

A Focus on Climate During the Past 100 Years

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Abstract The past 100 years are a key period for understanding climate variability and climate change as it marks the changeover from a climate system dominated by natural influences to one significantly dominated by anthropogenic activities. This volume is a compilation of contributions to a workshop dealing with different aspects of climate change, variability, and extremes during the past 100 years. The individual contributions cover a broad range of topics, from the re-evaluation of historical marine data to the effect of solar variability on the stratosphere. In this introductory chapter we provide an overview of the book in the context of recent research.

1 Introduction

The latest report of Working Group I of the International Panel on Climate Change (IPCC 2007) concludes that, very likely (probability >90%), human influence has contributed to a warming of the planet during the past 50 years and that it virtually certainly (>99%) will lead to an accelerated warming in the future. In order to understand, quantify and predict anthropogenic impacts on climate, the study of past climate variability is of utmost importance. The current understanding of past climate and its variability on interannual-to-decadal time scales is still fragmentary owing to the complexity of the coupling between atmosphere, ocean, and land surface and

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complex dynamics within the atmosphere itself. Open questions also concern the chemical and dynamical variability of the stratosphere, which arguably plays an important role in climate variability.

Probably the most important time period for assessing climate variability and understanding the underlying processes is the past 100 years. First, with respect to climate trends, this period marks the changeover of a climate system driven by natural forcings to a climate system in which anthropogenic forcings play a major role. Second, with respect to climate variability, the past 100 years are the closest precedent to the present and future. Understanding the processes governing interannual climate variability during the recent past may eventually contribute to improved seasonal forecasts. Third, with respect to extremes, the period is important as it comprises several volcanic eruptions, severe droughts and floods, heat waves and other climate extremes. Some of these extremes are expected to become more frequent in the future and therefore past extremes may serve as an analogue. Finally, the past 100 years are accessible, at least to some extent, for both direct observational analyses and modelling studies.

The aim of this book is to give a broad overview of climate variability and extremes during the past 100 years through a compilation of selected contributions to a workshop held in Gwatt, Switzerland in July 2006. Some of the contributions present new research results on a specific topic, while others have a review character and provide an overview with a broader scope. However, the book is by no means complete. This introductory chapter gives an overview of this volume in the context of recent research, highlighting some of the key findings and concepts.

The chapter is structured as follows. In Sect. 2 we use the example of Arctic warming to demonstrate why the focus on the past 100 years is important and how science might progress in the future. The remaining sections then follow the structure of the book, starting with the observational record (Sect. 3), climate trends and processes (Sect. 4), climate variability and extremes (Sect. 5) as well as chemical changes and the variability of the stratosphere (Sect. 6). Conclusions are drawn in Sect. 7.

2 The Focus on the Past 100 Years: An Example

An example that illustrates the importance of a focus on the past 100 years as well as the challenges in understanding the complex interactions in the climate system is Arctic warming (see also Alekseev et al. and Stroeve and Maslowski, this volume). During the past 100 years the Arctic experienced two pronounced warming periods (Fig. 1). Between 1915 and 1945, annual mean temperatures increased by about 1.8°C. This period was followed by a cooling and a more recent warming, which started around 1970 and is still ongoing. Stroeve and Maslowski (this volume) report accelerated melting of sea ice during the past decades, particularly since the late 1990s. In fact, the two most recent winters (2005/2006 and 2006/2007) were extremely warm and were accompanied by record low sea ice extent (see also Serreze et al. 2007). The change of Arctic climate is a major research focus (reflected in the Arctic Climate Impact Assessment and the International Polar Year 2007/2008) because of the sensitive nature of Arctic systems.

The recent warming appears to be related to global warming, which is expected to be amplified in the Arctic due to feedback effects (e.g., ice-albedo feedback or cloud feedback). Due to this “polar amplification”, most climate models produce an accelerated future Arctic warming of three to four times the global average (Holland and Bitz 2003). As a consequence, Arctic sea ice might change abruptly in the future (Holland et al. 2006). How sure are we about the mechanisms underlying the current and future Arctic warming in the light of the early 20th century warming? Global temperatures (land and sea) also increased in the first part of the 20th century (Fig. 1), but anthropogenic influences were much smaller at that time, and the Arctic warming could also just reflect an extreme realisation of internal variability in the climate system (Zhang et al. 2007). Some authors postulate a low-frequency oscillation (Polyakov et al. 2003) in the Arctic ocean–sea ice system which underlies any long-term trend, producing phases of accelerated warming. Bengtsson et al. (2004) suggest that the early 20th century warming was caused by increased transport of warm Atlantic water into the Arctic due to stronger westerly winds between Norway and Spitsbergen, reinforced by ocean–atmosphere–sea ice feedbacks in the Barents Sea (see also Fig. 2). For the recent warming, Stroeve and Maslowski (this volume) found that atmospheric circulation does not contribute considerably. Figure 1 shows that Arctic temperature is correlated with both global temperature and with the sea-level pressure (SLP) gradient between Norway and Spitsbergen (an approximation of the strength of westerly winds, bottom curve in Fig. 1). Given the similarity of response in Arctic temperatures during the early and late 20th century warming, the question remains: To what extent is Arctic temperature controlled by global warming, by the regional atmospheric circulation, or by lower frequency oceanic processes?

Our understanding of the climate mechanisms operating in the Arctic on different timescales is still limited. From the data side, it would be important to have better

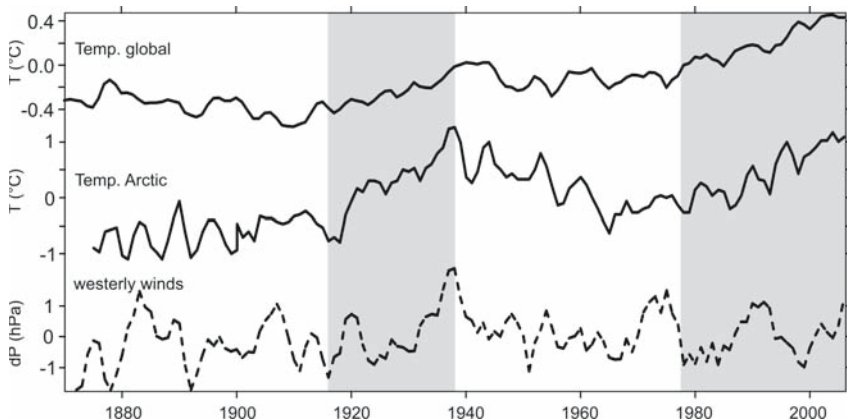


Fig. 1 Annual surface air temperature for the globe (top, HadCRUT2v, Jones and Moberg 2003) and north of 62°N (middle, Polyakov et al. 2003). The bottom curve shows the annual difference in sea-level pressure between Tromsø and Spitsbergen (HadSLP2, Allan and Ansell 2006). All series are anomalies with respect to 1961–1990, smoothed with a 3-year moving average. Grey shadings denote the two phases of pronounced Arctic warming.



