskattad från observationer, 2) uppskattningar av amplituden av en kvasiperiodisk 50-80 årig oskillation i den globala medeltemperaturen från 0,03 K till 0,17 K samt för dess fasutveckling och den synkroniserande effekten av yttre drivningar, 3) igenkännande av ett potenslagsformat spektrum $S(f) \propto f^{-\alpha}$ för den globala medeltemperaturen med $\alpha \sim 0.8$ mellan tidsskalor av flera decennier och El Nino -tidsskalor samt en mindre exponent i simulationer utan yttre drivning, 4) åtskiljning av aerosols egenskaper och klimateffekter i Indien på basen av plats och årstid, 5) en mera effektiv spridning av primära sulfataerosoler än sekundära kolaerosoler i simulationerna, 6) en ökning i monsunsnederbörden i Indien p.g.a. absorption av ljus av aerosoler och en sannolikt större minskning p.g.a. mattning av ljus av aerosoler och 7) en uppskattning på 0,19 K för den maximala kylningen p.g.a. större vulkanutbrott på höga och medelhöga breddgrader av det norra halvklotet.

Resultaten kan vara nyttiga i isolering av klimatförändringssignalen orsakad av människor, i noggrannare och mer detaljerade studier av processerna, i klimatprognoser med årtiondeskala, i värdering av modeller och som stöd för politisk beslutsfattning i samband med partikelutsläpp i Indien och andra länder.

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Contents

List of publications

Henriksson, S. V., Arjas, E., Laine, M., Tamminen, J., and Laaksonen, A.: Comment on "Using multiple observationally-based constraints to estimate climate sensitivity" by J. D. Annan and J. C. Hargreaves, Geophys. Res. Lett., 2006, Clim. Past, 6, 411-414, doi:10.5194/cp-6-411-2010, 2010.

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Henriksson, S. V., Pietikäinen, J.-P, Hyvärinen, A.-P., Räisänen, P., Kupiainen, K., Tonttila, J., Hooda, R., Lihavainen, H., Backman, L., Klimont, Z., and Laaksonen, A., Spatial distributions and seasonal cycles of aerosol climate effects in India seen in global climate-aerosol model (submitted to Atmospheric Chemistry and Physics Discussions)

1 Introduction

Understanding Earth's climate is built upon observing it and modeling its physical processes. Carbon dioxide and other greenhouse gases warm the climate through their capacity to absorb and re-emit longwave radiation. Aerosols affect the climate mainly through scattering and absorbing sunlight and modifying cloud properties, which presently appears to result in a net cooling effect [Forster et al. (2007)]. Meanwhile, the atmosphere and ocean have their own internal and coupled dynamics. These are key processes causing climate change and variability. When observing the complex climate system, it is often challenging to quantify the role the different processes have had to produce the observed change and variability. On the other hand, modeling the processes to have the right strengths and characteristics is also challenging.

Carbon dioxide emitted into the atmosphere has a lifetime of decades, centuries or much longer [Archer et al. (2009)]. Greenhouse gas warming of the climate thus happens on the timescales from decades to centuries, even millennia. Aerosols in the troposphere, on the other hand, have short lifetimes, from days to weeks. Thus, their immediate effect vanishes as soon as emissions stop. Internal variability is known to happen at a wide range of timescales [Huybers and Curry (2006)].

The 20th century record of instrumental observations of the climate has relatively good coverage globally and provides the best observational data of the climate overall. Therefore, understanding these observations is central to understanding the climate and its change and variability. The goal of this thesis is to contribute to the understanding of the three factors: greenhouse gas warming, aerosols and internal variability. These are all known to have had significant contributions to the development of observed global mean temperature between 1850 and present, shown in Figure 1. Although greenhouse gas warming is estimated to have been the most important contributor to the increase in global mean temperature and is expected to dominate in the future if greenhouse gas emissions develop according to projections assuming a fossil-fuel based economy, its contribution can not be exactly quantified and is dependent among other things on the contributions that aerosols and internal variability (and other processes such as varying solar radiation) have had in producing the observed temperature changes.

An important part of the scientific literature on climate change is formed by the assessment reports (ARs) by the Intergovernmental Panel on Climate Change (IPCC). The ARs have been published in 1990, 1995, 2001 and 2007 and AR5 is expected to be completed in 2013/14 [IPCC (2013)]. In the AR4, the uncertain magnitude of the *radiative forcing* [Forster et al. (2007)] caused by aerosols was mentioned to be the most important single factor limiting understanding of past and future climate changes. The uncertainty range of the estimate of radiative forcing due to carbon dioxide in 2005 was 1.49 to 1.83 W/m^2, while the whole radiative forcing uncertainty was 0.6 to 2.4 W/m^2, much larger mostly due to aerosols. Since the AR4 and also before it, there has been a debate on the reasons behind the lack of rise in global mean temperatures in the 1950s and 1960s and between 1998 and present despite greenhouse gas warming; whether aerosol cooling or internal variability has been more important [Booth et al. (2012), Zhang et al. (2013)]. As will be discussed later in this thesis, a relatively regular oscillation with a period of 60-70 years seems to appear in the instrumental temperature record, and internal variability at frequencies corresponding to periods of 50-80 years seem to be strong, compared to that at other timescales, also based on model simulations.

The layout of this thesis is as follows. Firstly, the uncertainty of sensitivity of the climate to greenhouse gas warming and the important contributions of internal variability and aerosol climate effects to the uncertainty are discussed. Then, the contributions of internal variability and aerosol climate effects to climate change and variability are discussed in more detail. Internal, and externally-forced, climate variability at a wide range of timescales is discussed, especially *quasiperiodic* internal variability with a period of 50-80 years. Aerosol effects on the climate are discussed in general, and in particular through the examples of volcanic eruptions and aerosols over Asia.

The main tools applied are climate models developed at the Max Planck Institute for Meteorology. Observations are also analysed and used for model evaluation. Specific research questions this thesis tries to answer are the following:

* What is the uncertainty of the global mean temperature response to a doubling of atmospheric CO_2 concentration or other well-known forcing?

* How strongly and regularly does the climate vary globally and regionally with a period of 50-80 years due to internal dynamics?

* How much does the climate vary at other timescales due to internal dynamics and external forcing?

* How do aerosols from northern hemisphere mid- and high-latitude volcanic eruptions affect the climate?

* What are the characteristics of aerosols in India and China presently and how do they affect the climate?

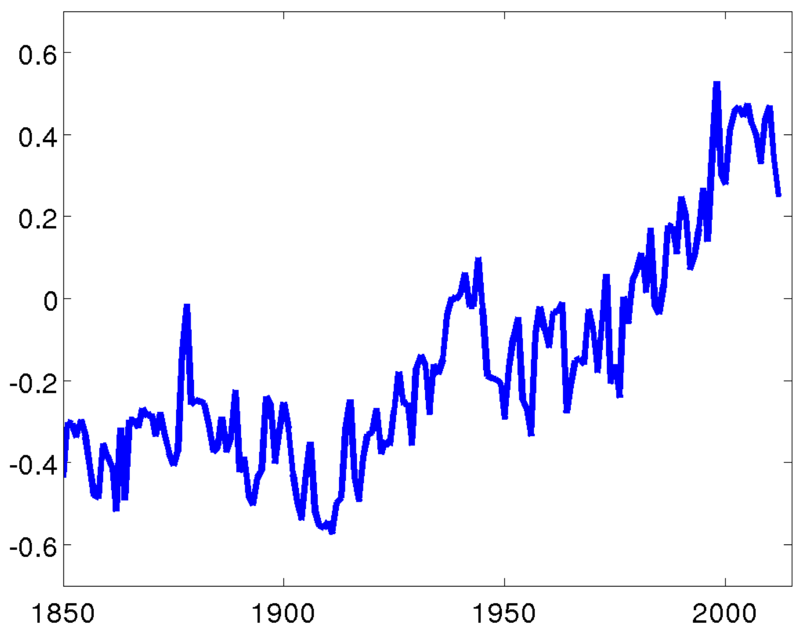

Figure 1: Global annual mean temperature (anomaly in degrees K) between 1850-2012 from the Hadley Center HadCRUT3 dataset.

2 Climate modeling

2.1 General principles

The climate can be mathematically modeled with various levels of detail. As relevant processes range in scale from nanometers (and below), such as new particle formation, to global phenomena, such as the greenhouse effect or El Nino Southern Oscillation, many of which are not fully understood, approximations are necessary and unavoidable. General circulation models or global climate models (GCMs) solving the full equations for atmospheric (and oceanic) flow and energy transfer are employed in the papers included in this thesis, although in Paper **I** also one simple equation describing the global climate is used.

Between simple one-equation models and GCMs resolving the full flow, there are a variety models that can be arranged in hierarchies based on many classifications: the number of spatial dimensions in the model, the extent, to which physical processes are explicitly resolved, the level at which empirical *parametrizations* are involved and the computational cost of running the model [Houghton et al. (1997)]. *Regional climate models* [Jacob (2001), Christensen et al. (2007), Rummukainen (2010)] allow higher resolution modeling in a region of interest than global climate models and may also be viewed as forming part of a climate modeling hierarchy [Randall et al. (2007)]. Earth system models of intermediate complexity (EMICs) [Randall et al. (2007)] lie between simple models and GCMS by describing the same processes as GCMs, only in more parametrized form. Simple models usually have very few degrees of freedom and may have many more adjustable parameters, but EMICs are already assumed to have more degrees of freedom than adjustable parameters by many orders of magnitude [Randall et al. (2007)]. EMICs are often applied in simulations over very long timescales or in experiments scanning wide regions of parameter spaces, made possible by computational ease. Simpler ocean models include mixed-layer models, ignoring processes below the turbulent mixed layer and slab models, which ignore ocean dynamics completely. GCMs allow studying some complex phenomena or interactions between several processes not described by simpler models, this thesis being full of such examples, but complexity is not univocally a positive property either, as a model is only as good as its assumptions. Additionally, simpler models may allow for a better understanding of climate processes [Le Treut et al. (2007), Randall et al. (2007)]. They may help the person interpreting the data to see the forest from the trees.

General circulation models for the atmosphere are based on solving the *primitive equations* for atmospheric flow. The standard variables included in these equations are the three velocity components of the flow, temperature and humidity. Formulating these equations is based on conservation of mass, momentum and energy. Corresponding ocean GCMs solve the equations for the three velocity components of the flow, temperature and salinity. In addition to describing the atmosphere and the ocean, *earth system models* include other components such as aerosol, cryosphere, carbon cycle and ocean biogeochemistry models interacting with the atmospheric and ocean models. *Aerosol-climate models* consider emissions of aerosols, their transport due to atmospheric flow, chemistry and removal from the atmosphere due to rain and dry scavenging.

