Reconstruction of global upper-level circulation 1880-1957 for analyzing decadal climate variability

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Three little words achingly familiar on a Western farmer’s tongue, rule life in the dust bowl of the continent – if it rains.

Robert Geiger, April 15, 1935, in Surviving the Dust Bowl

The cover displays the reconstructed geopotential height field at 700 hPa in January 1881 together with the wind at 3000 meter.
Abstract

The climate system exhibits strong interannual to interdecadal variability on a regional to continental scale among others becoming apparent in wet spells and droughts. Particularly droughts have widespread consequences concerning agricultural production, human health and transport and energy sector.

This thesis focuses on droughts in Europe and the US, both regions with a high population density and a high agricultural productivity. Therefore both regions are highly vulnerable to drought conditions. Europe and the US experienced several severe drought events. Well known is the “Dust Bowl” drought in the US Great Plains in the 1930s which had devastating consequences. In the 1940s and 1950s anomalous low precipitation was observed in Europe. For the year 1947 a complete crop loss for Switzerland is recorded.

In Europe and the US, drought occurrence is linked to sea surface temperature anomalies. Hence, besides other factors, oceanic forcing seems to be important for the development or intensification of droughts. The oceanic signal is transferred to the continents via the upper-level circulation, visible in quasi-stationary circulation anomalies. It is expected that the upper-level circulation plays a crucial role in causing and pertaining drought events. With a better understanding of the upper-level circulation, future drought events could be predictable since sea surface temperatures are predictable to some extent.

So far drought events in the US and in Europe have been analyzed from a ground based perspective since for the first half of the 20th century no upper-air data have been available or are not in a suitable format. However, upper-air measurements started in the early 20th century. Records from the first half of the 20th century are still available on paper in various archives. Through an effort of the scientific community many records were digitized and became available in the last few years. Since historical upper-air measurements are unequally distributed in space and have longer or shorter gaps, they can not directly be used to resolve the global circulation patterns. Therefore upper-air measurements before 1957 together with additional ground measurements are used to statistically reconstruct upper-level fields back to 1880.

Global monthly upper-level temperature and geopotential height fields are reconstructed for the period 1880 to 1957 on six levels (850, 700, 500, 300, 200, and 100 hPa). The predictor comprises several thousands of upper-air and ground-based measurements from before 1957. In the calibration period (1957-2002) the statistical models are fitted using the ERA-40 reanalysis as predictand. The statistical model is based on a principal component analysis and a multiple linear regression approach. Due to the changing station network in the past, for each single month from 1880 to 1957 a separate statistical model is build. To
make maximum profit of the available information present in the predictor data and to avoid overrepresentation of some regions or the northern latitudes, a complex weighting scheme is developed. The quality of the reconstructions is checked with a split-sample approach and with independent upper-air data. Based on the validation, the quality of the reconstruction skill is high over the Northern Hemisphere, especially over Europe and North America. In the Southern Hemisphere only poor reconstructions are expected. Generally geopotential height is better reproduced than temperature and lower levels better than higher levels, except in the Tropics.

In a second step the reconstructed upper-level fields, supplemented with historical pilot balloon data and climate model output, is used to analyze the 1930s “Dust Bowl” drought in the US Great Plains. In the Great Plains most precipitation occurs from April through September related with convective systems. The so-called Great Plains low level jet carries moisture from the Gulf of Mexico into the Great Plains region. In the 1930s this jet was weakened and diverted to the east. In the upper-levels this anomaly was accompanied by a high pressure anomaly in the middle troposphere and a rather zonal pattern in the higher troposphere. Through the pressure anomaly in the middle troposphere the convection was suppressed. The upper-level anomaly resembles the pattern described by Schubert et al. (2004a), relating tropical SSTs with the extra-tropical circulation, though only overall agreement is achieved. Therefore the author concludes that in fact an oceanic forcing is very likely but further research has to be carried out. The reconstructed fields allow a detailed assessment of future model results with respect to the upper-level circulation.

In the European sector the reconstructed upper-level fields are analyzed with a focus on drought events from 1890 to 2002 affecting Western Europe, especially Switzerland. With two different drought indices 14 and 16 drought events have been defined, respectively. These drought events were examined with respect to event-non-event, inter-event and intra-event variability. Drought events in Western Europe are related to anomalous high pressure over north-western Europe in the middle and upper troposphere in late summer, whereas in early summer a low pressure anomaly over the Atlantic is dominating. In winter and early summer the single drought events mainly differ in the position of the high pressure anomaly over north-western Europe whereas in late summer the strength of the pressure anomaly is varying. The drought events go together with colder than normal sea surface temperatures in the tropical and extra-tropical Atlantic. Cassou et al. (2005) linked the high pressure anomaly in the middle troposphere over north-western Europe (“blocking”) with the SST anomalies in the tropical Atlantic, implying an oceanic forcing. Our results support this assumption since comparable patterns are found. Besides the influence from the tropical Atlantic we found significant SST anomalies in the Indian Ocean. That is in agreement with findings by Black and Sutton (2006) and a Rossby wave response to the Indian monsoon heating. Without any knowledge about winter precipitation only small regions in the Indian Ocean and the tropical Atlantic show significant SST signals in the preceding winter. Therefore we assume that European drought events are only as predictable as SSTs in the Atlantic and in the Indian Ocean. The observed link between winter SSTs and drought events is mainly due to winter precipitation and
therefore soil moisture memory. There is only little evidence that an oceanic winter signal is transferred to the following summer via atmospheric circulation anomalies.
Zusammenfassung