Fig. 2 Sea ice cover in the Barents sea region plays an important role in Arctic climate variability (Bengtsson et al. 2004). True colour image (Terra/MODIS) of the Novaya Zemlya islands and the coast of Russia on 12 June 2001. The Barents sea (left) is ice free, whereas the Kara sea (right) is still ice covered (Jacques Descloitres, MODIS Land Rapid Response Team, NASA Visible Earth v1 ID 8126).

surface air temperature data as well as sea ice coverage and upper-air data, for both warming periods and the intermittent cooling phase. The data then need to be analysed with respect to seasonal to interannual variability of Arctic climate, including extremes, and with respect to the large-scale atmospheric circulation (Overland and Wang 2005). Land-surface feedback might play a particularly important role (Chapin et al. 2005). From the model side, one would wish to have models capable of reproducing both the early 20th century warming and the recent trend. However, there are still strong limitations in this respect (e.g., Wang et al. 2007).

Perhaps Arctic warming can only be understood by taking atmospheric chemistry into account. Pollutants emitted in the midlatitudes can be transported to the Arctic, where they lead to the phenomenon of “Arctic Haze”. Its effects on clouds and long-wave radiation in the Arctic are suggested to be significant contributors to the recent warming (e.g., Lubin and Vogelmann 2006; Garrett and Zhao 2006; Law and Stohl 2007).

Arctic weather and climate are also affected by the stratosphere and the stratospheric ozone layer. Through dynamical coupling of the stratosphere and troposphere,

stratospheric disturbances can propagate downward and affect weather and climate at the ground (e.g., Baldwin and Dunkerton 2001).

This list of research directions – data work, analyses of trends, variability, and extremes in observations and models, analyses of coupling processes and feedbacks, analyses of chemical climate and aerosols, analyses of stratospheric processes – not only holds for the problem of Arctic warming, but also for many other open questions in climate research. Therefore, we have structured the volume along these lines.

3 The Observational Climate Record

In this section we attempt to give an overview of the current efforts and remaining problems related to climate data over the past 100 years. The examples in this book cover the most important observation platforms and data types, though they are by no means exhaustive. Figure 3 shows the platforms and measurement systems used to study the Earth’s atmosphere and climate since the late 19th century. Many of the widely used global climate data sets for the Earth’s surface reach back to the second half of the 19th century (Fig. 3). These data sets, comprising sea-surface and surface-air temperature, pressure, precipitation, and some other variables, are the most important climate and weather data prior to the more recent period (1948–present), which has been reanalysed using data-assimilation techniques. There are, however, changes within these data sets that affect the quality of the analyses. National meteorological networks, the Brussels Maritime Conference of 1853 and, after

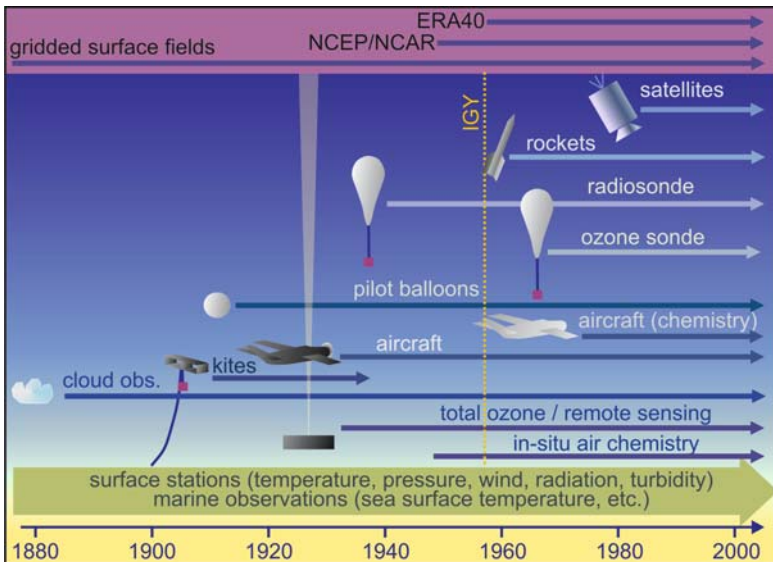


Fig. 3 Measurement platforms for 20th century climate data (adapted from Brönnimann et al. 2005).

1873, the International (now World) Meteorological Organization established standards for weather observations worldwide. The data quality has increased ever since, in some cases gradually, sometimes in steps – as has generally data coverage over both the terrestrial and marine domains.

The observational data sets are constantly being reanalysed and improved, but they could also be complemented and extended back in time. A large fraction of the meteorological data has still not been digitised. Concentrated efforts are needed (and are partly underway) to digitise and re-evaluate these data. However, changes in instrumentation, measurement practices and reporting, location, station environments, and lack of meta information inevitably lead to heterogeneous data sets. It is a challenging task to homogenize the data in order to assess climate trends reliably. The contributions by Woodruff et al. and Alekseev et al. (this volume) summarize these efforts for the examples of sea-surface temperature (SST) and sea ice. The situation is very similar for land (station) temperature and pressure data (e.g., Klein Tank et al. 2002; Allan and Ansell 2006; Ansell et al. 2006; Nicholls et al. 2006; Moberg et al. 2006; Della-Marta and Wanner 2006). Della-Marta and Beniston (this volume) report the additional efforts needed for homogenizing daily temperature data for the analysis of temperature extremes. With daily climate data it is necessary to homogenize the entire distribution function, not just the inhomogeneities in the mean.

Climate and weather data near the Earth's surface only provide a very limited view of the large-scale atmospheric circulation. Upper-air measurements started in the late 19th century (Fig. 3). During the first decades these measurements were in the early stages of development. Lüdecke (this volume) describes this pioneering work for the case of the Arctic, where kites, tethered and free-flying balloons, and airships were used as platforms. In many countries, operational upper-air networks using aircraft, kites, and pilot balloons gradually established during the 1900s–1920s, while radiosonde networks started only in the late 1930s and 1940s (Fig. 3). An important event in the history of climate observations was the International Geophysical Year (IGY) 1957/58 (Fig. 3), which not only led to the establishment of meteorological stations in Antarctica, a ground-based global total ozone network, and to measurements of CO₂ in the atmosphere, but also to an improved, global radiosonde network. Still, as these measurements were predominantly obtained for operational weather forecast, achieving trend quality remains a major problem (Free and Seidel 2005; Sherwood et al. 2005). In the contribution by Sterin et al. (this volume) issues related to the quality of upper-air observations and differences between data sets are discussed.

The availability of global radiosonde data also marks the beginning of current reanalysis data sets, which have become invaluable for atmospheric research. Using numerical weather prediction models combined with a filtering and interpolation scheme, radiosonde data, observations from the Earth's surface, from ships, satellites, and aircraft are assimilated. The most widely used global reanalysis data sets are NCEP/NCAR reanalysis (Kistler et al. 2001) and ERA40 (Uppala et al. 2005), reaching back to 1948 and 1957, respectively (Fig. 3). Uppala et al. (this volume) assess the value of reanalysis data products for climate analysis. While caution is still necessary when using the data for trend analysis (see Sterin et al. this volume),

they are invaluable for the analysis of global atmospheric circulation variability and offer opportunities to reconcile information from independent platforms simultaneously. Efforts are currently underway to extend three-dimensional atmospheric data sets back to the 19th century based on surface data and historical upper-air data using statistical reconstruction techniques and data assimilation (Brönnimann et al. 2005; Compo et al. 2006).