To solve the equations numerically, they are *discretized* and solved on a grid. Grids may be longitude-latitude meshes with singularities at the poles (which with their surroundings are excluded from the calculations), but also other grids are used. Even when resolving the primitive equations on a grid, by necessity a large number of essential quantities have scales smaller than the resolution of the grid. Such features include boundary layer turbulence, cloud processes and detailed topography. Although some of these processes could be described in more detail using high resolution, it is usually computationally too expensive to include them in climate models. The standard solution is *parametrization*, expressing the sub-grid scale processes with equations including the values at the grid points. Parametrizations of sub-grid processes can strongly influence the nature of large-scale processes explicitly computed, such as winds and ocean currents [Houghton et al. (1997)]. They are therefore a major issue and a large part of the work in developing climate models.

Parametrizations include uncertain and even non-observable quantities represented as numbers in the parametrized equations. Thus, their exact numerical values are to some degree up to the subjective decision of the model developers. Examples of quantities that are usually parametrized are drag and gravity waves caused by sub-grid scale orography, autoconversion describing conversion of cloud water to rain and parameters related to the entrainment rate describing mixing of environmental air into convective clouds. The parameters are standardly used in *tuning* the models for desired properties, such as the radiative balance at the top of the atmosphere, global mean temperature and large scale wind fields. If the model is out of balance at the top of the atmosphere, its climate will drift away from the state it is in. Tuning is thus correcting the imperfections of the model. Tuning is justified as long as there are more degrees of freedom than adjustable parameters [Mauritsen et al. (2012)], which is believed to be true, although formal studies are few and determining the true or efficient number of degrees of freedom reduced from potentially millions to more limited numbers due to spatial, temporal and inter-variable correlations is a highly non-trivial task [Bretherton et al. (1999), Randall et al. (2007), Yokohata et al. (2011)]. In practice, these degrees of freedom are seen, for example, as modes of variability of the climate system, such as seasonal cycles [Wang and Shen (1999)] or the North Atlantic Oscillation [Randall et al. (2007)]. Most climate phenomena studied in this thesis are properties the climate models are not tuned for and are therefore good additional evaluation of the models. The recent study of Mauritsen et al. (2012) went through tuning of an earth system model in detail and concluded that its effects were less than expected.

The global climate models used in the papers included in this thesis have been developed at the Max Planck Institut for Meteorology (MPI-M). Models belonging to the ECHAM5 family were employed in Papers **II-VI**.

2.2 The ECHAM5 model family

In Papers **V** and **VI**, where we study explicit aerosol distributions and their impacts in Asia, the coupled climate-aerosol model ECHAM5-HAM is employed [Stier et al.(2005)], while the COSMOS earth system model [Jungclaus et al. (2010)] is employed in Papers **II-IV**. In the earth system model, ECHAM5 [Roeckner et al.(2003), Röckner (2006)] and the ocean model MPI-OM [Marsland et al. (2003)] are coupled with the PRISM/OASIS3 coupler [Valcke (2003)] with the carbon cycle and land-use model JSBACH [Raddatz et al. (2007)] included in the atmospheric model and the ocean biogeochemistry model HAMOCC [Wetzel et al. (2006)] included in the ocean model.

ECHAM5 is a fifth generation atmospheric climate model originally developed from the weather model of the European Center for Medium Range Weather Forecasts (ECMWF). It solves prognostic equations for vorticity, divergence, surface pressure and temperature, derived through approximations from the primitive equations and expressed in terms of spherical harmonics with a triangular truncation. The vertical coordinate is a flexible hybrid of terrain-following and pressure levels and in the default version of

ECHAM5 used in the papers of this thesis, the upper level is at 10 hPa. Time integration is done semi-implicitly. The T42L19 and T31L19 spatial resolutions used in this thesis imply horizontal resolutions of about 2.8 degrees and 3.75 degrees, respectively, and 19 vertical levels. It is also possible to run simulations at lower and higher resolutions, from T21 truncation to T159. Non-linear processes, including parametrizations, are treated in an a Gaussian grid with almost regularly spaced grid points. Water vapor, cloud liquid water, cloud ice and other tracers are transported with a flux form semi-Lagrangian scheme on this grid. Radiation is calculated for 4 shortwave and 16 longwave bands.

MPI-OM simulates the ocean and sea ice using the seven primitive equations for the ocean. The model uses a conformal orthogonal grid with poles placed on the continents and locations chosen based on high local resolution near the poles. The Bousinessq approximation is applied for the density of sea water, meaning that variations in density are only considered in the vertical momentum equation. The simulations analysed in this thesis have a nominal horizontal resolution of 3 degrees and 40 vertical levels, with finer and coarser resolutions also being possible.

The aerosol model HAM describes aerosol transport, removal and chemistry by representing the five chemical species included in the model (sulfate, black carbon, organic carbon, mineral dust and sea salt) in seven log-normal modes. The modes are assumed to be externally mixed from each other with each mode being internally mixed. Microphysics is treated in the M7 module [Vignati et al.(2004)]. Emissions of all species except of sulfate are in particular form and emissions of sulfate except for marine DMS emissions are 97.5% in the form of SO_2 and 2.5% in particular form. The model with AEROCOM emissions [Dentener et al. (2006)] and other emissions described in [Stier et al.(2005)] and references therein was in excellent agreement with global average aerosol optical depth (AOD) estimated from AERONET stations and in relatively good agreement with a satellite-measured MODIS-MISR composite. In Section 5.1. and Papers **V** and **VI**, the model's perfomance in present-day India and China is evaluated.

The COSMOS earth system model was used in the millennium simulations to assess and compare the impact of human activities, external natural forcings and internal variability on the climate and carbon cycle since the year 800 [Jungclaus et al. (2010)]. Solar, volcanic, land-use, orbital, greenhouse gas and aerosol forcings are included in the millennium simulations.

We studied several aspects of the climate in the millennium simulations in the papers included in this thesis, including impacts of northern hemisphere volcanic eruptions in Paper **IV** and quasiperiodic variability with a period of 50-80 years in Paper **II**. In paper **III**, we studied the full spectrum of temperature variability at different timescales and how this changed when including only certain forcings. A control simulation without external forcings was used in Papers **II** and **III**. The control simulation was run at FMI, while the forced simulation data were downloaded from the CERA database (see papers **II-IV**).

3 Climate sensitivity and feedbacks

Climate sensitivity is defined as the equilibrium response of global mean temperature to the doubling of atmospheric carbon dioxide concentration. Under certain conditions, it is also a more general simple metric for determining how much equilibrium global mean temperature reacts to a certain amount of heating by greenhouse gases or to *any* external forcing. Firstly, the equilibrium global mean temperature is assumed to react linearly to external forcing:

$$\Delta T = \lambda \Delta Q, \tag{1}$$

where λ is called the climate feedback parameter and has the unit $K/(Wm^2)$. The linearity seems to hold quite well in most GCMs, although it is changing from model to model, and holds more accurately for global than for regional forcings [Ramaswamy et al. (2001)]. Over wide ranges of parameter values the feedbacks are most probably nonlinear [Colman and McAvaney (2009)], in addition to which *tipping points* might change the qualitative climate state [Lenton et al. (2008)], thereby invalidating the linear model. This and possible irreversibility of changes [Solomon et al. (2009)] should be taken into account in any holistic risk assessment.

Secondly, the radiative forcing from the doubling of atmospheric CO_2 concentration is quite well known: 3.7 W/m^2 with an error margin of about 10% [Myhre et al. (1998), Ramaswamy et al. (2001)]. The radiative forcing by CO_2 is well approximated by the logarithmic relationship

$$Q = \alpha \ln(C/C_0), \tag{2}$$

with $\alpha = 5.35$ and C_0 the reference level of carbon dioxide concentration, meaning that the forcing from a doubling of CO_2 is well defined for a wide range of concentrations (calculated for about 280-1000 ppmv in [Myhre et al. (1998)]).

3.1 Blackbody response and feedbacks

An idealised blackbody response to a doubling of CO_2 keeping all other things fixed can be calculated based on the Stefan-Boltzmann relation. Denoting climate sensitivity by ΔT_s, we get [Räisänen (2008)]:

$$\Delta \mathrm{T_s} = \frac{\Delta \mathrm{Q}}{4\epsilon_{\mathrm{eff}}\sigma \mathrm{T_s^3}} = \frac{\Delta \mathrm{Q}}{(4\mathrm{F_{space}}/\sigma \mathrm{T_s^4})\sigma \mathrm{T_s^3}} = \frac{\Delta \mathrm{Q}}{4\mathrm{F_{space}}}\mathrm{T_s} = 1.1\mathrm{K} \tag{3}$$

where $\Delta \mathrm{Q} = 3.7\mathrm{W/m^2}$, $\sigma = 5.670 * 10^{-8}\mathrm{Wm^{-2}K^{-4}}$ is the Stefan-Boltzmann constant, $\epsilon_{eff} < 1$ is the effective emissivity of the Earth, $\mathrm{T_s}$ is the observed global mean temperature and $\mathrm{F_{space}}$ is the observed outgoing longwave radiation at the top of the atmosphere. Alternatively, one could do the calculation without needing the effective emissivity or the observed outgoing longwave radiation by using the radiative temperature of Earth: 255 degrees K. This yields $\Delta \mathrm{T_s} = 0.97$ K, i.e. a sensitivity of about 1 degree, and although the result is similar to the calculation above, it might be less exact as it does not consider the atmosphere as a whole [Manabe and Wetherald (1967)].

For describing the real climate, this simple blackbody model is not sufficient. Besides the *negative feedback* caused by increasing outgoing longwave radiation limiting the temperature response, there are other feedbacks, which seem to sum up to be positive [Randall et al. (2007)], with the evidence described more carefully below when reviewing the literature. Well-known important feedbacks are the water vapor, lapse rate, surface albedo and cloud feedbacks. A warmer atmosphere may contain more water vapor and seems to do so, i.e. there is a positive water vapor feedback. The lapse rate, i.e. the rate of decrease of temperature with height, probably weakens in a warming climate producing a negative feedback. A warming climate generally implies less snow and ice at the surface, a resulting higher albedo and more warming, i.e. a positive feedback. Cloud feedbacks are more complex, but the consensus among state-of-the-art climate models is that both low-level and high-level cloud fields produce a positive feedback, although its magnitude has high uncertainty [Andrews et al. (2012)].

While the paradigm of equilibrium surface temperature and radiative forcing applied in this chapter seems to have strong support from climate model experiments [Randall et al. (2007)], the models have also been critisized for producing climates that are too stable, underestimating centennial variability caused by internal dynamics as well as sensitivity to abrupt change [Valdes (2011), Essex (2013)]. Thus, the first assumption mentioned in the previous section could be invalid if a long-term equilibrium does not exist, if low-frequency internal variability is strong or if a forcing exceeding some threshold qualitatively changes the climate state. Variability in global mean temperature without external forcing is dealt with in the next chapter of this thesis for multidecadal and faster variability.