Das Klimasystem zeigt sowohl auf regionaler als auch kontinentaler Skala zwischen Jahren und Jahrzehnten eine starke Variabilität, ersichtlich in Feuchtphasen und Dürren. Insbesondere Dürren haben weitreichende Konsequenzen, unter anderem für die Landwirtschaft, die menschliche Gesundheit und den Transport- und Energiesektor.


Höhenmessungen von vor 1957 zusammen mit zusätzlichen Bodenmessungen für die statistisch Rekonstruktion der Höhenfelder von 1880 bis 1957 verwendet.


Über Westeuropa wurden die rekonstruierten Höhenfelder im Hinblick auf Dürreereignisse in der Zeitperiode von 1890 bis 2002 untersucht, welche die Schweiz betrafen. Mit zwei
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Curriculum Vitae
1 Introduction

1.1 Motivation

The fourth IPCC report specifies the increase in global mean surface temperature over the period 1901 to 2005 with about 0.7 °C (IPCC 2007). But global warming was not linear. Periods of rapid warming alternated with more stagnant phases. After a phase of rapid warming at the beginning of the 20th century, which was most pronounced in the Arctic, global mean temperature stagnated from about 1946 to 1975 before it increased again.

In addition to these multi-decadal variations, the record is punctuated by strong multi-annual events such as droughts and wet spells of a regional to continental scale. Due to predicted future temperature increase it is expected that extremes become more frequent and longer lasting. Because of the complexity of the coupling between atmosphere, ocean and land surface the processes which cause the interannual to interdecadal variability are not fully understood. In order to understand the anthropogenic impact on climate it is of interest to understand these phenomena. Different studies showed that extreme events like droughts are predictable to some extend since they are linked to sea surface temperatures. It is still under debate how in detail the oceanic signal is transferred to the continents.

This thesis focuses on droughts in the US and Western Europe. Both are regions with a high population density and a high agricultural productivity and therefore vulnerable to drought conditions. In both regions the occurrence of droughts is linked to oceanic forcing. For the US, influences from the Pacific and the Atlantic are expected whereas Europe is mainly influenced from the Indian Ocean and the Atlantic.

The 1930s “Dust Bowl” drought in the US Great Plains is analyzed in detail. The US Great Plains experienced a number of multiyear droughts during the last century, most notably the droughts of the 1930s and 1950s. The 1930s drought affected almost two-third of the United States and parts of Mexico and Canada. Throughout the decade, with only two short interruptions, the climate was too warm and too dry. The drought and its associated dust storms created one of the most severe catastrophes in the history of the United States. While many non-climatic factors contributed to the disastrous effects of the Dust Bowl drought, the climatic conditions were certainly extreme. In the US the 1930s was the warmest decade on record together with the 1990s. Although the meteorological record from the Earth’s surface is well documented for this period, the causes for this climatic anomaly are still not understood. Recent studies imply an oceanic influence (Schubert et al., 2004a&b;
Zhang and Mann, 2005; Shabbar and Skinner, 2004; McCabe et al., 2004). A meridional displacement of the jet stream and mean storm track is assumed. Zhang and Mann (2005) suggest that cold ENSO events lead to anomalous high pressure over the North Pacific. Resulting from the weakened low pressure system the north-westerly flow toward the US is reduced. Additionally there were negative sea level pressures over the tropical and subtropical North Atlantic during the drought periods leading to a less moisture advection from the Gulf of Mexico.

Western Europe is a region regularly but not permanently affected by droughts. Due to the geographical position, Europe is mainly influenced by westerly flow from the Atlantic year-round and therefore exhibits generally a moderate wet climate. Despite this fact, in some years drought conditions prevail. Most notably in the mid 1940s to the early 1950s there was a period with prolonged droughts (Dai et al., 1998). A variety of studies link drought occurrence with a high pressure anomaly in the middle troposphere over Europe (Beniston and Diaz, 2004; Black et al., 2004; Carril et al., 2007; Della-Marta et al., 2007; Fink et al., 2004; Fischer et al., 2007b; Schär et al., 2004). This “blocking” alters the horizontal advection and therefore the precipitation and temperature patterns. Some studies link variations of sea surface temperatures (SSTs) in the Atlantic with droughts in Europe (Della-Marta et al., 2007; Cassou et al., 2005; Sutton and Hodson, 2005). Other studies relate European droughts with SST anomalies in the Indian Ocean (Black and Sutton, 2006; Rodwell and Hoskins, 2006).