Information from a reanalysis can also help in the process of homogenizing radiosonde temperature records. Instrument-dependent radiation errors as well as breakpoints in groups of stations have been identified and corrected by these methods (Andrae et al. 2004; Haimberger 2007; see also Sterin et al., this volume).

Another important step in the history of climate observations was the advent of space-borne Earth observation in the 1970s (Fig. 3). Satellites potentially provide a near-global coverage of numerous climate variables. For instance, the contributions by Alekseev et al., Mears, Norris, and Stroeve and Maslowski (this volume) address products based on satellite observations such as sea ice coverage (see also Fig. 2), cloud cover, and the vertical temperature structure. At least as important as for surface data, quality remains a fundamental issue for satellite data. In particular, the overlap between different sensors is often too short to obtain reliable transfer functions and the necessary corrections have been a matter of debate (Mears and Wentz 2005; Mears, this volume).

Not only was the physical state of the atmosphere measured during the past 100 years, but also its chemical composition. Spectral measurements in the UV wavelength range that provide information on the amount of ozone present locally in the Earth's atmosphere (i.e., total ozone) were performed since the 1920s (see Brönnimann et al., this volume). Starting around the 1950s to 1970s, in situ chemical measurements were performed at the ground, in the troposphere (e.g., using aircraft) and in the stratosphere (using balloons, see Staehelin et al., this volume). Since the 1970s, space-borne sensors have been delivering a wide range of chemical information, including ozone concentrations. Moreover, satellites measure aerosol properties and land-surface information that is extremely important for climate studies (e.g., MODIS on NASA's Terra Aqua satellites, see Fig. 2). Monitoring the atmosphere from space not only allows the climate system to be studied from a chemical and physical perspective, but also allows external (e.g., volcanic, solar) climate forcings to be quantified. Examples are found in the contributions by Jackman and Fleming and by Rozanov (this volume).

Large efforts have been devoted to obtaining data sets for the past 100 years that are suitable for climate applications. Reprocessing and homogenizing the available data in order to obtain a sufficient level of quality to enable the analysis of climate trends is an important, ongoing task. At the same time, constant efforts are also needed to ensure the quality of the present and future monitoring systems. In particular, space-borne observations offer new possibilities, but also new challenges with respect to long-term data quality.

While quality is one problem, data quantity is another, particularly in the early decades. Much could still be learned from analysing the past 100 years and concentrated data rescue efforts are needed to extend and supplement the early part of the record.

4 Climate Trends: Forcings and Processes

During the past 100 years global climate has experienced major changes. Several contributions in this volume deal with trends in the climate system and the underlying physical processes. In this section we provide a background for these chapters by briefly discussing climate forcings and temperature trends.

As shown in Fig. 1, two phases of distinct global (and Arctic) warming can be addressed. However, these two warming periods had a different spatial signature, as is shown in Fig. 4 for the boreal winter (January–March) period. The 1910–1945

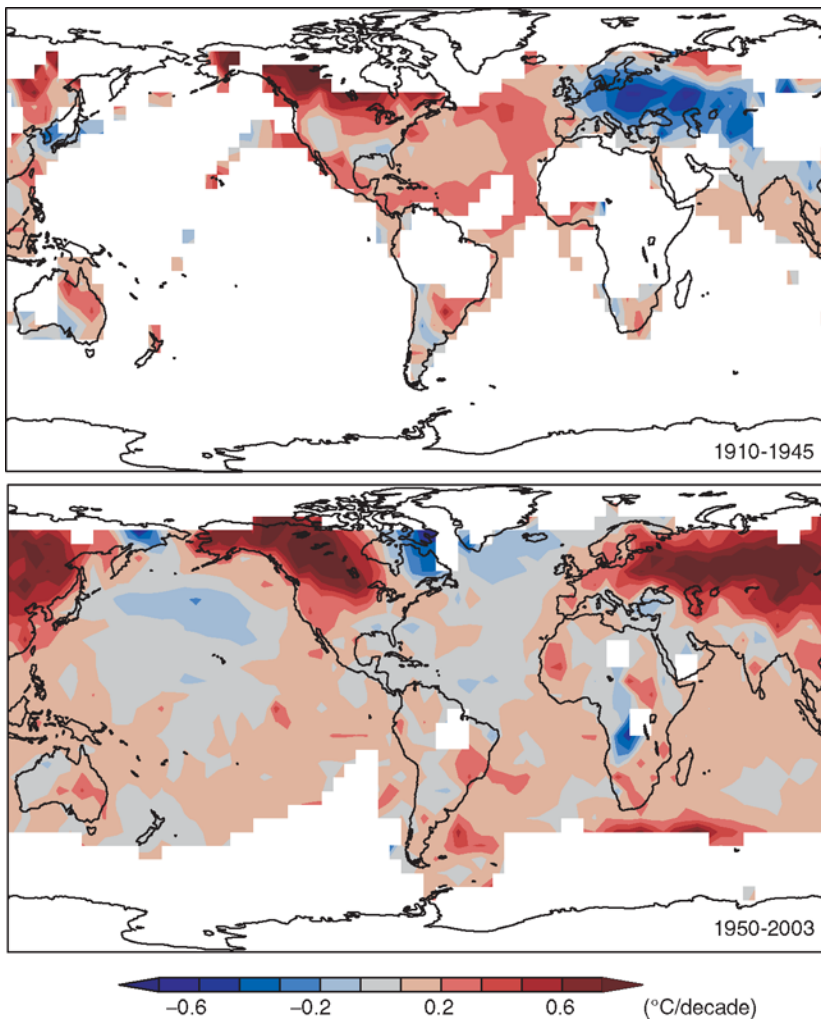


Fig. 4 Linear trend of boreal winter (January–March) surface air temperature for the period 1910–1945 (top) and 1950–2003 (bottom) using HadCRUT2v data (Jones and Moberg 2003). At least 20 winter mean values must be available in order to calculate a trend.

trend was most pronounced over the Atlantic and North America, while Europe experienced a winter cooling. In contrast, the 1950–2003 period exhibited a strong winter warming of the northern hemispheric land masses (see also Jones et al. 1999; Trenberth et al. 2007). What may have caused this difference in climatic response?

Multidecadal climate trends can be understood as the product of external forces (due to both natural and anthropogenic factors) and internal variability of the climate system. The development of the external factors (expressed as radiative forcing relative to 1870–1880) is shown in Fig. 5. Despite remaining uncertainties, which mainly concern the solar and aerosol effects, it is clear that the past 100 years mark the transition of a climate system where changes are dominated by natural forcings to one where changes are dominated by anthropogenic forcings. Was the 1910–1945 trend a result of “natural variability” and the 1950–2003 trend an “anthropogenic” warming?