3.2 Estimating climate sensitivity

There are two fundamentally different approaches to estimating climate sensitivity in the real climate, including the feedbacks. In a 'bottom-up' approach, the climate system including all the feedbacks is modeled and the model will then provide an estimate as the difference of equilibrium temperature with carbon dioxide concentration doubled as compared to the reference concentration and temperature. The uncertainty in the estimates can then be estimated by varying the model parameters according to best understanding of uncertainty. In a 'top-down' approach, a given measured temperature time series and measurements of forcing agents are used to estimate climate sensitivity. This necessarily also involves estimates of the thermal inertia of the climate system in reacting to external forcing, as the forcing in practical cases usually does not stay constant for long enough for the climate to reach equilibrium. The uncertainty of an estimate in this latter case can be estimated by estimating the uncertainties of the forcing data, the thermal inertia of the climate system and the temperature time series. A simple energy balance model may be used in the latter method:

$$\mathrm{c}\frac{\mathrm{d}\Delta \mathrm{T}}{\mathrm{dt}} = \Delta \mathrm{Q} - \frac{1}{\lambda}\Delta \mathrm{T}, \tag{4}$$

where ΔT is the global mean temperature, $\Delta \mathrm{Q}$ is the radiative forcing, c is the ocean heat capacity and λ is the climate feedback parameter. The equation reduces to Equation (1) in the steady state. The feedback parameter can be represented as a sum:

$$\lambda = \lambda_{\mathrm{P}} - \lambda_{\mathrm{WV}} - \lambda_{\mathrm{LR}} - \lambda_{\mathrm{A}} - \lambda_{\mathrm{C}}, \tag{5}$$

with the negative Planck (P) longwave radiation feedback and the water vapor (WV), lapse rate (LR), surface albedo (A) and cloud (C) feedbacks. It is standard in the literature to approximate the feedback parameter as the sum of these known feedbacks as they are, based e.g. on climate model experiments, thought to form most of the total feedback. This equation summing up the feedbacks can be rewritten:

$$\lambda = \lambda_{\mathrm{P}}(1 - \Sigma_{\mathrm{i}}\lambda_{\mathrm{i}}/\lambda_{\mathrm{P}}), \tag{6}$$

where the sum $\Sigma_{\mathrm{i}}\lambda_{\mathrm{i}}$ includes the water vapor, lapse rate, surface albedo and cloud feedbacks. The resulting equilibrium temperature anomaly is:

$$\Delta \mathrm{T} = \frac{1}{\lambda_{\mathrm{P}}} \frac{1}{1 - \Sigma_{\mathrm{i}}\lambda_{\mathrm{i}}/\lambda_{\mathrm{P}}} \Delta \mathrm{Q}. \tag{7}$$

From this form it can be seen that the feedback parameters affect temperature change non-linearly, and uncertainty in the λ_{i}s may cause a large uncertainty of climate sensitivity if the sum $\Sigma_{\mathrm{i}}\lambda_{\mathrm{i}}/\lambda_{\mathrm{P}}$ approaches 1 and the factor $\frac{1}{1-\Sigma_i \lambda_i/\lambda_P}$ (the gain factor) thus becomes large. This possibility of explaining typical long tails in climate sensitivity probability density functions was discussed by Roe and Baker (2007).

Early estimates for climate sensitivity were 5.5 degrees by Svante Arrhenius [Arrhenius (1896)] and, in more recent times, 3 degrees in the well-known Charney report from 1979 [Charney et al. (1979)]. Charney and coauthors reported a most likely value of 3 degrees and an uncertainty interval of 1.5-4.5 degrees. The estimate of the Charney report has stayed perhaps even surprisingly little challenged [Kerr (2004)], despite a lot of development in process description in climate models since 1979. The IPCC AR4 quotes 2-4.5 degrees as a *likely* range of climate sensitivity [Hegerl et al (2006)], meaning a probability exceeding 66%. A real possibility for the climate sensitivity value lying outside that interval remains. Many references in the IPCC AR4 such as [Andronova and Schlesinger (2001), Frame et al. (2005), Forest et al. (2006)] report upper bounds of the 95% confidence interval of the order of 9-10 degrees or higher, while others [Annan and Hargreaves (2006), Hegerl et al (2006), Schneider von Deimling (2006)] report 95% confidence intervals close to the IPCC likely range. The possibility of high values of climate sensitivity based on observations remains from the possibility that aerosol cooling could have masked a large part of the greenhouse gas warming up until now [Andreae et al. (2005)], showing up in Equations (1) and (4) as a small total radiative forcing ΔQ having caused a large temperature anomaly ΔT. Even the studies reaching higher upper bounds for climate sensitivity are critisized by Tanaka et al. (2009) to underestimate the true uncertainty as they only account for uncertainty in historical radiative forcing by scaling an assumed forcing time series with different constants. The studies reaching lower upper bounds include other information than the 20th century observations, which do not exclude high sensitivity due to uncertainty in aerosol radiative forcing. For example information from uncertain paleo-records [Jansen et al. (2007)] or models describing the climate feedbacks can be used if the evidence is evaluated to be strong enough.

Compared to observationally-based studies, global climate models tend to give narrower uncertainty intervals for climate sensitivity [Kerr (2004)]. The range of climate sensitivities in CMIP3 models cited in the IPCC AR4 was 2.1-4.4 K [Randall et al. (2007)], while the range of climate sensivity in the newer generation CMIP5 models is 2.1-4.7 K [Andrews et al. (2012)]. In the CMIP5 models, the differences in cloud feedbacks are an important contributor to the spread. However, there are also examples of higher modeled sensitivities, like in the multi-thousand ensemble [Stainforth et al. (2005)], reaching climate sensitivity values of up to 11 K in model simulations and converting the results to a 95% confidence interval of 2.2-8.6 K with a certain internally consistent representation of model-data discrepancy, though with a simpler ocean model than used in models of full complexity. Models have also been critisized for producing results too similar to each other as compared to uncertainty of the underlying variables [Schwartz et al. (2007), Kiehl (2007)]. Lemoine (2010) made calculations for climate sensitivity based on the possibility that models share uncertainties and biases, also relevant for the discussion related to Paper **I** below. The conclusion was that high climate sensitivity may be more probable than thought based on scatter between different model results. It would be desirable to explore the range of model uncertainty further by scanning tuning parameters in wider, more systematic extent than done up until now, for example with methods like those presented in [Hakkarainen et al. (2012), Järvinen et al. (2010), Solonen et al. (2012)]. Perhaps it will turn out that the models have included information independent of climate observations through the laws of physics describing the dynamics of the system and that the narrower range of uncertainty is justified, but this remains to be confirmed.

3.3 Combining different lines of evidence

In Paper **I**, we comment on and critisize the article by Annan and Hargreaves (2006; hereinafter referred to as AH06), that claimed to have reached a narrow uncertainty for climate sensitivity from observations. AH06 assumes three different lines of evidence to be independent and combine them in a Bayesian estimate of climate sensitivity. The Bayesian framework assumes that there is prior information represented in the form of a prior probability density function, which is then combined with the new data through Bayes formula:

$$f(\mathrm{x}|\mathrm{O},\mathrm{H}) = f(\mathrm{O}|\mathrm{x},\mathrm{H})f(\mathrm{x}|\mathrm{H})/f(\mathrm{O}|\mathrm{H}), \tag{8}$$

where x is the parameter to be estimated, i.e., climate sensitivity, H is the old data, O is the new data and f is a notation for conditional probability density functions. $f(\mathrm{x}|\mathrm{H})$ is called the prior, $f(\mathrm{x}|\mathrm{O},\mathrm{H})$ the posterior and $f(\mathrm{O}|\mathrm{x},\mathrm{H})$ the likelihood function.

In Paper **I**, we point out that of the three constraints used: 20th century warming, volcanic cooling and the last glacial maximum (LGM), the volcanic cooling constraint ignored radiative forcing uncertainty [Wigley et al. (2005)] and likely contains information on the climate system already included in the 20th century warming data. In combining the 20th century warming and volcanic cooling estimates, the likelihood function $f(\mathrm{O}|\mathrm{x},\mathrm{H})$ thus does not reduce to f(O|x) as assumed in AH06. In addition to this, the end result was dominated by the tight LGM constraint derived on a few lines in AH06. We also argued against assuming that the results had come from studies using constant priors.

Annan and Hargreaves reply to the independence concern with three arguments [Annan and Hargreaves (2011)]. Firstly, they note that the same assumption in their view had passed through in earlier articles without attracting particular criticism. Secondly, they claim that neglecting evidence is always expected to result in exaggerated uncertainty but that an assumption of independence may overestimate or underestimate the uncertainty compared to more precise calculations. Thirdly, they claim that the sensitivity calculations of our Comment article strengthen their original result rather than contradict it. As a reply to the first argument, certainly we could have gone through other studies critically as well, but we focused on AH06 in our comment. Additionally, the most recent reference listed by Annan and Hargreaves (2011), [Hegerl et al (2006)] combining observations from the instrumental record with reconstructions had also received a critical response [Schneider (2007), Hegerl et al (2007)] based on exaggerated certainty claims in the reconstruction data, also relevant for our discussion. The second argument is the whole topic of discussion, more a question than an argument and without calculations, it is not at all clear how strong such an impact is expected to be. As noted by [Lemoine (2010)], Bayesian models do reach points where adding additional constraints do not narrow down uncertainty and as a sidenote, it is also possible in theory for a combined estimate of two sources to have larger uncertainty than any of the original ones. Replying to the third argument requires going through the different aspects of our sensitivity calculations: We dropped the volcanic cooling line of evidence in the first sensitivity calculation, which is justified considering the probable dependence and the fact that the constraint did not consider radiative forcing uncertainty. The narrowing down of the end result from the 20th century observational evidence is thereafter dominated by the LGM constraint and the assumptions done in deriving it. The following can probably safely be concluded from the discussion: 1. the volcanic eruption line of evidence should be dropped from the original calculation as it is likely not to be independent from the 20th century warming evidence and it does not consider radiative forcing uncertainty, 2. for the LGM line of evidence, the independence assumption is probably a reasonable one, if not exact, but in including it in the evidence one should be very critical if it is claimed that the LGM reconstructions alone provide more information than 20th century warming as paleoclimate records are known to be very uncertain [Jansen et al. (2007)]. The LGM constraint in AH06 was sketched in rough fashion and did not result from a thorough study. The fact that its maximum conveniently coincided at around 3 degrees K also contributes to the narrowing. 3. The result can be narrowed with a narrower choice of prior. This remains a partly subjective choice. However, [Lemoine (2010)] pointed out that there are weakly informative priors giving far wider posteriors than those in AH06 and in the subsequent article [Annan and Hargreaves (2009)].

Based on the research done for this thesis and surveying the literature, the consensus estimate for a climate sensitivity of 3 degrees seems like a best estimate with current knowledge. Additionally, while models seem to speak against sensitivities over 5-6 degrees at high levels of confidence, it would be desirable to explore full model uncertainty further than inter-model comparisons, at least by running through full spaces of tuning parameter values. Excluding high values of climate sensitivity from observations requires further improving the estimates, which could be made possible for example by improving estimates on historical radiative forcing in one way or the other, waiting for the instrumental record to get longer or by improving observational estimates for ocean heat uptake (for example through the Argo float measurement system [Roemmich et al. (2012)]).