It is evident that upper-level circulation plays a crucial role in the development and evolution of droughts. Data sets consisting of direct upper-air measurements, such as radiosondes or pilot balloons, exist only for the second half of the 20th century (e.g. CARDS: Eskridge et al., 1995, see also Lanzante et al., 2003; HadRT: Parker et al., 1997; IGRA: Durre et al., 2006). Widely used are the so called reanalysis datasets combining a climate model and surface, upper-air and satellite measurements. These data sets are available back to 1948 (NCEP/NCAR: Kistler et al., 2001) or 1957 (ERA-40: Uppala et al., 2005). Regular upper-air observations started already around 100 years ago. In the early years, kites, pilot balloons, aircraft, and registering balloons were the most common measurement platforms, whereas from the early 1930s on more and more radiosondes were used. These early records were never used because they are only available on paper or are not in a suitable format. In the last years many old records have been digitized and their data quality re-evaluated (Brönnimann, 2003a&b; Ewen et al., 2008a&b; Grant et al., 2009).

In this thesis the historical upper-air measurements are used to statistically reconstruct gridded upper-level fields of temperature and geopotential height. The reconstructed fields back to 1880 allow a much more detailed analysis of the upper-level circulation during the 1930s US “Dust Bowl” drought and during European drought events in the early 20th and late 19th century.
1.2 Objectives and Outline

In the previous section droughts in Europe and the US and their links to large scale climate variability are addressed. Although various studies have been conducted with the aim to explain the occurrence, development and evolution of droughts, the knowledge is still fragmentary, especially with respect to droughts occurring in the first half of the 20th century. The reconstruction of global upper-level fields of temperature and geopotential height will provide useful insights with respect to large scale circulation variability.

The main objectives covered in this thesis are:

- to reconstruct global upper-level fields of temperature and geopotential height up to the lower stratosphere for the period 1880 to 1957.
- to analyze the reconstructed data set with a focus on the “Dust Bowl” drought in the 1930s in the US.
- to define circulation anomalies related with droughts in Western Europe during the 20th and late 19th century.

The thesis is subdivided into three sections with the following content:

Chapter 2 (Griesser et al., 2009) describes the reconstruction of upper-level temperature and geopotential height fields. The data set covers 78 years (1880 to 1957) and is available globally with a resolution of 2.5° for six levels (850, 700, 500, 300, 200 and 100 hPa). The reconstruction procedure is outlined in detail. The underlying data together with the applied weighting scheme is explained explicitly. Furthermore the validation procedure with a split-sample approach and with independent upper-air data is presented and the quality of the reconstructions discussed. The usefulness of the reconstructed fields is demonstrated comparing reconstructions with upper-level climate anomalies over the US in the 1920s and 1930s.

Chapter 3 (Brönnimann et al., 2009b) examines the 1930s “Dust Bowl” drought. Pilot balloon soundings over the US are presented together with reconstructed upper-level fields and climate model output. For the drought years (1931-1939), the structure and strength of the Great Plains low level jet is analyzed based on the pilot balloon soundings. Furthermore the “Dust Bowl” drought is linked to circulation anomalies in the mid- to upper-troposphere. The author is second author of the paper and contributed to the work by providing the reconstructions, analyzing various climate variables and reviewing and editing the manuscript. An Appendix gives a more detailed view of the upper-level circulation during the “Dust Bowl” period.

Chapter 4 (Griesser et al., 2008b) presents the reconstructed upper-level fields together with surface fields with respect to circulation anomalies occurring during European drought events. Droughts events affecting Western Europe, especially Switzerland in the
period 1890 to 2002 are identified using two different drought indices. The circulation anomalies are examined comparing drought years with non-drought years. Additionally inter-event and intra-event variability of drought events is determined using principal component analysis. The drought years 1911, 1947 and 1972 are studied in detail.

**Chapter 5** summarizes the work presented, and discusses the layout of subsequent projects.

**Appendix A** (Schaller et al., 2008) describes a related project using the same reconstructed upper-level fields. The circulation anomalies after the Krakatoa eruption in 1883 is examined with a special focus on the polar vortex in the subsequent winters after the eruption. Additionally the eruption is embedded in the historical context. The author is second author of the paper and provided the reconstructed fields and reviewed the manuscript.
2 Reconstruction of global upper-level fields

Reconstruction of global monthly upper-level temperature and geopotential height fields back to 1880

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Abstract

This work presents statistically reconstructed global monthly mean fields of temperature and geopotential height (GPH) up to 100 hPa for the period 1880 to 1957. For the statistical model several thousand predictors were used, comprising a large amount of historical upper-air data as well as data from the earth’s surface. In the calibration period (1957-2002) the statistical models were fit using the ERA-40 reanalysis as predictand. After the weighting of the predictors, principal component (PC) analyses were performed on both the predictand and predictor dataset. Multiple linear regression models relate each principal component time series from the predictand with an optimal subset of principal component time series from the predictor. To assess the quality of the reconstructions, statistical split-sample validation (SSV) experiments were performed within the calibration period. Furthermore, the reconstructions were compared with independent historical upper-air and total ozone data and with a historical reanalysis. Based on the SSV experiment we obtained good reconstructions for temperature and GPH in the Northern Hemisphere, however the skill in the Tropics and the Southern Hemisphere was much lower. The reconstruction skill shows a clear annual cycle with the highest values in January. The lower levels were better reconstructed except in the Tropics where the highest levels showed the best skill. With the inclusion of a considerable amount of upper-air data after
1939 the skill increased substantially. The fields were analyzed for selected months in the 1920s and 1930s to demonstrate the usefulness of the reconstructions. It is shown that the reconstructions are able to capture regional to global dynamical features.