Detection and attribution studies use climate models to predict the effect of the forcings and to estimate internal variability and then use these results for statistical analyses of the observed climate record (Allen et al. 2006; Hegerl et al. 2006; 2007). These studies often find that the early 20th century warming was caused by both increasing solar irradiance and large natural variability, while anthropogenic effects could have also contributed. The “large natural variability” could be related to the so-called Atlantic Multidecadal Oscillation (AMO; Delworth and Mann 2000) – the basin-wide average of North Atlantic SSTs, which is believed to be related to the strength of the thermohaline oceanic circulation (Knight et al. 2005; Zhang et al. 2007).

The global temperature trend during the past 30 years is dominated by the anthropogenic climate forcing (e.g., Allen et al. 2006). The anthropogenic influence is also detectable in SLP, ocean heat content, and tropopause altitude (e.g., Santer et al. 2003). The rate of change observed during the past 30 years was particularly large and affected not only surface air temperatures but also tropospheric and stratospheric temperatures (Mears, Sterin et al., this volume; for an overview on

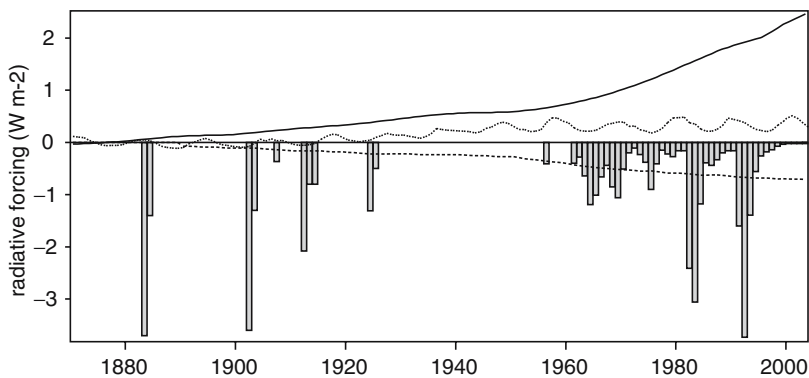


Fig. 5 Radiative forcing of climate since 1870. Solid: Greenhouse gases, dotted: solar, dashed: tropospheric aerosols (updated from Crowley (2000) and Lean (2004)). All series are shown as differences with respect to 1870–1880.

stratospheric temperature trends see Ramaswamy et al. 2001, 2006), Arctic sea ice coverage (Stroeve and Maslowski, this volume) and thickness (Alekseev et al., this volume) and many other variables. Menzel et al. (this volume) demonstrate that particularly rapid changes have also occurred in several phenological time series. Trends in cloud cover and solar radiation are discussed in Norris (this volume) and Wild (this volume), respectively.

In terms of radiative forcing, natural factors dominate only immediately after strong tropical volcanic eruptions (Fig. 5). At least five major tropical eruptions occurred during the past 125 years, which allows volcanic effects on climate to be studied and better understood. Figure 6 (top) shows surface air temperature and SLP anomalies averaged over the first post-eruption winters (January–March). Volcanic eruptions illustrate that the climate response to a radiative forcing is more complex than a uniform temperature change. Rather, there are pronounced seasonal and regional patterns, and short and long-term effects need to be distinguished. While over the oceans, volcanic eruptions lead to cooling related to decreased shortwave radiation (with long-term effects, see Church et al. 2005), they cause a winter warming of the northern extratropical land masses (see Robock 2000; Fischer et al. 2007). This is an indirect dynamical effect that might proceed via the stratosphere (see further discussion in Sect. 6). Depending on the time period, this pattern might contribute to winter temperature trends (e.g., from 1950 to 2003; Fig. 4 bottom).

Among the anthropogenic forcings, greenhouse gases are the most important and arguably the best understood. However, other influences such as tropospheric aerosols and land-use change also play a role. These influences are addressed in the contributions by Lohmann, Wild, and Seneviratne and Stöckli (this volume). Figure 7 shows schematically the fluxes of energy, matter, and momentum in the troposphere and how they are affected by anthropogenic activities. Greenhouse gases primarily affect the long-wave radiation balance, while the effect of aerosols on the energy and water balance is more complex. Lohmann (this volume) provides an excellent overview.

The aerosol effect is often addressed in terms of direct and indirect effects. A consequence of the direct effect is a reduction of total solar radiation reaching the Earth's surface (Fig. 7), which is known as “global dimming” (Ohmura and Lang 1989). Wild (this volume) shows that for the last 20 years, this trend has reversed over the industrialised regions (in fact, over much of the globe) in line with a decrease in sulphur emissions (Wild et al. 2005). In order to disentangle the effects of increasing long-wave radiation (due to greenhouse gases) and changing short-wave radiation (due to aerosols, affecting only daytime temperature), the diurnal temperature range is often analysed. Wild (this volume) shows that its change agrees with a trend reversal from “global dimming” to “global brightening”.

The indirect aerosol effect (see Lohmann, this volume) comprises the effects on the cloud optical and microphysical properties as well as on the cloud amount and lifetime (see Fig. 7). Its magnitude is one of the largest remaining uncertainties in the assessment of anthropogenic climate influences (Forster et al 2007). Addressing trends in cloud cover (which aerosols and other factors also influence) is much more difficult than

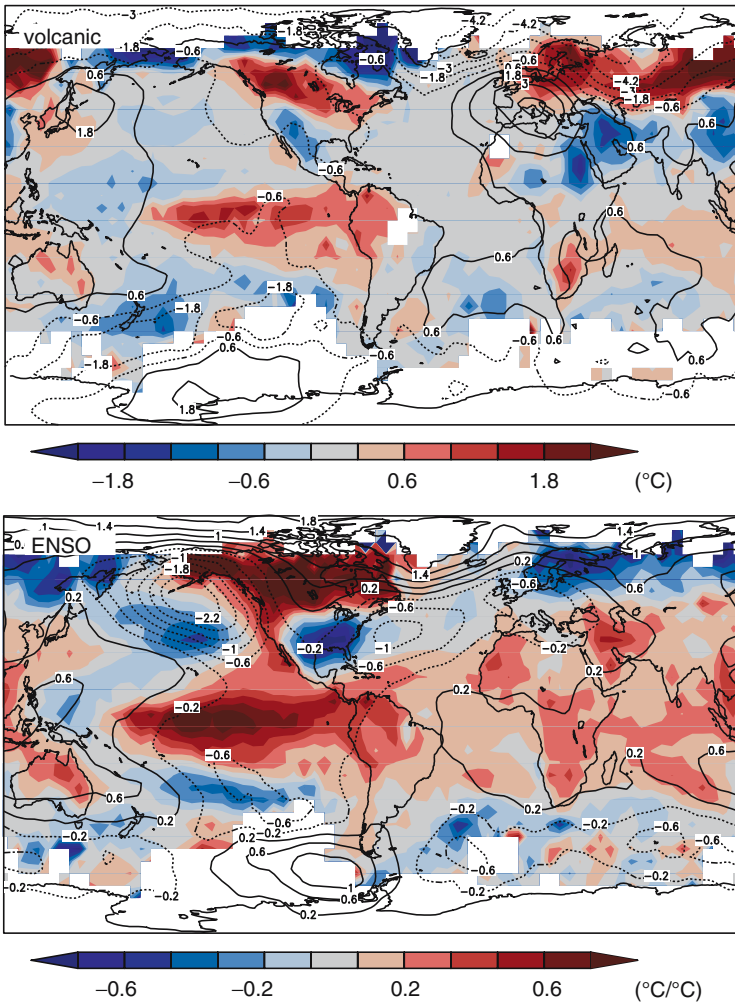


Fig. 6 Effect of tropical volcanic eruptions (top) and ENSO (bottom) on boreal winter (January–March) surface air temperature and SLP since 1880. Temperature and SLP data were detrended. The top panel shows a composite of the first winters after five tropical eruptions (Krakatoa, Santa Maria, Mt. Agung, El Chichón, Pinatubo). The bottom panel shows regression coefficients using a NINO3.4 index (September–February average) after removing two winters after each major volcanic eruption (same as above plus Katmai, see Brönnimann 2007).

addressing temperature trends. Norris (this volume) reports that upper-level cloud cover has decreased over the past decades, while the data show large discrepancies and inconsistencies when it comes to low clouds (see also Evan et al. 2007).