4 Climate variability at different timescales

The real climate varies at all timescales from months to geological timescales for different reasons. In this thesis, timescales from years to a millennium are at focus. The previous chapter dealt with equilibrium climate change caused by greenhouse gases and other external forcings. In practice, this equilibrium will never be completely reached (here leaving out the discussion to what extent such an equilibrium actually exists) and it will not be reached smoothly as could be suggested by equation (4) if applying a constant or smoothly increasing forcing. Aerosol emissions and thereby also concentrations typically change significantly over years and decades, these being the most relevant timescales to study aerosol climate effects. Natural variability occurs due to varying external forcing in the form of changing incoming solar radiation and volcanic aerosols and because of internal variability of the climate system, which might interplay non-linearly with the external forcing. Well-known internal modes of variability in the atmosphere ocean system are e.g. El Nino Southern Oscillation, the Quasi-biennal Oscillation, North Atlantic Oscillation and Southern Annular Mode, all operating on interannual timescales. Significant internal variability also occurs at longer timescales and of particular interest in this thesis is quasiperiodic variability with a period of 50-80 years as deduced from reconstructions and modeling and showing up in the instrumental measurement record as a more regular 65-70 year oscillation, discussed in more detail in the following section.

There is a debate about how much of the cooling of global temperatures in the 1950s and 1960s was caused by aerosol forcing and how much was due to internal variability. For the North Atlantic, there are recent papers advocating both factors to be dominant [Booth et al. (2012), Zhang et al. (2013)], although acknowledging both, with the latter favoring natural variability and stating to have refuted the results of the first. The IPCC AR4 favored the explanation of aerosol forcing [Hegerl et al (2006)]. More recently, the lack of rise in global mean temperature since 1998 has been suggested to be caused except by natural variability and declining solar radiation, by increased sulfate load in Asia [Kaufmann et al. (2011)], studied in Papers **V** and **VI**. Based on the phase information presented in Paper **II**, the internal multidecadal oscillation's phase progresses in such a way that it could have contributed to both halts in the rise of global mean temperature. Meanwhile, it is also very probable that aerosol forcing contributed to cooling in the 1950s and 1960s, illustrated e.g. by the IPCC model trend used as one alternative in detrending the results. The debate of aerosol effects versus natural variability on interannual to multidecadal timescales is an important link between the different papers included in this thesis.

Decadal prediction of the climate [Latif et al. (2004), Keenlyside et al. (2008)] is attracting more and more societal and scientific interest. Success of decadal prediction is dependent on understanding the processes involved, and knowing their relative strengths could be applied in decadal prediction of the climate.

In studying climate variability at different timescales, it can be noted that defining the terms climate, climate change and climate variability exactly is a non-trivial task, is usually not done explicitly but assumed to be clear enough from the context. The definition of climate according to the World Meteorological Organisation [WMO (2013)] is:

Climate in a narrow sense is usually defined as the "average weather," or more rigorously, as the statistical description in terms of the mean and variability of relevant quantities over a period of time ranging from months to thousands or millions of years. The classical period is 30 years, as defined by the World Meteorological Organization (WMO). These quantities are most often surface variables such as temperature, precipitation, and wind. Climate in a wider sense is the state, including a statistical description, of the climate system.

The climate system in the WMO definition means the system consisting of the atmosphere, the hydrosphere (the ocean, lakes, rivers, groundwater and water in the atmosphere), the cryosphere (frozen water), the land surface (or litosphere) and the biosphere and their interactions (thus, no circular argument in using it in the definition of climate). Further, climate variability is defined as the mean state and variability of the climate at all spatial and temporal scales beyond individual weather events and climate change as a statistically significant change in the mean state or the variability persisting for an extended period. Climate change is alternatively defined by the United Nations Framework Convention on Climate Change as a change of climate attributed directly or indirectly to human activity through an alteration of the composition of the atmosphere [UNFCCC (2013)]. Thus, the definitions may vary somewhat from source to source, and the WMO definition is not totally explicit in what kind of means (spatial/temporal, resolution) and other statistical quantities from an infinite amount of possibilities are preferable in describing the climate, though the WMO climate normals are defined as 30-year periods [WMO (1989), WMO (2007)]. The lack of an exact general definition is not a weakness of the definition but an illustration of the inevitable challenges in the task to conseptualize the complex climate system. The climate normals have also been critically reviewed and alternatives proposed in particular to account better for climate change that continuously makes 30-year normals obsolete even before they are calculated [Milly et al. (2008), Arguez et al. (2011)].

The statistical descriptions of the climate presented in this chapter can be considered as describing the climate in the wider sense of the WMO definition. The statistical descriptions are to our knowledge partly new additions to the climate literature and we think that for example studying the phase progression of an oscillation in a spectral analysis or a spectrum obtained can provide useful information compared to alternative statistical descriptions.

4.1 Fourier analysis with a flexible time window, or Welch's method

Spectral analysis, representing a time series as sines and cosines or more complex oscillatory functions, provides tools for separating possible periodic functions, or more generally, variability at different timescales (or spatial scales), with care required in interpreting the results nonetheless. Spectral analysis of climate data has involved a wide selection of methods [Yiou et al (1996)], including wavelet analysis, singular spectrum analysis, multitaper analysis and other Fourier methods.

The *Fourier transform* represents a time series as a sum of sines and cosines. We chose this traditional method with a modification for the spectral analysis to facilitate understanding for non-specialists in time series analysis, who are interested in climate variability and because its pitfalls and limitations are better known and more extensively studied than for other methods. Well-known potential issues are aliasing, spectral leakage, only resolving certain discrete frequencies and frequency and amplitude modulation. Aliasing occurs from frequencies above the highest resolved frequency, called the Nyquist frequency, to the other frequencies. Aliasing is not likely to be an issue in analysing the annual mean temperatures, because the amplitudes in the frequency range $f > 1/(2y)$ are very small (except for the annual cycle that gets averaged away in the analysed annual means), as can be seen in Figure 4 in Paper **III**. Spectral leakage [Harris (1978)] may occur, but with varying the length of the time window to suit an oscillation at interest, we can combat this problem, especially when its amplitude is large compared to amplitudes of oscillations at other frequencies. Similarly, we can also resolve much more frequencies than usual since we are doing Fourier transforms using different lengths of time window. Effects of amplitude modulation slower than a full period of an oscillation are also resolved by the moving time window, but effects of possible frequency modulation will be visible in the results, a limitation of the method to be kept in mind. In Paper **II**, we evaluate the magnitude of different possible errors caused by the method by several sensitivity calculations.

In studying the short global mean instrumental temperature record from the year 1850 to present, the standard Fourier analysis was also preferred as the rectangular window function allows us to retain all the data with full weight. *Multitaper analysis*, whereby the data is multiplied with a number of non-constant window functions, treats data close to the endpoints with smaller weight. An important advantage of the chosen method is its flexibility in that we can adjust the time window to be a multiple of the period of an oscillation at interest. Another important advantage is being able to track the amplitude and phase progression of a certain component. This tracking is done for the sinusoidal wave with a period of 66 years in Paper **II**, found to correspond closely to the the time window, that gives the maximum amplitude for the largest amplitude. Fitting the time window this way, spectral leakage is reduced. We can also average over several spectra obtained by different parts of a certain time series as the quantities $S(f) \propto |\hat{T}(f)|^2$ and $|\hat{T}(f)|$ are non-negative and phase differences will not cause cancellation in the averaging. This is utilised in Paper **II** to evaluate the amplitude of the quasiperiodic oscillation with a period of 50-80 years and in Paper **III** to study the full spectrum. We are not aware of any other climate study performing such averaging. The averaging improves the statistical estimate of the amplitude at a certain frequency at the expense of decreasing the spectral resolution, illustrated by the narrowing of the 95% confidence intervals in Paper **III**. Averaging over spectra obtained from different segments was probably first done by Welch in 1967 to facilitate computation, core storage and stationarity testing [Welch (1967]. It would be possible to extend this averaging to multiple time series for an ensemble of simulations to further improve the statistics. If one assumes that the processes generating climate variability remain constant or nearly constant through a time series, averaged spectra describe more characteristic climate variability at different timescales than single stochastic realisations and can thus be considered important climate variables. Comparing these averaged spectra between datasets will then be more meaningful than comparing spectra of single time series. This is done in Paper **III** to compare climate variability in the earth system model used and in measurements. As further work, an intercomparison of interannual variability of global mean temperature in the CMIP5 models has been started.

4.2 Quasiperiodic variability with a period of 50-80 years

An early study reporting the finding of a 65-70 year oscillation in the global climate system was [Schlesinger and Ramankutty (1994)], using singular spectrum analysis to analyse the instrumental temperature record. Several similar results have been reported thereafter [Delworth et al. (1997), Delworth and Mann (2000), Semenov et al. (2010)], especially for the North Atlantic, where the oscillation has been named Atlantic Multidecadal Oscillation (AMO) [Kerr (2000), Enfield et al. (2001), Latif et al. (2004), Knudsen et al. (2011), Wei and Lohmann (2012)]. The consensus is that the oscillation is generated internally by the atmosphere-ocean system, but probably affected by external forcing [Otterå (2010)]. The quasiperiodic oscillation has also been found in tree-ring reconstructions [Gray et al. (2004)].

Some mechanisms contributing to or producing a quasiperiodic multidecadal oscillation have been discovered from model data [Dima and Lohmann (2007)]. In the North Atlantic, important processes are related to the meridional overturning circulation (MOC) and

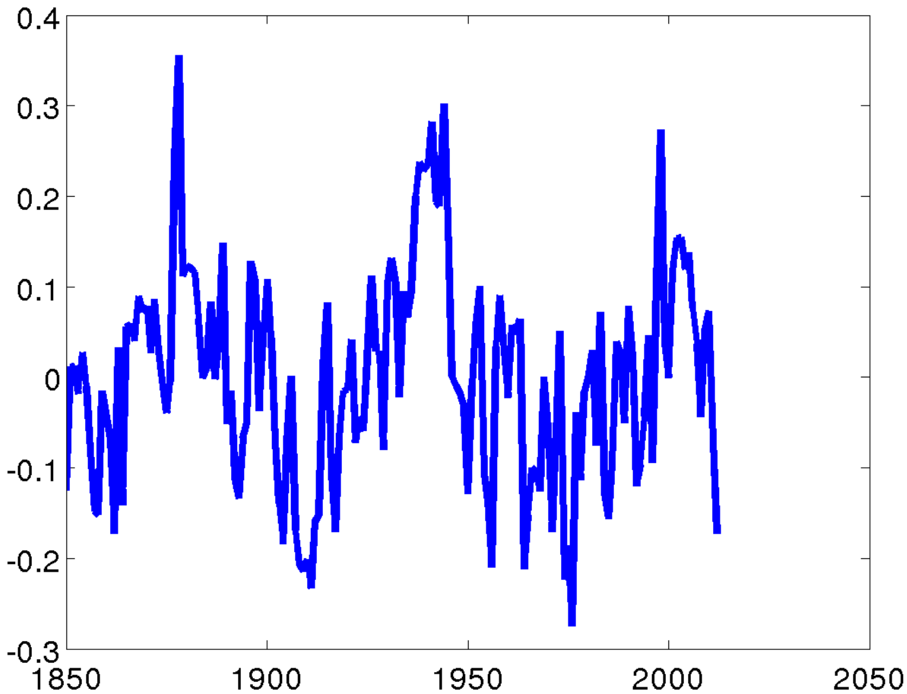

Figure 2: The instrumental global annual mean temperature record from the HadCRUT3 dataset detrended by a quadratic fit (degrees K).

salinity anomalies in the important downwelling regions of the Gulf stream north and east of Greenland. The negative salinity anomaly feedback could come either from the Arctic as freshwater or sea ice export through the Fram Strait [Delworth et al. (1997), Delworth and Mann (2000), Jungclaus et al. (2005)] or from the tropical Atlantic through moving of the intertropical convergence zone (ITCZ) [Vellinga and Wu (2004)]. In paper **II**, we find medium high correlations for filtered data supporting both possibilities.