2.1 Introduction

For the study of interannual-to-decadal climate variability in the 20th and the late 19th centuries, a variety of global gridded datasets of different variables on a monthly or daily basis at the surface are available (e.g., HadSLP2, Allan and Ansell 2006; CRU TS 2.1, Mitchell and Jones 2005; HadCRUT3, Brohan et al. 2006; HadISST, Rayner et al. 2003; ERSST, Smith and Reynolds 2004). Gridded datasets going further back in time (e.g., Luterbacher et al. 2002, 2004; Casty et al. 2005; Pauling et al. 2005; Xoplaki et al. 2005) are on a more regional scale (e.g. Europe) and lose temporal resolution.

Although many phenomena can be addressed to some extent based on surface data, an interpretation of these phenomena requires information on atmospheric circulation, which necessarily involves features at upper levels. Datasets consisting of direct upper-air measurements, such as radiosondes or pilot balloons, exist for the second half of the 20th century (e.g., CARDS, Eskridge et al. 1995, see also Lanzante et al. 2003; HadRT, Parker et al. 1997; IGRA, Durre et al. 2006). Parts of these datasets, supplemented with additional information from the surface and satellites, were assimilated into weather prediction models to generate probably the most widely used 3D datasets: ERA-40 and NCEP/NCAR reanalyses (Uppala et al. 2005; Kistler et al. 2001). Although, the continuously updated NCEP/NCAR reanalysis now provides data for the past 60 years (1948-2008), there is still a general lack of knowledge about the variability of the upper-level circulation in earlier times and on an interdecadal timescale.

Several authors have tried to fill in this gap. Klein and Dai (1998) presented a method to statistically reconstruct 700 hPa geopotential heights (GPH) for North America, extending to the Pacific and Atlantic, using surface air temperature (SAT) and sea level pressure (SLP) data. Schmutz et al. (2001) reconstructed 700, 500, and 300 hPa GPH for the European and Eastern North Atlantic region based on SLP, SAT and precipitation (RR) data back to 1900. Gong et al. (2006) derived 500 hPa GPH for the Northern Hemisphere based on SLP and SAT fields back to 1871. All the studies mentioned above have one major shortcoming: they do not include any upper-air measurements. Brönnimann and Luterbacher (2004) pointed to the availability of upper-air measurements before 1948, and they reconstructed 700, 500, 300, and 100 hPa GPH and temperature back to 1939 using surface and newly available upper-air measurements. In this paper we extend and refine this approach and present statistical reconstructions of GPH and temperature for three regions: the Northern Hemisphere (15°N-90°N) (NH), the Tropics (20°S-20°N) (TP) and the Southern Hemisphere (15°S-90°S) (SH). The dataset consists of monthly reconstructions on the levels 850, 700, 500, 300, 200, and 100 hPa for the period 1880-1957, in order to allow a seamless connection to the ERA-40 reanalysis.
Another method of filling this gap is data assimilation. A new reanalysis (Compo et al. 2009) has been produced based only on sea-level pressure data and monthly sea-surface temperature and sea ice fields (hence, no upper-air information), reaching back to 1908. The two data sets are complementary and almost completely independent. Some comparisons of the two are presented in Compo et al. (2009).

This paper is organized as follows: In Section 2.2, the data used for reconstruction and validation are presented. Section 2.3 describes the reconstruction method. Validation results are shown in Section 2.4. Some analyses of reconstructed fields, demonstrating the potential value of these data, are presented in Section 2.5. Conclusions are presented as Section 2.6.

2.2 Data

2.2.1 Terminology

For the reconstruction we define two major time periods. The calibration/validation period (1957-2002) and the reconstruction period (1880-1957). The statistical model relates a predictand (Y) to a predictor (X) dataset. In this reconstruction approach the predictand consisted of the upper-level temperature and GPH fields. The predictor comprised upper-level and surface-based measurements. A third dataset consisting of independent upper-air data was required for validation in the reconstruction period; this is in addition to an initial quality assessment which was already performed in the calibration/validation period based on a split-sample validation approach (see section 2.3.3). In the following sections the datasets are briefly described and their quality discussed. For a deeper discussion of each dataset, the reader is referred to the cited literature.

2.2.2 Predictand data in the calibration/validation period

As predictand a long, global and homogeneous 3D dataset was required. The two datasets used in the majority of cases are the ERA40 and NCEP/NCAR reanalyses. NCEP/NCAR starts in 1948 and is continuously updated to the present (Kistler et al. 2001). ERA-40 starts in 1957 and ends in 2002 (Uppala et al. 2005). The operational forecasting system and assimilation procedure in NCEP/NCAR was designed in the mid-1990s, while the core of ERA-40 was developed after 2000. Therefore, the two reanalyses belong to two different generations. In direct comparisons ERA-40 clearly outperforms NCEP/NCAR (Simmons et al. 2004; Santer et al. 2004; Bengtsson et al. 2004). For this reason we choose ERA-40 as predictand for the reconstruction.