Another important player in the climate system is the land surface. It affects climate through the storage of water and CO₂ in soils and vegetation, through controlling

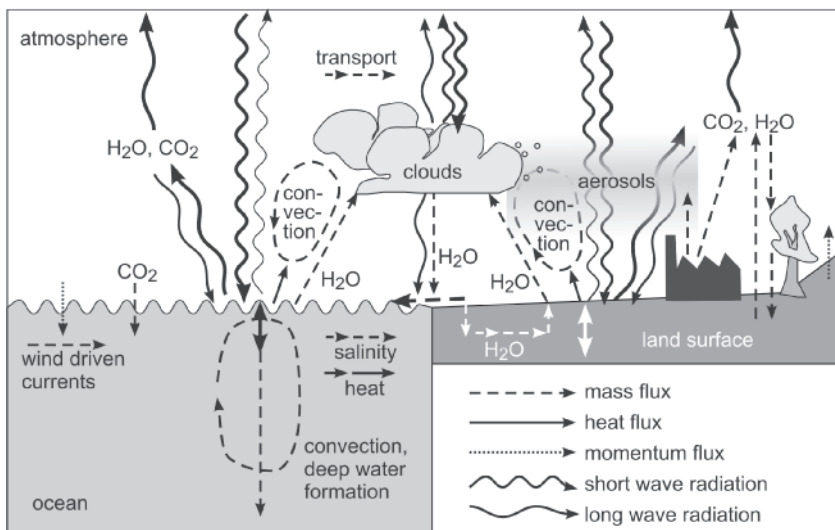


Fig. 7 Schematic depiction of the fluxes of matter, energy, and momentum in the climate system.

their exchange with the atmosphere and through modifying the energy balance (Fig. 7). Land-surface processes are not static but themselves depend on climate (e.g., albedo). Moreover, human induced land-use changes can affect climate via land-atmosphere interactions. Seneviratne and Stöckli (this volume) give an overview of the processes relevant for land-atmosphere interactions and their implications for climate variability.

Marked climate trends have been observed in the global climate system during the past 100 years, most importantly during the past 30 years, when the anthropogenic influence was strongest (IPCC 2007). These trends are increasingly well documented; examples are found throughout this volume. While some of the processes are relatively well understood today, this is not the case for others, in particular when coupling processes and feedbacks come into play. Focusing on the past 100 years, which comprise periods with strong and weak trends in anthropogenic forcing, might help to narrow down some of the remaining uncertainties.

5 Climate Variability and Extremes

The processes depicted in Fig. 7 show the direct, “static” effect of climate forcings on the budgets of energy, matter, and momentum. As mentioned for the case of volcanic eruptions, there are important indirect, “dynamic” effects as these processes also change the large-scale atmospheric circulation and its variability. For instance, they might affect the strength of the Hadley cell (see Diaz and Bradley 2004 for an overview) or the extratropical circulation. In order to properly assess future climate scenarios it is therefore important to study the interannual climate variability that

was observed or modelled during the past 100 years. Moreover, a better understanding of interannual climate variability helps to improve seasonal forecasts. The interest is particularly strong for extreme climate events that have large economical impacts such as large-scale droughts and heat waves. In fact, several contributions in this volume address heat waves and droughts.

One main line of research on interannual climate variability and extremes concerns the concept of variability modes and their impact on climate, predictability, and change in a future climate. Arguably the most important mode is El Niño/Southern Oscillation (ENSO). El Niño is an episodic warming of the eastern and central tropical Pacific, associated with an eastward shift of convective activity from the western tropical Pacific (Philander 1989; Diaz and Markgraf 2000; McPhaden et al. 2006). Via changes in the Hadley circulation and interaction with the extratropical circulation, ENSO affects temperature and SLP globally as is shown for the boreal winter period in Fig. 6 (bottom, regression coefficients using the September–February average of the NINO3.4 index, see Brönnimann 2007). El Niño often brings drought to different parts of the world (e.g., Diaz et al. 2001). Apart from affecting the Indian monsoon, it also influences monsoonal rainfall over southern West Africa (Paeth, this volume). Because ENSO is to some extent predictable (e.g., Chen et al. 2004), studying the relationship between the tropical Pacific and climate in more remote regions is an important area of research.

The dominant mode of interannual variability in the North Atlantic-European sector is the North Atlantic Oscillation (NAO, see Wanner et al. 2001; Hurrell et al. 2003). It is defined as the SLP difference between the Azores high and the Icelandic low (the two dominant centres of action over the Atlantic) and affects climate in Europe, North Africa, and eastern North America. The NAO is an important concept for assessing impacts of climate variability. For instance, Menzel et al. (this volume) show that the phase of the NAO affects spring leaf unfolding in Germany. Large efforts are currently devoted towards a seasonal forecasting of the NAO, even though its predictability seems to be limited.

Climate extremes are often related to extremes in the above-mentioned variability modes, although in general, different factors act together to produce extremes. This is shown in Fig. 8 for the example of droughts. The key factors identified here are oceanic forcing, aerosol effects, and land–atmosphere interactions. As a consequence, global change can affect the frequency of extremes in various ways.