This quasiperiodic oscillation might also be relevant for Finland, illustrated in Figure 3, showing 20-year running averages of the AMO index derived from the HadSST2 dataset [Rayner et al. (2006)] and from measured mean temperature in Finland [Tietäväinen et al. (2010)]. The filtered AMO signal and the Finnish mean temperature follow each other quite closely. This suggests that the approach utilising observed sea surface temperatures in model initialization [Latif et al. (2004), Collins et al. (2006), Keenlyside et al. (2008)] that has given some predictability for the North Atlantic could be tried also for Finland. Naturally, more adjustments would be needed to catch the local features of variability at faster timescales in Finland.

The spatial distribution corresponding to the quasiperiodic oscillation looks quite different depending on how it is extracted. Zanchettin et al. (2013) discuss this issue for Atlantic multidecadal variability in more detail by going through patterns obtained by three different definitions for describing Atlantic multidecadal SST variability, two based on spatial averages and one based on the first empirical orthogonal function of North Atlantic SSTs and reached clearly different patterns with the different methods. In Paper **II**, we derived spatial distributions with two methods: maximum minus minimum and local discrete Fourier transform, again leading to somewhat different results. In general, though, northern ocean and continent areas tend to have

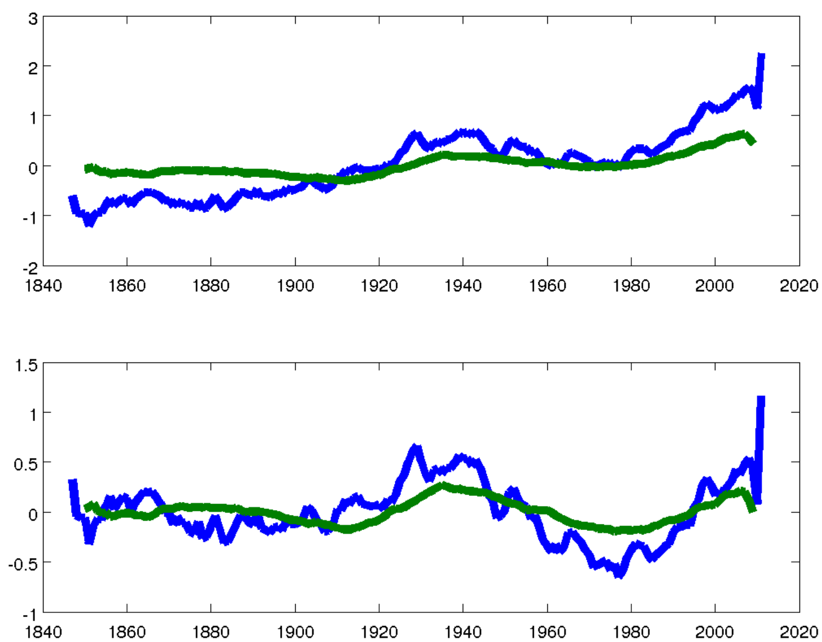

Figure 3: Anomalies of Finnish mean temperature (blue) and AMO index (green) in measurements, without detrending (above) and with quadratic detrending applied (below) (degrees K).

larger positive anomalies in such distributions than other regions. As will also be corrected in the Erratum of Paper **II** below, in the local Fourier transform estimates we mistakenly used the term 'amplitude' in place of 'coefficient' in the context of Figs. 4 and 7. This method gives a value zero if the local temperature anomaly has a 90 degree phase difference with the refence index and a negative value for phase differences between 90 and 270 degrees. A new map showing the absolute value of the amplitude and disregards the phase, is shown in Figure 4. The Barents Sea, the North Atlantic, areas near the Bering Strait and the Amundsen Sea have the highest amplitudes (all areas with relatively large climatological temperature gradients). Local amplitudes in Finland are also relatively high.

A 50-70 year oscillation in measured temperature in the North Pacific was reported by [Minobe (1997)]. Multidecadal variability in the North Pacific and North Atlantic in the Kiel climate model were studied in [Park and Latif (2010)], where it was concluded that the memory of the North Pacific low-frequency oscillation is related to the subtropical gyre, while the North Atlantic low-frequency oscillation is related to the meridional overturning circulation. It remains to be seen whether the 50-80-year oscillations are regional and independent in nature or whether the oscillation is a hemispheric or global phenomenom. While there have been arguments that the North Atlantic could have the ability to drive multidecadal variability in the global climate [Zhang et al. (2007)], others have speculated that the oscillation might be hemispheric, or even global in extent [Semenov et al. (2010)]. [d'Orgeville and Peltier (2007)] studied measured $\sim 60-$year temperature variability in the North Atlantic and North Pacific, found that the North Atlantic variability leads that of the North Pacific, and speculated that variability in the two ocean basins could be connected. Data that could be used in such studies is plotted in Figure 5 showing mean temperature in the North Atlantic (AMO index area $0-60^\circ$ N, 70° W -0° E)

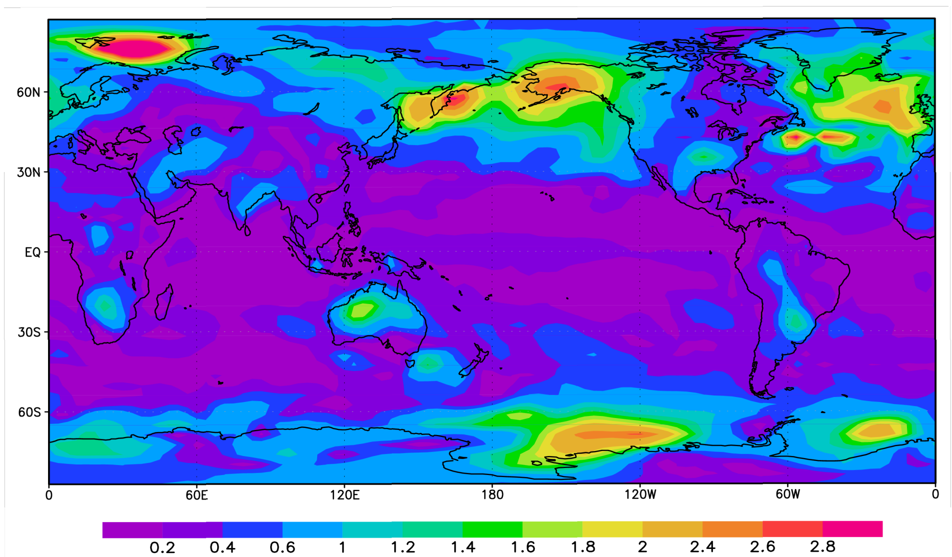

Figure 4: Local amplitude of 66-year oscillation in discrete Fourier transform in unforced earth system model simulation (degrees K).

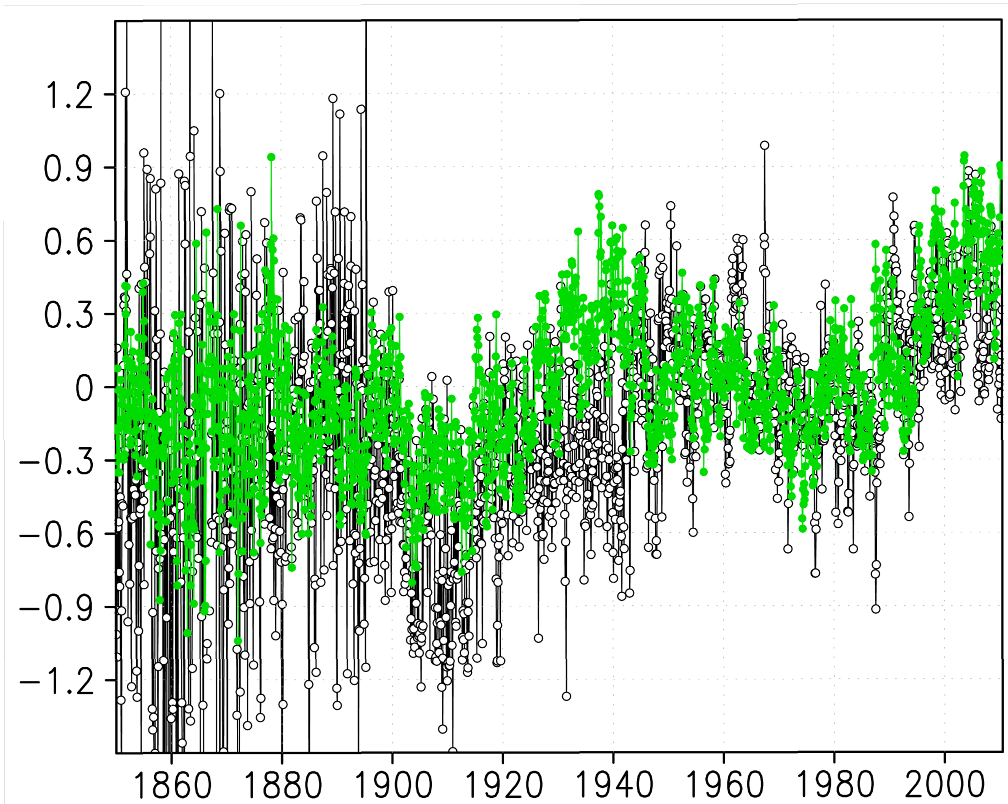

Figure 5: Sea surface temperature in the North Atlantic (north of 0° N; green) and in the North Pacific (north of 30° N;black) in the HadSST2 dataset (degrees K).

and in the North Pacific ($30 - 60°$ N, $120°$ E $- 120°$ W) from the measured HadSST2 dataset [Rayner et al. (2006)].

Choosing the terminology related to the topic, including the title of this section, is not straightforward. There is no consensus in the literature as to how regular the oscillation is and for the (North) Atlantic some prefer Atlantic Multidecadal variability (AMV) [Vincze and Jánosi (2011), Zanchettin et al. (2013)] over Atlantic Multidecadal Oscillation (AMO). This could be motivated as the oscillation is not completely regular, but on the other hand for example the phase progression plots in Paper **II** show quite regular progression in the instrumental record, which would perhaps make AMV too general a term to describe the oscillatory behavior since 1850. Figure 2 shows the instrumental temperature timeseries detrended with a quadratic trend and by visual inspection a relatively regular amplitude and length of the multidecadal oscillation.