Although most deficits apparent in NCEP/NCAR were removed in ERA-40, some problems remain unsolved and some new problems were added. In the Southern Hemisphere the data coverage is still poor in the early years, especially before 1967 (Uppala et al. 2005). Small jumps in the mean temperatures in the troposphere are present, resulting from differences in the bias correction of satellite measurements, with the largest
inhomogeneity expected around 1975-76. In the pre-satellite years the extratropical Southern Hemisphere exhibits a cold tropospheric bias (Bengtsson et al. 2004). In the same years a cold bias in winter and springtime in the Antarctic lower stratosphere is apparent.

The disadvantage of the shorter time period, compared to NCEP/NCAR, was expected to be at least compensated by the increased data quality. Furthermore, despite the problems with ERA-40 mentioned above, there were other good reasons to use it as predictand. Our reconstruction approach primarily focused on spatial variability patterns for the whole troposphere and the lowermost stratosphere and was therefore less affected by inhomogeneities in either a subregion or a specific layer. Additionally, the month-to-month variability is large relative to the observed jumps. Inhomogeneities in the predictand dataset only affect the quality of the reconstruction to the extent to which they project onto patterns of variability which occur naturally. Also, they do not introduce trends in the reconstruction period. The quality of the reconstruction can be assessed with a statistical bootstrap procedure in the calibration period and additionally with the independent validation data in the reconstruction period. Finally, we reconstructed three independent regions (see section 2.1). The reported shortcomings in ERA-40 are mainly in the Southern Hemisphere and do not influence the reconstructions of the other regions.

In our case we used monthly mean fields of GPH and temperature at the 850, 700, 500, 300, 200, and 100 hPa levels (thereafter termed Z850, T850, Z700, T700 etc.) interpolated to an equal area grid. Hence, the number of grid points on a latitudinal circle decreases towards the poles. The distance on a longitudinal circle was kept constant with a resolution of 2.5°. We have a maximum number (144) of gridpoints at the equator, equivalent to a resolution of 2.5°. This number decreases towards the poles according to the cosine of the latitude.

2.2.3 Predictor data

The predictor data was divided into two major groups: surface data and upper-air data. The surface data again consisted of gridded SLP (HadSLP2, Allan and Ansell 2006) and homogenized surface station temperature data (GISSTEMP, Hansen et al. 1999). The SLP dataset was incorporated “as is” with a spatial resolution of 5° by 5° and spanning from 1880 to 2002.

For the surface temperature predictors, stations with high data quality and good spatial and temporal coverage are preferred. Therefore, the GISSTEMP station network was reduced according to the following criteria: first, all stations with less than 90 percent of possible data available in the historical period were eliminated. Second, we calculated the Pearson correlation between the temperature anomalies of each individual station and ERA-40 925 hPa temperature anomalies interpolated to the station location. Stations with a correlation <0.8 were removed. Third, because the US still showed an overrepresentation relative to other regions, which is potentially problematic with regard to the weighting, the station network over the US was further reduced. US stations with an incomplete record in the 20th century were discarded. Based on criteria just described, a global subset of 760 total
stations was extracted covering the period from 1880 to 2002. The location of the surface temperature stations and the temporal evolution of the number of predictors are presented in Fig. 2.1. After subtracting the annual cycle based on the period 1961 to 1990, the data were standardized, and the few remaining missing data points in the calibration period in the extracted surface station network were filled with standardized 925 hPa anomalies from ERA-40 in order to have complete data series. Brönnimann and Luterbacher (2004) showed that this is justified by the high median correlation of 0.85 between the reanalysis and the station series.

For the upper-air data we distinguished between measurements taken by radiosondes, kites, aircraft, and pilot balloons. All upper-air measurements were from the period before 1958 and originated from many different sources. The radiosonde data were collected from digital archives and were compiled and quality controlled (Grant et al. 2009). They were supplemented with additional historical radiosonde, aircraft, and kite measurements which were processed at ETH Zurich (Brönnimann 2003ab; Ewen et al. 2008b). In addition, reevaluated upper-level wind data from the global TD52 and TD53 pilot balloon datasets provided by NCAR (available online at http://dss.ucar.edu/docs/papers-scanned/papers.html, documents RJ0167, RJ0168) and from the African pilot balloon dataset of MétéoFrance were used. The pilot balloon data were checked for errors with the same procedure as described by Grant et al. (2009) for radiosonde data. In the cases where it was not clear if the station should be accepted or rejected, generally because of a too weakly correlated reference series, the variance and the mean of the historical period were plotted against the same variables from ERA-40, at the same location to assess the quality. Station series with a bias of more than two standard deviations, or a difference in the variance of more than 1.5 standard deviations between the historical period and the reanalysis were rejected if the historical time series was longer than one year. Whenever the majority of the levels from a station showed inconsistency with the reanalysis, the complete station was removed.