Oceanic forcing of the atmosphere can alter the large-scale moisture flux, including the monsoon systems and other circulation systems that normally transport humidity to the drought-prone areas. Several major droughts were observed during the past 100 years and were analysed in relation to oceanic forcing. The importance of large-scale SST anomalies in forcing atmospheric circulation patterns that are conducive to drought has been demonstrated for the Great Plains (Schubert et al. 2004), the Sahel region (Giannini et al. 2003; see also Paeth, this volume), and central Europe (Sutton and Hodson 2005; Black and Sutton 2007; see also Hoerling and Kumar 2003). For instance, large-scale SST anomalies related to a positive phase of the AMO cause dry conditions across the USA, while a cold tropical Pacific (e.g., La Niña) often leads to droughts in the south-west of the USA and the Great Plains (McCabe et al. 2004). Both effects acted together in the 1930s and set

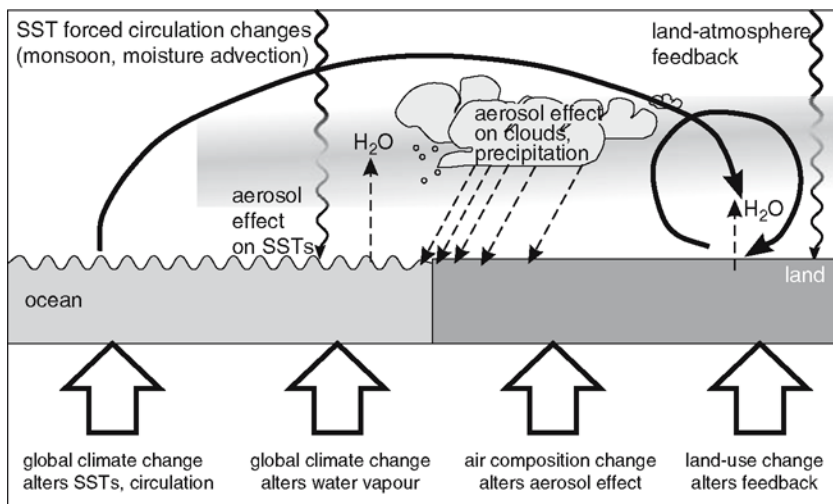


Fig. 8 Processes involved in the development of droughts and pathways through which they can be affected by climate change.

the stage for “Dust Bowl” droughts in the USA (Schubert et al. 2004). However, other processes such as land–atmosphere feedbacks also played an important role (Schubert et al. 2004; see also Seneviratne and Stöckli, this volume). The “Dust Bowl” is not the only region where land–atmosphere coupling is important. In a future climate, this might also be the case for central Europe (Seneviratne et al. 2006).

Paeth (this volume) gives an overview of drought conditions in the Sahel region. While large-scale SST anomalies and changes in atmospheric circulation are an immediate cause, the long-term trends can also be explained by other causes, including soil properties, vegetation cover, and aerosols (see also Rotstayn and Lohmann 2002). Lohmann (this volume) discusses the effect of aerosols on precipitation and points to droughts in Southern China that may have been linked to a decreased land–sea temperature gradient (due to the presence of aerosols) and a weak monsoon circulation (see also Lau et al. 2006). However, the suggested mechanisms differ between the studies and the results are not yet conclusive, as current climate models do not account for all relevant microphysical properties.

Droughts are also experienced in Europe. The Mediterranean region is particularly vulnerable (e.g., Xoplaki et al. 2004), but other regions are also affected by droughts. The most prominent recent example is the heat wave in central Europe in 2003 (Schär et al. 2004; Beniston 2004). Have droughts and heat waves in Europe become more frequent? Della-Marta and Beniston (this volume) analyse changes in the occurrence of summer heat waves in Europe during the past 126 years. Apart from important data obstacles they had to overcome, the authors report that heat waves have increased both in number and in persistence.

The topic of climate extremes became apparent to the participants of the meeting from which this book results, as the event took place during a heat wave. In many parts

of Switzerland, July 2006 was the hottest July on record (Fig. 9 top left). Fall 2006, winter 2006/2007 and spring 2007 were equally extreme (shown for the series from Bern, Switzerland, in Fig. 9 bottom, see also Luterbacher et al. 2007). Had the meeting taken place a few months earlier, the participants would have witnessed another extreme: In August 2005 the conference venue was flooded (Fig. 9 top right).

Is this evidence that climate has become more variable? Several authors have suggested that increasing temperature variability contributes to the increased frequency in heat waves (Schär et al. 2004). For the occurrence of extremes, changes in variability are as important as changes in the mean (Katz and Brown 1992).

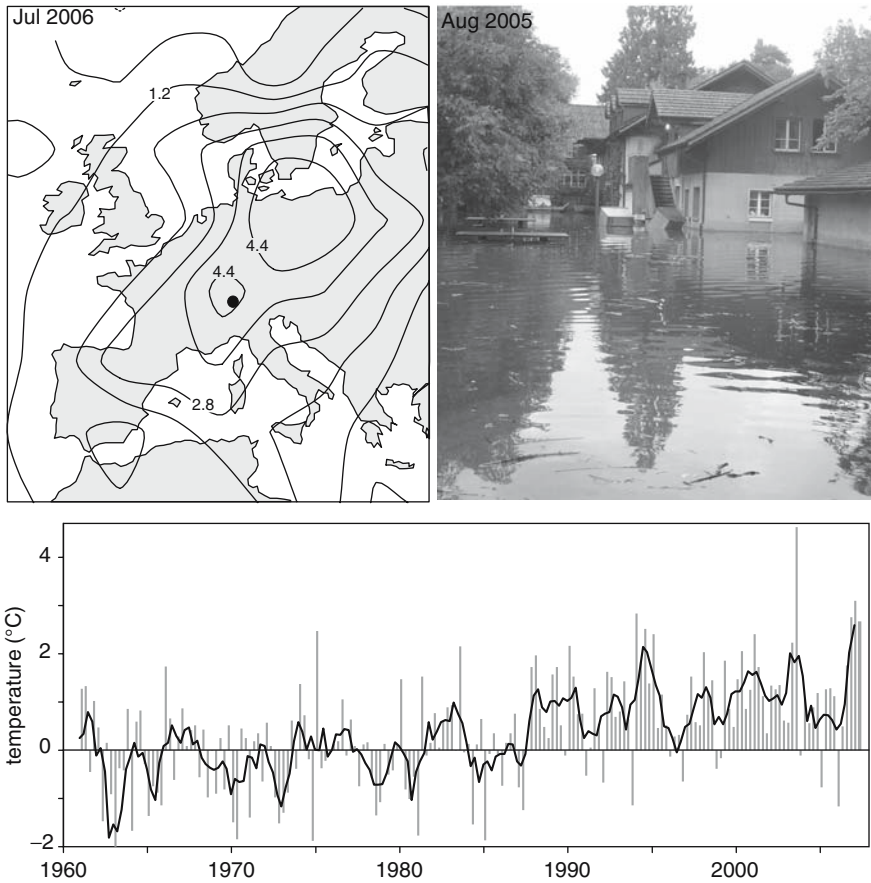


Fig. 9 Surface air temperature anomalies in July 2006 (top left, in degree Celsius with respect to the 1961–1990 average), Jones and Moberg 2003. The conference site in Gwatt, Switzerland (dot in left panel) in August 2005 (top right) following an extended period of much above normal precipitation. Lower panel displays seasonal temperature anomalies (with respect to the 1961–1990 average) in Bern, Switzerland from 1961 to 2007 (winter 1963 is off the scale), data provided by MeteoSwiss. The solid line shows a running average over four seasons.

Scherrer et al. (this volume) analysed climate data as well as climate scenario runs for the 21st century with both respects, focussing on central European temperature. For the summer season, they found an increase in the mean during the past 40 years and in the variability. Analyses based on daily summer temperature from Della-Marta and Beniston (this volume) show that summer temperature variability has increased significantly over the last 126 years in central western Europe. Relatively small changes in the variability of daily summer temperature (+11%) have contributed approximately 40% to the increase in extremely hot days. For the 21st century, Scherrer et al. (this volume) report large relative increases in mean temperature for all seasons with maxima in summer, accompanied by an increase in variability. In Central and Eastern Europe, this appears linked with a mean decrease of soil moisture and a consequent strengthening of land–atmosphere coupling as concentrations in greenhouse gas increase (Seneviratne et al. 2006; see also Seneviratne and Stöckli, this volume), which could also explain part of the recent trend during the 20th century.