4.2.1 Erratum to Paper II

As mentioned and discussed in the section above, the term 'amplitude' was mistakenly used in place of 'coefficient' in the context of Figs. 4 and 7 in Paper **II**.

4.3 The full spectrum and power laws

This section deals with the full spectrum of global temperature anomalies. In addition to the 50-80 (or 65-70)-year cycle having been found, significant amplitudes at periods of 20-30 years in the North Atlantic have been found in several GCMs [Timmermann (1998), Cheng et al. (2004), Dong and Sutton (2005), Frankcombe (2010)].

A power spectrum can follow a power law:

$$\mathrm{S(f)} \propto \mathrm{f}^{-\beta}, \tag{9}$$

where $\mathrm{S(f)} = |\hat{\mathrm{T}}(\mathrm{f})|^2$ or $\mathrm{S(f)} =< |\hat{\mathrm{T}}(\mathrm{f})|^2 >$, with $< ... >$ denoting averaging when it is used. A power law shows up as a line in a log-log plot of frequency vs. power S(f). Power laws are found for a wide number of spectra and frequency distributions [Clauset et al. (2009)], with the energy wavenumber (spatial) spectrum of fully developed turbulence in the inertial subrange of the spectrum perhaps most well known [Tennekes and Lumley (1972)]. Power laws are usually intuitively characterised as the system lacking a characteristic length or time scale in the range, where the power law is valid. Self-similar *fractals* have a constant exponent in the scaling of the frequency distribution when going to smaller and smaller scales, while *multifractality* means that the exponent is changing with scale.

Paper **III** goes through earlier power-law fits to temperatures in climate data in a wide range of timescales and proceeds to describe power-law fits made to COSMOS earth system model data. Of earlier results, [Huybers and Curry (2006)] present results for timescales ranging from monthly variations to timescales of hundreds of millennia and obtain different exponents for different frequency ranges. In our results, after averaging, we find two frequency ranges where power laws fit well: from multidecadal ($\sim 50 - 80$ years) to El Nino ($\sim 3 - 6$ years) timescales and from El Nino timescales up to the Nyquist frequency. Averaging is essential to narrow down the confidence interval of the power estimate S(f) for each frequency and to see the spectral form clearly.

We studied power laws in temperature anomalies, but also for example multifractality in rainfall and application of such found laws to study rainfall extremes has been performed in previous studies [Veneziano et al. (2006)].

Based on Paper **III**, averaged spectra seem to be a good way to compare variability at different timescales for different data sets representing conditions of constant or nearly constant climatic conditions, especially when time series are long enough to provide large enough samples of fluctuations at timescales of interest. The spectra in the COSMOS model seem to follow power laws quite well for frequencies between about 1/(65 y) and 1/(6 y), and the forced simulation spectra correspond to estimates done for the instrumental temperature record. The results can also be sensitive to the choice of frequency range and averaging reveals the power-law ranges better than spectra of single time series.

5 Aerosols and the climate

Aerosols affect the climate directly, by scattering and absorbing sunlight and by scattering, absorbing and emitting longwave radiation [Haywood and Boucher (2000)], and indirectly, through their interactions with clouds [Lohmann and Feichter (2005)].

Scattering and absorption can be be calculated by the Mie solution of the Maxwell equation in the case of spherical particles. The particles are very often assumed to be spherical in climate models and the Mie solution thus applied, but radiation calculations for more complex shapes, such as ellipsoids have also been made and implemented in climate models [Räisänen (2012)].

The possibility of aerosols increasing cloud droplet number concentration and thereby the reflectance of a cloud is called the *first indirect effect* [Twomey (1959), Twomey (1977)] and the possibility of aerosols increasing cloud lifetime and thereby reflection of sunlight is called the *second indirect effect* [Albrecht (1989)]. Light absorption by aerosols can cause cloud droplets to evaporate after warming of the droplets and their surroundings, thereby reducing the cloudiness, referred to as the *semidirect effect*. In their review, [Lohmann and Feichter (2005)] also list the following possibilities for aerosols to affect the climate through their effects on clouds: the thermodynamic effect (smaller droplets delay freezing), the glaciation indirect effect (more ice nuclei increase precipitation efficiency), the riming indirect effect (smaller droplets decrease riming efficiency) and the surface energy budget effect (less radiation absorbed at the surface). In short, the indirect effects include the possible ways that aerosols can effect the cloud properties through the modification of cloud droplets. The direct and indirect effects may also modify the climate even more indirectly through feedbacks operating differently than feedbacks due to forcings that operate globally. Some such possible effects are discussed in Section 5.3. below.

Estimates for the radiative direct effect of aerosols tend to have quite large relative uncertainty. The IPCC AR4 5 to 95% confidence interval was -0.5 ± 0.4 W/m^2 (all-sky at the TOA) and scientific understanding was rated medium-low [Ramaswamy et al. (2001)]. [Myhre (2009)] reported to have constrained the negative forcing value to below 0.5 W/m^2 by combining satellite observations and modeling, and the new AeroCom Phase II modeling intercomparison reported results between -0.58 and -0.02 W/m^2 [Myhre et al. (2012)]. Advanced satellite products include such variables as fine-mode fraction and effective radius in addition to AOD. Remote sensing of aerosol absorption, on the other hand, is currently limited to a semi-quantitative UV index [Moosmüller et al. (2009)]. A large part of the uncertainty of direct-effect estimates in modeling studies also comes from absorption as emissions of carbonaceous aerosols are more uncertain than those of SO_2 and the absorption depends on assumptions related to mixing state of the aerosols, particle compositions, vertical distributions and shape, age and wavelength [Bond and Bergstrom (2006), Stier et al. (2007), Andreae and Gelencsr (2006), Vignati et al. (2010), Bond et al. (2013)].

Estimates of indirect aerosol radiative forcing both from modeling and observations are uncertain and the scientific understanding of the cloud albedo effect is deemed to be low by the IPCC. The best estimate for the radiative forcing due to this first indirect effect in the AR4 was -0.7 W/m^2 with a 5 to 95% range of -0.3 to $-1.8\mathrm{W/m^2}$ [Ramaswamy et al. (2001)].

Sulfates have a strong radiative forcing, possibly in magnitude second only to CO_2, but opposite in sign [Forster et al. (2007)]. Sulfates form mainly heterogeneically, e.g. in cloud water, or from emitted SO_2 being photo-oxidized by UV radiation and converted into sulfate particles [Kulmala and Kerminen (2008), Laaksonen et al. (2008)]. Black carbon has a positive net forcing and might be a major contributor to the total anthropogenic radiative forcing [Jacobson (2001), Bond et al. (2013)]. As a side-note, the term black carbon is not totally clear, but it is normally defined through optical properties of the aerosols, although this might not be a complete characterisation and wavelength dependence of absorption through the introduction of 'brown carbon' alongside black carbon has been suggested as one solution to improve the description of light-absorbing aerosols in climate research [Andreae and Gelencsr (2006)]. Other anthropogenic aerosol species showing significant impacts on the climate are nitrates formed through oxidation of nitrogen oxides NO_x [Kulmala et al. (1995), Makkonen et al. (2012)], organic carbon and industrial dust [Forster et al. (2007)]. Sulfate, carbonaceous and nitrate aerosols as well as for example pollen, spores and bacteria are also emitted by natural sources (biogenic, volcanoes, wildfires etc.). Additionally, mineral dust and sea salt emissions are caused by surface winds and their average concentrations may be changing as a result of changes in the climate or land-use [Forster et al. (2007)].

Light-absorbing carbonaceous aerosols are produced in biomass and fossil fuel burning, wildfires and biogenic and other processes and have significant climate impacts in many regions of the world. Atmospheric brown clouds with high aerosol optical depths and high absorption were identified by Ramanathan et al. (2008) to exist in East Asia, the Indo-Gangetic Plain in South Asia, Southeast Asia, Southern Africa and the Amazon Basin. In the following two sections, aerosol-climate interactions of absorbing and other aerosols in Asia are considered. Effects of black carbon aerosols in Arctic areas and Finland have become of interest as black carbon through absorbing light in the air or when deposited on snow or ice can have significant climate effects and may under suitable circumstances be transported from far to the Arctic [Hyvärinen et al. (2011c)]. As an example, in spring 2006, large amounts of absorbing aerosols emitted by agricultural fires in eastern Europe were transported to the Arctic [Stohl et al. (2007)]. McConnell et al. (2007) reported Arctic black carbon forcing estimates from 1788 to present and found that industrial emissions have caused ice core black carbon concentrations to be seven-fold compared to preindustrial values since about 1850 and peaking in 1906 to 1910 when surface forcing was estimated to be about 3 W/m^2. The concentrations were smaller after 1951 than before, but increased again toward the end of the century.

Smith et al. (2011) estimated anthropogenic sulfur dioxide emissions for the years 1850-2005 and [Junker and Liousse (2008)] estimated anthropogenic carbonaceous aerosol emissions for the years 1860-1997. In the last few decades, emissions of sulfate, nitrogen oxides and carbonaceous aerosols have decreased in Europe and North America and lately increased in Asia, with global emissions having peaked in the early 1970s. Carbonaceous aerosol emissions were dominated by North America, Germany and the United Kingdom until 1950, after which the USSR, China and India became substantial contributors. Global black carbon emissions have mostly increased during the 20th century, with a local peak around 1910 and some decrease after the 1980s, but estimates have relatively large uncertainty and there is scatter between inventories [Junker and Liousse (2008), Ito and Penner (2005)]. Skeie et al. (2011) estimated a historical timeseries of radiative forcing and uncertainty for the present-day forcing. Seemingly there is no study concerning the historical development of radiative forcing uncertainty, but one would be desirable for improving climate sensitivity estimates and detrending for the anthropogenic signal in studying internal climate variability.

5.1 Modeled and observed aerosol distributions and optical properties in India and China

India and China are presently experiencing a high load of aerosols, that have dimmed their surface by at least 6% since preindustrial times [Ramanathan et al. (2008)]. In addition to their climate effects, the aerosols have adverse effects on human health and the environment. Negative health effects are caused by exposure to aerosols both indoors and outdoors and include cardiovascular and pulmonary effects leading to chronic respiratory problems, hospital admissions and premature deaths. Surface ozone formed from precursors emitted simultaneously with aerosols in combustion have through experiments been estimated to reduce the harvests of the wheat, rice and legumes by $10 - 40\%$ at levels of 30-45 ppbv, typical values for present-day Asia [Ramanathan et al. (2008)].