All upper-level series cover only a part of the historical period and most do not reach the present time or have long gaps, often because stations were relocated or closed or the measurement platform changed; for instance no kite data are available after the 1930s. In these cases, a substitute must be found for calibration. Therefore, ERA-40 was used to supplement all historical upper-level series after 1958. The only exceptions were the TD52 and TD53 datasets after 1948, which were rigorously quality-checked in a previous study (Ewen et al. 2008a) using the NCEP/NCAR reanalysis. Data from that study was used after supplementation with NCEP/NCAR after 1948. The location of all upper-air predictors as well as the measurement platforms is shown in Fig. 2.1. Globally, 15 394 upper-air predictor variables in the historical period were used for the reconstruction (7752 kite/aircraft or radiosonde, 7642 pilot balloon).
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Fig. 2.1 a) Map of surface and upper-air stations used as predictors. Shaded areas indicate the three regions (NH: Northern Hemisphere 10°N-90°N, TP: Tropics 30°S-30°N, SH: Southern Hemisphere 10°S-90°S) used for the reconstruction. Green triangles represent surface temperature stations, blue crosses denote pilot balloon stations, black circles denote upper-air series taken by radiosondes, kites or aircrafts and red circles are upper-air stations used for the validation. b), c) and d) Number of available predictors from 1880 to 1957 separated by measurement platforms for the regions NH, TP and SH, respectively.
The quality of historical data (especially upper-air data) is lower than that of more recent measurements. Therefore, the predictor data after 1957 were perturbed with normally distributed noise. The noise consisted of two parts, a random (i.e., time independent) bias for each station and a purely random component. The standard deviation of the normal distributions of the noise was deduced from our quality assessment (Brönnimann 2003a; Grant et al. 2009). For upper-air temperature data we assumed a random station bias with a standard deviation of approximately 0.5°C and a completely random component with a standard deviation of roughly 1.1°C. For all wind data we deduced 0.7 m/s for the standard deviation of the random station bias and 1.1 m/s for the purely random part. In contrast to temperature and wind, where the error is kept constant with height, the error for GPH increases from the lower to the higher levels. From a standard deviation of 7.5 gpm in the 850 hPa layer, the noise grew to a standard deviation of 20 gpm in the 100 hPa level for the station bias, and from a standard deviation of 11.5 gpm to a standard deviation of 53 gpm for the completely random noise. After perturbation, all predictor variables were standardized and expressed as anomalies with respect to the 1961 to 1990 annual cycle.

The data availability for any given month in the historical period was much more limited than suggested by Fig. 2.1. Except for the SLP data, all data series had longer or shorter gaps in the historical period. The earliest upper-air series used for the reconstruction started in 1920, and a large amount of data was confined to the lower troposphere. The coverage was much better for the continents and the Northern Hemisphere. For the Tropics and the Southern Hemisphere, the coverage was poor and upper-air data were available only from the beginning of the mid 1930s.

### Validation data

For the purpose of validation, some upper-air stations were retained and not used for the reconstruction. Stations were selected according to the following criteria: first, the stations had to cover as much of the historical period as possible, preferably with no gaps. Second, to keep the validation of the reconstruction independent from the quality control procedure of the predictors, only stations which did not need any correction were used. Based on these criteria, three stations were withheld: Oakland (USA), Ellendale (USA) and Lindenberg (Germany) (See Fig. 2.1 for their exact position). Lindenberg was the upper-air station with the longest available record, going back to 1905. Furthermore, the reconstruction was compared with independent reconstructions from Brönnimann and Luterbacher (2004) and Schmutz et al. (2001).

In addition to historical upper-air data, we used historical total ozone data to assess the reconstruction (see Brönnimann and Staehelin (2004) for a brief description of the technique of validating upper-level reconstructions with total ozone). At mid-latitudes total ozone is known to be well correlated with meteorological variables in the tropopause region. Historical total ozone data were available for time periods (e.g., the 1920s) and regions (e.g., Australia or China) for which no upper-air data were available and thus allowed some conclusions about the skill of the reconstructions in these cases. We used
monthly mean values of total ozone at Arosa (Switzerland), 1926-2002 (Staehelin at al. 1998), Tromsø (Norway), 1935-1972 (Hansen and Svenøe 2005), Oxford (UK), 1924-1975 (Vogler et al. 2007), New York (USA), 1941-1944, Shanghai (China), 1932-1942 (Brönnnimann et al. 2003), and Canberra (Australia), 1929-1932 (unpublished; re-evaluated as in Brönnnimann et al. 2003). Where no station data were available after 1978, TOMS total ozone data (Version 8) were used to supplement the series.

2.3 Reconstruction method

2.3.1 Weighting scheme

The available historical predictors were unequally distributed in space. In general, there was an overrepresentation of the earth’s surface compared to the middle and upper troposphere, and continents were better covered than oceans. This fact potentially results in an unintentional focus on small scale variability near the surface and over land masses. Hence, the station series must be weighted to better represent the whole variability present in the predictor dataset.