The contributions in this volume show that progress has been made in addressing and documenting changes in climate extremes. Such progress is important, as extremes are rare events that occur infrequently enough for statistical analyses, and are difficult to model numerically. Concerning the underlying processes, the contributions show that several factors act together and therefore must be considered together, similar as outlined in Sect. 2 for the problem of Arctic warming. For droughts, for instance, the large-scale oceanic forcing of atmospheric circulation must be considered as well as land-surface processes and aerosols. A lot still has to be learned in order to better assess extremes and their possible change in a future climate and in order to obtain better forecasts. In this effort, the past 100 years have been and still are an important learning ground.

6 Chemical Changes and the Variability of the Stratosphere

In the past 100 years, climate has undergone changes not just near the Earth's surface but also in the middle and upper atmosphere. Also, changes have been observed in the chemical properties of the atmosphere, some of which have affected the climate near the ground. The link between chemical and physical aspects of climate is a comparably new and rapidly evolving topic of climate research that is covered only to a limited extent by the contributions in this book. A key role is played by ozone and by stratospheric processes – the focus of this section.

Trace gases naturally occur in the atmosphere due to emissions by living organisms, geogenic emissions, lightning, or forest fires. Humans have started to influence the composition of the atmosphere substantially and on a global scale at least since the advent of the industrialisation (see Staehelin and Schnadt Poberaj, this volume). In addition to the greenhouse gases CO₂ and methane (CH₄), reactive gases are being emitted. Combustion processes produce nitrogen oxides (NO_x), carbon monoxide (CO) and, depending on the fuel, sulphur dioxide and a range of organic compounds. Industrial processes and households have emitted ozone depleting substances, and man-induced biomass burning produces large amounts of aerosols. These constituents have profound effects on both air quality and climate.

A particularly important tropospheric constituent is ozone. It is an irritant to humans, a greenhouse gas, and one of the main determinants of the oxidizing capacity of the troposphere (and hence the lifetime of other greenhouse gases). It is formed in the stratosphere via oxygen chemistry in the presence of UV radiation. In the troposphere, it can be formed from the oxidation of CO , CH_4 , or higher hydrocarbons in the presence of NO_x . How much of the present tropospheric ozone burden is a consequence of the anthropogenic influence and how much is natural (e.g., influx from the stratosphere)? Staehelin and Schnadt Poberaj (this volume) assess the changes in tropospheric ozone during the past decades. They find that tropospheric ozone levels over Europe have doubled since World War II due to anthropogenic emissions.

Ozone plays a particularly important role in the stratosphere, where it forms the ozone layer that shields life on Earth from harmful UV radiation. Ozone depleting substances have caused a considerable depletion of stratospheric ozone since the 1970s and have led to the phenomenon of the Antarctic ozone hole (Solomon 1999; Staehelin et al. 2001; WMO 2007; see also Brasseur and Brönnimann et al., this volume). Figure 10 shows the trend of the ozone column in February between 1979

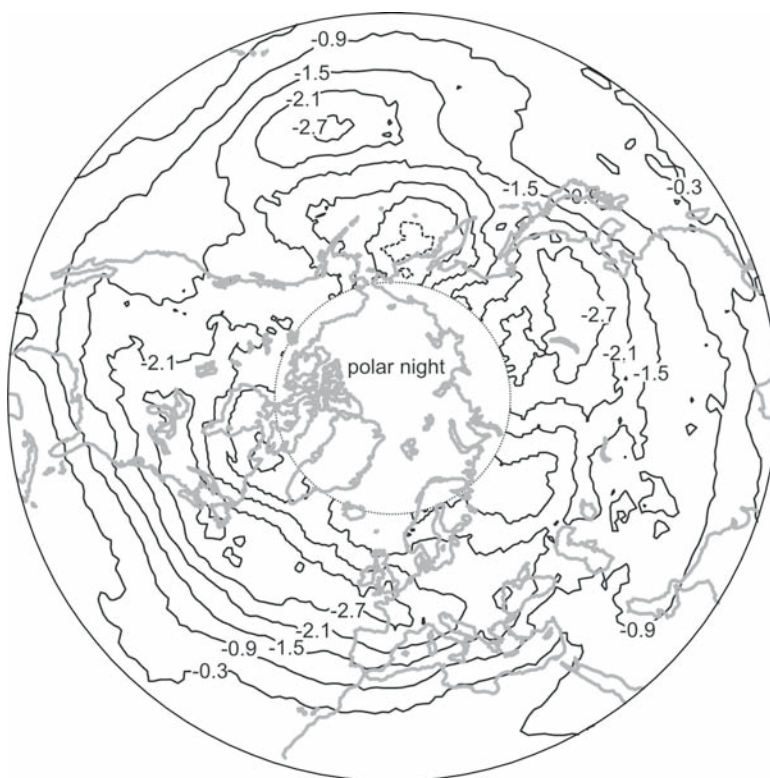


Fig. 10 Linear trend of the ozone column (DU/Year) in February from 1979 to 2001 based on Total Ozone Mapping Spectrometer data.

and 2000. Apart from the ozone depletion signal, the trend has a clear spatial pattern which can not be explained by chemical depletion but rather points to dynamical processes. Hence, understanding the ozone layer forces us to combine chemical and physical perspectives and to use different methods and concepts in order to understand the atmosphere. Using stratospheric ozone as an example, Brasseur (this volume) shows that scientific understanding can only develop due to the interplay of observations, models, and laboratory measurements. All three methods are necessary to confirm, contradict, and constrain each other – until new observations (such as the discovery of the ozone hole) reveal an inadequacy in our understanding, leading to a new learning cycle.

How good are the chemical and dynamical stratospheric processes represented in current models? Dameris and Deckert (this volume) integrated a model over a relatively long period of 40 years. They show that their model is able to capture the main features of temporal variability of stratospheric ozone on different time scales. However, in addition to ozone depleting substances, other important forcing factors (greenhouse gases, SSTs and sea ice cover, Quasi-Biennial Oscillation, solar variability, volcanic aerosols) also need to be included. Currently, large efforts are being devoted to a process-oriented validation of chemistry climate models (Eyring et al. 2006). This allows the models to be used for simulating future changes in the stratospheric ozone layer with sufficient confidence.

Apart from affecting UV radiation at the Earth's surface, changes in stratospheric ozone might also change climate at the ground (Thompson and Solomon 2002; Gillett and Thompson 2003). Of particular interest in this context are the stratospheric polar vortices which form in winter. On the one hand, the polar vortices both affect and are affected by ozone chemistry and are a location of chemical-dynamical interaction. On the other hand, the polar vortices also take part in the dynamical coupling between the stratosphere and the troposphere (Holton et al. 1995). Langematz and Kunze (this volume) discuss trends in the intensity and persistence of the vortices at both poles during the past decades. They find a high variability on multiannual timescales, which makes trend detection very difficult. For instance, a series of very cold winters in the Arctic stratosphere during the 1990s was then followed by a trend to a more disturbed stratosphere in recent years. Longer time series are required to assess trends in the dynamics of the polar vortices.