Plenty of aerosol station measurements in India and China have been done [Lelieveld (2001), Carrico et al. (2003), Franke et al. (2003), Monkkonen et al. (2004), Monkkonen et al. (2005), Muller et al. (2006), Nakajima et al. (2007), Kanaya et al. (2008), Beegum et al. (2009), Moorthy et al. (2008), Hyvärinen et al.(2009), Hyvärinen et al.(2010), Hyvärinen et al. (2011a), Hyvärinen et al. (2011b), Zhang et al. (2012)]. Satellite measured MODIS aerosol optical depths (AOD) and ground-based network AERONET AODs were used for comparison with models in Paper **V**. The modeled load of carbonaceous aerosols is uncertain, judging from much higher uncertainties in emissions than for SO_2 [Ohara et al. (2007), Lu and Streets (2011)], from high spatial and temporal heterogeneity of measured black carbon concentrations [Beegum et al. (2009)] and because of poor correspondence of modeling results with satellite AOD measurements in the Ganges valley with high absorption [Paper **V**, Ganguly et al (2012)]. Nair et al. (2012) performed a comprehensive validation study of South Asian aerosols in a regional climate-aerosol model with observations at AERONET stations, by the MODIS and MISR satellite instruments and at Aerosol Radiative Forcing over India (ARFI) network stations. The BC mass concentrations were underestimated by a factor of 2 to 5 at most stations. The many measurement and modeling efforts show promise in constraining the BC concentrations better in the near future.

In modeling studies on Indian aerosols, anthropogenic aerosol concentrations have a maximum in the winter and natural aerosol concentrations have a maximum in the summer, resulting in a total average AOD with maxima both in summer and in winter [Adhikary et al. (2007), Carmichael et al. (2009), Paper **V**], qualitatively corresponding to MODIS measurements [Ramachandran et al.(2007), Paper **V**]. Figure 6 illustrates the large-scale properties of the climate in India and China by showing the average monthly 10-meter wind speed and precipitation in the longitude-latitude boxes taken to represent India and China in an ECHAM5-HAM simulation. Precipitation and 10-meter wind speed are largest in the summer, which are connected with the larger washout of anthropogenic aerosols and larger emissions of natural aerosols in the summer. In China, the seasonal variations are not as large, with a spring maximum, mainly caused by dust aerosols originating in Mongolia and northern China and a second maximum in AOD in fall mainly explained by a high contribution from sulfate and its hygroscopic growth to

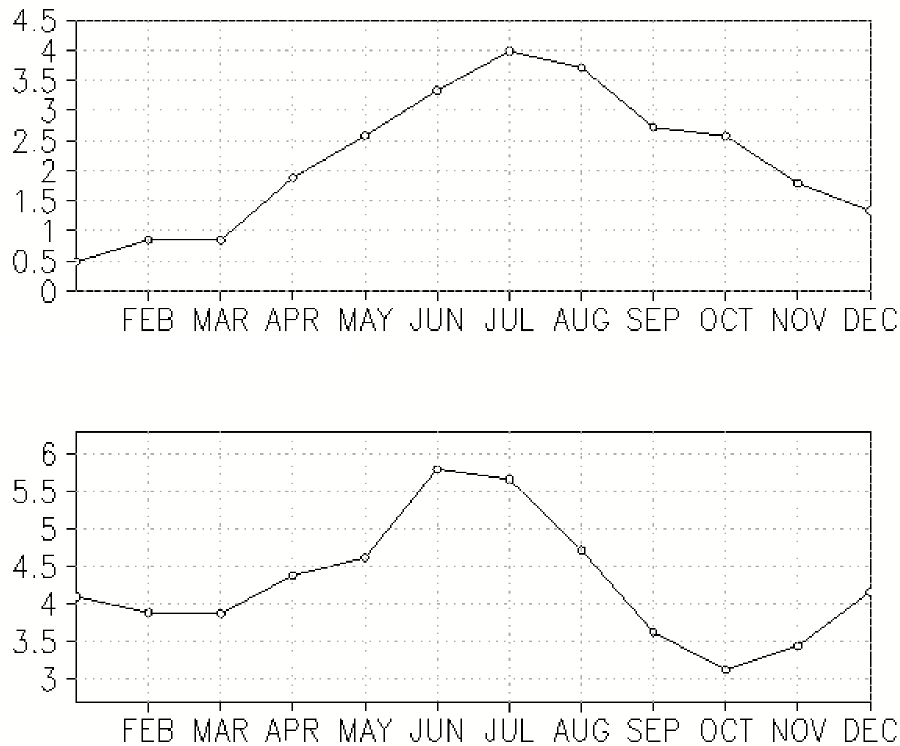

Figure 6: Mean precipitation (above; mm/d) and 10-meter wind speed (below; m/s) in India (represented by longitude-latitude box $5 - 35°$ N, $65 - 90°$ E), average in 5-year simulation with GAINS emissions and without aerosol cloud activation.

AOD, again qualitatively consistent with MODIS data [Paper **V**].

5.2 Climate effects of aerosols in India

India and southern Asia in general, the home of over a billion people, is dependent on its climate to produce an environment suitable for food production. Monsoon rainfall and river runoff are important variables. Glacier melt through heating at elevated levels and black carbon deposition on the glacier surfaces may diminish the glaciers and threaten runoff in the dry season [Ramanathan et al. (2008), Kehrwald et al. (2008)]. Both positive and negative effects of aerosols on monsoon rainfall have been proposed. The elevated heat pump (EHP) hypothesis [Lau et al.(2006), Lau and Kim(2006)] states that light-absorbing aerosols increase atmospheric heating and thereby convection at elevated levels, leading to enhanced rainfall. The solar dimming mechanism (SDM) reducing the temperature gradient in the monsoon season could, on the other hand, lead to diminished rainfall [Chung and Ramanathan(2006)]. Solar dimming might also reduce precipitation through reduced evaporation resulting from diminished solar radiation absorbed at the surface. Very strong radiative atmospheric heating by absorbing aerosols has been measured on urban sites in India [Tripathi et al. (2005)]. Sreekanth et al. (2007) reported high forcing values in the winter and moderate forcing values in the summer from a tropical station in eastern India, with net forcing at the top of the atmosphere being positive in all seasons.

Correspondingly with the aerosol distributions studied in Paper **V**, we find in Paper **VI** that total aerosol radiative forcing in India is largest in summer and anthropogenic forcing is largest in winter. In Paper **VI**, we also find that in northern India the elevated heat pump mechanism works to increase rainfall and that solar dimming works to reduce rainfall. Ramanathan et al. (2005) report a weakening of the climatological SST gradient in March to June from having been from about 303 K at 20° N on average in the Arabian Sea and Bay of Bengal to about 301 K at the equator by about 0.5 K since the 1950s and a decreasing trend in monsoon precipitation. As sea surface temperatures are prescribed in ECHAM5-HAM, we artificially decreased the gradient similarly in two model experiments. If the observed SST gradient reduction over the decades is caused by aerosols, then the best model estimate is that rainfall is decreased by $\sim 20-25\%$ in northern India in July and August because of the aerosol load. The corresponding spatial distribution for the whole India is shown in Figure 7. There is large uncertainty in the model estimates for aerosol effects on rainfall as the phenomena are complex and as modeling of aerosol-cloud droplet interactions is uncertain due to their complexity and small, sub-grid scale [Roelofs et al. (2006)]. An illustration of the uncertainty is Fig. 9 of Paper **VI**, showing that depending on whether aerosol-cloud interactions are applied or not, the rainfall in a region can change even by a factor of 2.

5.3 Climate effects of volcanic eruptions

Volcanic aerosols influence the climate. Emitted sulfur dioxide is oxidized and converted to sulfate aerosols in the stratosphere, where the particles have a long lifetime due to lack of removal by precipation, typical e-folding times being 12-14 months. In the direct aftermath of an eruption, also coarser particular ash can have an effect on the climate. After an eruption large enought to cause climate effects and of typical size, the sulfate aerosols are removed from the stratosphere in 2-3 years, but oceans have a longer memory of the eruptions [Stenchikov et al. (2009)]. Volcanic eruptions are considered essential in triggering the little ice age in approximately the years 1250-1850, alongside the probably smaller influence of solar radiation changes [Hegerl et al (2006), Jungclaus et al. (2010)]. Radiative forcing uncertainty of volcanic aerosols is typically very large, however, and often not even estimated [Andronova et al. (1999), Wigley et al. (2005)]. In general, the previous IPCC assessment report evaluated scientific understanding of radiative forcing caused by volcanic eruptions to be low [Forster et al. (2007)].

Very large eruptions have self-limiting effects when particles coagulate on each other and get removed from the stratosphere faster than they would

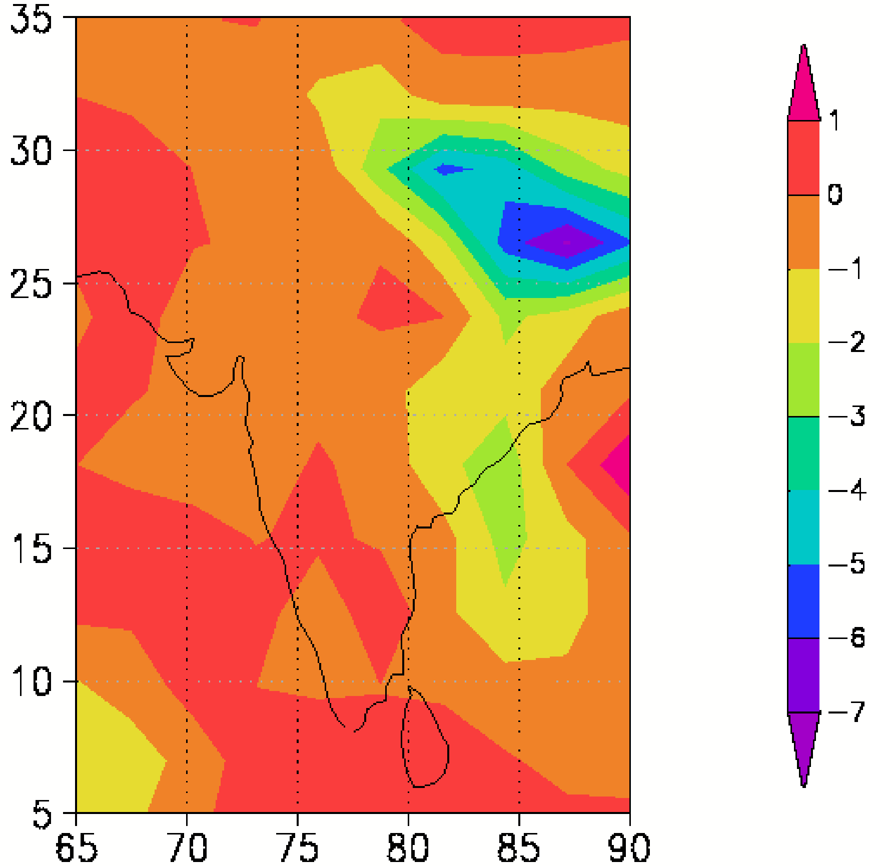

Figure 7: Difference in mean rainfall (mm/d) in July and August between simulation with GAINS anthropogenic aerosol emissions and artificially cooled sea surface in the northern Indian Ocean and the simulation with no anthropogenic emissions and standard sea surface temperatures, cloud activation included in both simulations and 5-year means used in plot.

otherwise [Pinto et al. (1989), Timmreck et al. (2009)], but also a prolonged lifetime of stratospheric sulfate due to dehydration caused by SO_2 works as an opposite mechanism [Bekki (1995)]. Timmreck et al. (2010) studied the very large Younger Toba Tuff eruption that happened 74000 years ago and has been proposed to be connected to a near-extinction of the modern human. They incorporated both the self-limiting effects of aerosol microphysics and the dehydration of the stratosphere in the ECHAM family models. They reported a shorter duration and a three times smaller value for the cooling of global mean temperature than in earlier studies on the Younger Toba Tuff eruption.