As a first step, all data series were assigned to an altitude level (L0: surface, L1: 250-3000 m or 925-700 hPa, L2: 3001-6000 m or 699-500 hPa, L3: 6001-9000 m or 499-300 hPa, L4: above 9000 m or below 300 hPa). For the reconstruction of the Northern Hemisphere (15°N-90°N) predictors from north of 10°N were included, for the Tropics (20°S-20°N) predictors from 30°S to 30°N were used and for the Southern Hemisphere (15°S-90°S) all predictors south of 10°S were used. Second, within each level and for the variables GPH, temperature, and wind (u- and v-winds were treated as a single variable), the average 0.5 decorrelation distance was calculated, yielding an estimation of the “influence radius”. Influence radii for each variable and level are given in Table 2.1.

Table 2.1 Average radius [km] for a given level (L0: surface, L1: 250-3000 m or 925-700 hPa, L2: 3001-6000 m or 699-500 hPa, L3: 6001-9000 m or 499-300 hPa, L4: above 9000 m or below 300 hPa), beyond which the spatial correlation drops below 0.5, which defines the “influence” radius of the stations.

<table>
<thead>
<tr>
<th>Level/Variable</th>
<th>Temperature</th>
<th>GPH</th>
<th>Wind</th>
</tr>
</thead>
<tbody>
<tr>
<td>L4</td>
<td>1529</td>
<td>1483</td>
<td>1311</td>
</tr>
<tr>
<td>L3</td>
<td>1379</td>
<td>1425</td>
<td>1267</td>
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<tr>
<td>L2</td>
<td>1398</td>
<td>1448</td>
<td>1142</td>
</tr>
<tr>
<td>L1</td>
<td>1421</td>
<td>1487</td>
<td>1017</td>
</tr>
<tr>
<td>L0</td>
<td>1266</td>
<td>-</td>
<td>-</td>
</tr>
</tbody>
</table>
The weight for each individual station and variable was the inverse of the number of all available stations with information from the same variable within the influence radius. Finally, the weights were adjusted such that the overall weight of a variable in a level was proportional to the total area covered by all the influence radii combined. The covered area as a percentage of the total area for each region and variable are shown in Fig. 2.2, and a map showing the coverage for the month indicated in Fig. 2.2 is given in Fig. 2.3. This procedure was repeated for each time step. Within the surface level (L0), 50% of the weight was attributed to SLP and 50% to the surface station temperature field. The SLP field was additionally area weighted.

![Graphs showing temporal evolution of coverage for different variables, regions, and levels.](image)

**Fig. 2.2** Temporal evolution of the percentage of total area which was covered for a given variable (Temperature (left), GPH (middle), and Wind (right)), region (NH (top row), TP (middle row), and SH (bottom row)) and level (L0-L4, color). The dashed vertical line indicates the month for which the total area coverage is presented in Fig. 2.3.

### 2.3.2 Statistical model: Setup

After the regridding of the predictand to an equal area grid and after the perturbation of the predictor, the regression model was set up. Here the model is defined as a set of regression equations linking the predictors and predictands for a specific month. The statistical model was fit in the calibration period and the derived relation was applied to the reconstruction
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period. The approach used here was based on a principal component (PC) regression model, similar to Brönnimann and Luterbacher (2004).

As described in the data section, the predictor network in the historical period changes over time and longer or shorter gaps existed in some predictors. To make use of all available data a separate statistical model for each historical month was created.

![Map showing the covered area, defined by the influence radii, for January 1944 and the variables GPH, temperature and wind. The different shadings indicate the levels L0-L4 (see text).](image)

Fig. 2.3 Map showing the covered area, defined by the influence radii, for January 1944 and the variables GPH, temperature and wind. The different shadings indicate the levels L0-L4 (see text).

To calibrate a model for a specific month in the past, a three month moving window centered on the associated calendar month was used. For the reconstruction of January 1944, for example, all data from the months December, January, and February in the calibration period were selected. In a further step, only those predictor series which were available in the defined month in the reconstruction period were selected for the calibration period. The extracted subset of predictor variables was multiplied by the square root of the weighting field pertaining to the specified historical month (for the weighting field, see section 2.3.1). Next, one PC analysis was performed on the predictand data set (standardized, all variables and levels combined) and a second on the predictor subset. Each predictand PC time series was then expressed as a linear combination of an optimal subset of predictor PC time series using linear regression (least-squares estimator). In order to obtain the best model, only a certain number of predictor and predictand PCs were retained. The retained variance was varied between 70% and 98% (independently on both the predictor and the predictand side) and the subset with the best performance was chosen for the reconstruction (where “best performance” is measured according to the split-sample validation: see section 2.3.3). The predictand PCs were generated by applying the coefficients to the corresponding predictor PCs in the reconstruction period. The reconstructed anomaly field is a function of the reconstructed predictand PCs which were retained and the predictand PC scores from the calibration period. Finally, the standardization procedure was inverted and the data were regridded to a 2.5° by 2.5° grid.
To reconstruct monthly values for 78 years (1880-1957) 936 individual models were formed for each of the three regions. Every model consisted of an iteratively solved set of regression equations based on least squares.

### 2.3.3 Statistical model: Validation

Before the final reconstructions were carried out, different sensitivity experiments were performed concerning the robustness of the reconstruction and weighting method, the definition of appropriate regions for the reconstruction (NH, TP, SH), and whether to reconstruct all levels and variables together or separately (not shown). To determine the potential benefit from the inclusion of upper-air predictors, reconstructions with only surface data were also conducted.