An important contribution to the variability of the northern polar vortex is that of “Major Midwinter Warmings” (see Labitzke et al., this volume). In these events, the vortex collapses completely in conjunction with an abrupt warming of the middle stratosphere by several tens of degrees within a few days. Such events are mostly triggered by anomalies in planetary wave activity propagating upward from the troposphere (see Limpasuvan et al. 2004). Once the vortex is significantly weakened in the middle stratosphere, the signal often propagates downward (Baldwin and Dunkerton 2001). Figure 11 shows the time–altitude evolution of the daily Northern Annular Mode index (measuring the strength of the polar vortex) in 2006. Day 0 marks the onset of a major warming in the middle stratosphere (30 km), which propagates downward within a week or two and affects the troposphere repeatedly until the end of February.

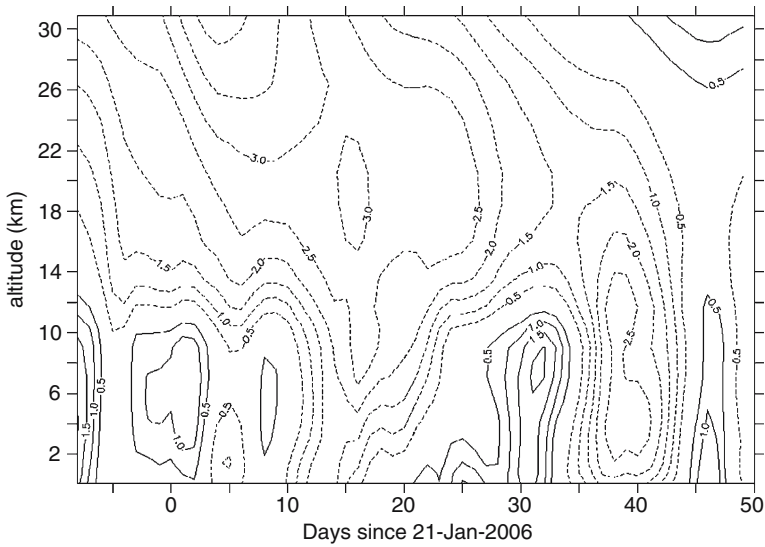


Fig. 11 Development of the standardized daily index of the Northern Annular Mode in January and February 2006 (data were provided by Mark Baldwin). Day 0 is defined as the first day when the index exceeds three standard deviations at 10hPa (ca, 30 km).

Downward propagation offers a mechanism by which forcings that strongly act on the stratosphere, such as solar or volcanic effects, can have an influence on climate at the ground. As schematically depicted in Fig. 12, strong tropical volcanic eruptions not only cool the Earth's surface by blocking short-wave radiation, but also heat up the stratosphere considerably. During the winter season this can lead to a differential heating between the sunlit areas and the polar night, which in turn may strengthen the polar vortex and, through downward propagation, weather and climate at the ground (see also Stenchikov et al. 2004). The winter warming following tropical eruptions (Fig. 5 top) could be related to this mechanism.

A combination of direct thermal effects (White et al. 1997; Meehl et al. 2002) and indirect dynamical effects via the stratosphere could also be responsible for solar effects on climate (see Calisesi et al. 2007 for a recent compilation of papers on Sun-climate links). Rozanov et al. (this volume) give an overview of the effects via the stratosphere (see also Shindell et al. 1999; Kodera and Kuroda 2002). The variability of solar radiation over an 11-year sunspot cycle is much larger in the UV wavelength range than total solar irradiance. Correspondingly, clear solar effects appear in upper-stratospheric ozone and temperature, which might lead to changes in stratospheric circulation and, through downward propagation, climate at the Earth's surface (see Rozanov et al., this volume). Changes in the Sun might also affect climate on Earth via energetic particles. Jackman and Fleming (this volume) provide an overview of the effect of solar proton events on the mesosphere and stratosphere. However, much about the solar influence still remains to be learned. For instance, Labitzke et al. (this volume) show that solar variability does affect the

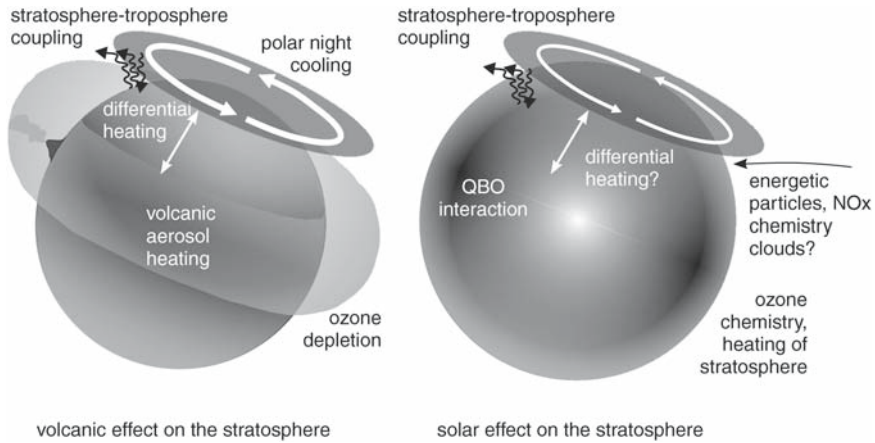


Fig. 12 Schematic depiction of the effects of volcanic eruptions and solar irradiance changes on the stratosphere.

northern polar vortex, but the effect depends on the phase of the Quasi-Biennial Oscillation in the tropical stratosphere – a demonstrated effect that is not well understood.

Chemical climate variability and stratospheric processes are today considered important areas of climate research. This is reflected, for instance, in the IGAC (International Global Atmospheric Chemistry) project of the International Geosphere-Biosphere Programme and the SPARC (Stratospheric Processes and their Role in Climate) project of the World Climate Research Programme. Recent research results have further demonstrated the importance of stratospheric processes (Baldwin et al. 2007). In particular, the concept of downward propagation provides hope that some of the long-standing issues in climate research such as the effect of solar variability can finally be addressed.

7 Conclusions

While major progress has been made in the past few years towards quantifying and understanding climate variability, trends, and extremes, there are still many open questions. Answering these questions will be important for our understanding of the climate system, in particular for assessing future climate change and for improving seasonal forecasts. A key period in this context are the past 100 years, which mark the changeover of a climate system dominated by natural forcings to a climate system dominated by anthropogenic influences.

The contributions in this volume provide a broad overview of the current state of the research on climate variability and extremes during the past 100 years. In this introductory chapter we embed the contributions in a scientific context and summarise

some important findings and key mechanisms that are currently discussed. At the same time, we show that our knowledge is still incomplete. We argue that the complexity of many current problems in climate research calls for a three-dimensional, physical-chemical view of the climate system.

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