Effects of volcanic eruptions on internal modes of variability such as El Nino Southern Oscillation, the Quasi-biennal Oscillation, the North Atlantic Oscillation and the Arctic Oscillation have been studied in several publications [Robock et al. (1995), Stenchikov et al. (2002), Stenchikov et al. (2004), Emile-Geay et al. (2008)]. Otterå et al. (2010) conclude that volcanic eruptions have affected the phasing of North Atlantic multidecadal variability discussed above in Section 4.2. in the past through deterministic effects of volcanic eruptions on the meridional overturning circulation. Therefore, the effects of their result on decadal climate predictability in the North Atlantic would be twofold: limiting on one hand, as eruptions can currently not be predicted on climatic timescales [Sparks et al. (2003)], and on the other hand, would improve chances for short-term climate prediction after an eruption has happened.

In addition to being of interest for climate understanding in general, Northern hemisphere volcanic eruptions fell under the spotlight with the eruptions of Eyjafjallajökull in 2010 and Grimsvotn in 2011 [Kerminen et al. (2011)]. Their climatic effects were most likely nearly negligible, but in addition to causing disruptions to air traffic, especially the Eyjafjallajökull eruption was interesting, because, due to a history of simultaneous activity, it might anticipate an eruption of the bigger volcano Katla, situated 25 kilometers away [Sturkell et al. (2003)]. Paper **IV** was largely motivated by this possibility. Generally, mid- and high-latitude eruptions can have significant climate effects [Robock (2000), Oman et al. (2005), Oman et al. (2006a]. One example from the past is the eruption of the Icelandic volcano Laki in June 1783 [Oman et al. (2006b)], that was followed by an exceptionally cold European winter 1783/4 [Thordason and Self (2003)].

The approach taken in Paper **IV** is to study a typical climate response of a northern hemisphere mid- or high-latitude volcanic eruption. The estimates can not be directly converted to a prediction of what the climate impact of a Katla eruption would be, as the volcanic aerosols are only resolved in four latitude bands. In addition to not resolving the local effects in the immediate aftermath of an eruption, an important difference of these coarsely resolved simulated eruptions and Katla eruptions is that the insolation at Katla's latitude ($\sim 64°$N) varies much more with season than insolation at mid-latitudes. Thus, real eruptions outside the summer season would, compared to our simulations, likely have a significantly smaller or even negligible climate effect through stratospheric sulfate aerosols, as such aerosols could not be formed through photooxidation [Kravitz and Robock (2011)].

Our best estimate for a typically sized northern hemisphere mid- or high-latitude eruption is -0.19 K for the maximum cooling and -0.095 K for average cooling during the 21 months following the eruption. The time of eruption does play a role. When lumping all the studied eruptions from the ensemble together, the negative precipitation anomaly signal was statistically significant at the 90% level and similar in magnitude compared to the temperature signal as projections for increased precipitation due to greenhouse gas warming: $\sim 2\%\mathrm{K}^{-1}$. Effects on carbon dioxide concentrations were relatively small and the monthly anomalies were not statistically significant at high levels.

6 Review of papers and the author's contribution

Paper **I** deals with the uncertainty of climate sensitivity deduced from observations. It comments on an earlier paper from literature and concludes that constraining climate sensitivity and ruling out its high values is not as simple as described in that paper. I was responsible for the climate-related physics, performed the calculations and wrote most of the paper.

Paper **II** studies 50-80 year quasiperiodic variability in the instrumental temperature record and in unforced and forced millennium simulations made with the COSMOS earth system model. Estimates for the amplitude of the quasiperiodic oscillation are given. For measurements, three different trends are used in detrending. Phase and amplitude progression were analysed. The average amplitude was larger and the North Atlantic sea surface temperatures are more synchronized with global mean temperature in the simulation including external forcing. I developed the modifications to the spectral analysis method, performed the analyses and wrote the paper with input and important ideas coming from my coauthors.

Paper **III** studies power-law behavior in millennium simulations. Power laws are found to describe the temperature spectrum in the frequency range from the multidecadal peak to the El Nino frequencies. These power laws are found by averaging many time series obtained by splitting up the original time series and both in unforced and forced simulations, with the exponent in the latter case corresponding to that estimated from the instrumental temperature record. I performed the analyses and wrote the paper with support and input from my coauthors.

Paper **IV** studies climate effects of northern hemisphere volcanic eruptions in the COSMOS earth system model. Estimates for maximum cooling and average cooling during the 21 months following eruptions are provided as well as estimates for radiation, precipitation and carbon dioxide concentration anomalies. The large ensemble allows us to assign 90% statistical significance to the decrease in precipitation. I wrote the paper based on Heidi Meronen's master's thesis, which I supported through technical support and advice. I also did additional analyses not included in the master's thesis.

Paper **V** studies spatial distributions and seasonal cycles of aerosol mass distributions and optical properties in India and China. The succesful evaluation against MODIS AOD data, other models and some measurements gave confidence in the climate-aerosol model to be applied in Paper **VI** and facilitated separating the effects of anthropogenic and natural aerosols. A finding made possible by the aerosol treatment in the model was that sulfate aerosols on average get transported further away from their sources than carbonaceous aerosols, and therefore sulfate concentrations are on average higher in remote locations. I did the simulations and the post-processing with help from my coauthors, drew most of the conclusions and wrote most of the paper.

Paper **VI** studies aerosol climate effects in India with ECHAM5-HAM, partly building on Paper **V**. Estimates on radiative forcing and temperature and precipitation response are provided relying on nine simulations with different model setups. A brief comparison with black carbon observations is included. The total negative aerosol forcing at the top of the atmosphere is largest in summer while anthropogenic forcing is largest in winter. Temperature anomalies are mainly negative with some exceptions, such as northern India in March-May. Aerosol effects on rainfall are studied with and without aerosol cloud activation, with and without aerosol light absorption and with and without modified sea surface temperatures. Absorption increases monsoon rainfall in the model, while solar dimming reduces it. I performed the simulations, the analyses and wrote the paper with important support and input from my coauthors.

7 Discussion and conclusions

Key processes causing climate change and variability were studied in this thesis. Climate models developed at the Max Planck Institute were employed and results were compared to measurements. Sensitivity of global mean temperature to greenhouse gas warming, when inferred from observations was shown to be uncertain to a large extent because of uncertain aerosol forcing and to some extent because of internal variability. In the instrumental global mean temperature record from 1850 to present, on the other hand, it was shown that isolating multidecadal internal variability is important to extract the aerosol effects. Thus, greenhouse gas warming, aerosol climate effects and internal variability are closely connected, even inseparable, research questions.

The thesis contains quantitative estimates for the effects of greenhouse gas warming, multidecadal internal variability and aerosols on the climate globally and regionally. The estimates show that all are important processes causing climate change and variability. Especially on decadal timescales, the strengths are comparable, illustrating the importance to take all into consideration in modeling when attempting decadal climate predictions. However, especially through climate sensitivity, also longer-term climate projections require an improved understanding of internal variability and aerosols, even if assuming that greenhouse gases in the atmosphere increase so much as to dominate the others in terms of radiative forcing and impact on temperatures. The estimates obtained and methods developed in this thesis may be applied in improving the estimates for greenhouse gas warming or other long term climate change through improved estimates of the other signals in climate data.

Of the more specific conclusions in the thesis, some highlights answering the questions presented in the Introduction are the following:

* Climate sensitivity deduced from observations still has relatively high uncertainty.

* A quasiperiodic oscillation in global mean temperature with a period of 50-80 years is observed in measurements and in the MPI models with its amplitude and typical frequency agreeing surprisingly well between the two. The oscillation is likely to explain part of the lack of rise of global mean temperature in the 1950s and 1960s, as well as from 1998 to present. However, external forcing has probably also been smaller during these times, which makes quantitative estimation of the oscillation's contribution to temperature records challenging. The amplitude can be estimated to lie between 0.05 K and 0.15 K, with large uncertainty remaining and with model results agreeing on a qualitative level with measurements.

* The power spectrum of global mean temperature is well approximated by a power law with exponent $\sim 0.7 - 0.8$ between multidecadal and El Nino frequencies in the earth system model when external forcing is included. The exponent is consistent with a best fit for the same frequency range of the spectrum of measured global mean temperature. Averaged spectra allow for better recognition of frequency ranges, where power laws are valid and are better suited for comparing climate variability in different datasets than single realisations.

* The model estimate for effects of northern hemisphere mid- and high-latitude volcanic eruptions during the last millennium with mean AOD north of 30°N exceeding 0.1 is -0.19 K for the largest hemispheric mean temperature anomaly and -0.095 K on average during the 21 months following the eruption with summer eruptions causing a larger integrated effect than winter eruptions. Precipitation decreases, but the signal is weak compared to internal variability. Especially for high-latitude eruptions, the time of the year

is important for photo-oxidation of sulfur dioxide, which was not explicitly modeled.

* Anthropogenic aerosols in India have a maximum in winter and natural aerosols in the summer. Secondary sulfate aerosols spread wider than primary carbonaceous aerosols.

* Absorption by aerosols seems to increase monsoon rainfall in India while solar dimming and the resulting weakened SST gradient in the monsoon season seem to reduce it. When including all aerosol processes in a simulation, assuming the observed cooling of the Northern Indian Ocean relative to the equator to be an aerosol effect, the total effect on monsoon rainfall is clearly negative.

The gained understanding about climate variability at different timescales and due to northern hemisphere volcanic eruptions could be helpful in decadal climate prediction, for example for Finland. The mechanisms of multidecadal internal variability could be studied further, especially for the Pacific and the deep ocean, which have been studied much less than the Atlantic part of the oscillation. It would also be interesting to study the nonlinearities involved: synchronization of the different regions by external forcing, coupling between variability at different timescales etc. It would also be important to study unforced climate variability at lower than multidecadal frequencies in more detail, to evaluate climate models in this aspect and evaluate the radiative forcing and equilibrium paradigm based on best knowledge of internal climate dynamics at each time. The effects of volcanic eruptions could be studied more carefully by looking at coupling with internal modes of variability and by resolving the eruption, which would account for the UV oxidation effects and climate effects immediately after the eruption. The aerosol-monsoon interplay in India could be studied further by analyzing the mechanisms more carefully or by using a regional climate model, which could also make the rainfall estimates more realistic. Secondary organic aerosols could be modeled to see if also they spread wider than primary carbonaceous aerosols like sulfate. The climate analysis could also be extended to China. Including more of the available aerosol observations in Asia in evaluating the model and in constraining it would be essential to estimate the effects on the radiative balance more accurately. This would facilitate separating temperature variability because of aerosols, internal dynamics and other factors from each other, not only for Asia but globally. Simulating greenhouse gas warming, aerosol effects and internal dynamics for the present climate all together with the best possible model setup would be desirable further research. Although challenging, if successful, it would help not only understanding past and present climate change and variability but also in developing the models for future predictions and projections.

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