The reconstructions were validated by using the split-sample validation (SSV) technique, a special case of a cross validation. The calibration period for the final reconstruction (1957-2002) was split into a calibration part and a validation part for the SSV model. The statistical model was derived from the data in the SSV calibration period and tested in the independent SSV validation period. This procedure was repeated twice with different time periods. The model was fit either in the period 1958-1987 or 1972-2001 and tested in the period 1988-2001 or 1958-1971, respectively. The potential skill of the model was measured with the reduction of error statistic (RE, Cook et al. 1994) defined as

\[
RE = 1 - \frac{\sum_i (x_{rec} - x_{obs})^2}{\sum_i (x_{null} - x_{obs})^2}
\]  

(2.1)

where \(t\) is time, \(x_{rec}\) is the reconstructed value, \(x_{obs}\) is the observed value and \(x_{null}\) is the null hypothesis. For reconstructed anomalies, the null hypothesis corresponds to a zero anomaly from the long-term mean annual cycle (1961-1990). Values of RE can be between \(-\infty\) and 1 (perfect reconstruction). An RE of 0 is indicative of a reconstruction not better than climatology, whereas an RE > 0 points to a model with predictive skill. Due to stochastic properties RE values can be above zero by chance. Therefore we consider reconstructions useful if RE values are above 0.2. This approximately corresponds to \(R^2\) equal to 0.2-0.25 (see Brönnimann and Luterbacher 2004). Because the validation period in the SSV procedure is 14 years long, equation (2.1) sums over 14 time steps. The result of each SSV experiment is a spatial field of RE values on the predictand grid. For the model validation it is useful to aggregate the information into a single number. As the RE skill score has a fixed upper boundary at one, distributions of RE values tend to be skewed. In this case the appropriate location estimator is the RE median. For the selection of an optimal subset of predictand and predictor PCs, the RE median over the entire field was calculated and maximized. For the analysis of the fields usually the average RE value from the two split sample validation was given.

In the work by Ewen et al. (2008a) a subsample of our predictor data set was used for the reconstruction of an upper-level index (the Pacific North American pattern) using a very similar approach. In that study, additional validations were performed in a surrogate
climate using NCAR CCSM3 model output (Collins et al. 2006). The reconstructions in the surrogate climate showed almost identical skill to that obtained from the SSV in the reanalysis data. The conclusion was that this reconstruction method is robust in the model environment. It would be beneficial to repeat the same validation experiments for the full reconstruction presented here, but that is beyond the scope of this paper.

2.4 Validation results

2.4.1 Split-sample validation

Results from the SSV are summarized in the form of time series of the field median value of RE for a region (NH, TP, SH), or as maps showing the RE field for a specific level and month or period. As the SSV used only 30 years for calibration, the results provide a rather pessimistic estimation of the skill. For the final reconstructions, when calibrating with the full ERA-40 dataset, better results are expected. Through the artificial disturbance of the predictors in the recent years, we account for the lowered data quality in the reconstruction period. Therefore, we assume that no overestimation of the model skill is expected due to the higher data quality within the ERA-40 period.

The SSV showed distinctly different results for the three independently reconstructed regions. Based on the SSV (see Fig. 2.4 for the median of the RE-fields for selected levels), the performance of the statistical model was best for the Northern Hemisphere, intermediate for the Tropics and poor for the Southern Hemisphere. Nevertheless, some findings concerning the quality of the reconstructions are valid in general. The \( \text{RE}_{\text{median}} \) time series from all regions and levels show an annual cycle, most evident in the GPH fields in the Northern Hemisphere, with the highest \( \text{RE}_{\text{median}} \) usually found in January in all regions. Geopotential height was generally better predicted than temperature and lower levels were better reproduced than higher levels. The inclusion of upper-air predictors increased the reconstruction skill in all regions and on all levels, most pronounced for higher levels and temperature.

Though similarities between the individual regions appeared in the SSV results, considerable differences were observable. The \( \text{RE}_{\text{median}} \) values in the Northern Hemisphere showed the clearest annual cycle with a maximum in January and a minimum in June. In the period without (before 1920) or with few (before 1939) upper-air data, the skill of the reconstruction was good during the Northern Hemisphere winter for GPH at all levels and for temperature in the lower levels (T850, T700 and T500). Furthermore, the SSV results suggest poor predictability for the T300, T200 and T100 levels before 1939.

The inclusion of a considerable number of upper-air predictors after 1939 increased the reconstruction skill substantially. The effect was largest for temperature and at the higher levels, bringing the \( \text{RE}_{\text{median}} \) of the different levels closer together. With \( \text{RE}_{\text{median}} \) values between 0.6 and 0.8 during the Northern Hemisphere winter, good reconstructions were found for GPH after 1940, whereas reasonable to good reconstructions are obtained for
Fig. 2.4 Time series of RE$_{median}$ as a function of variable, level, and region from the split-sample validation (left). RE values calculated from comparing the ERA-40 and NCEP/NCAR reanalysis for the period 1957 to 1971 (summed over the 14 years with respect to the calendar month) (right).
GPH during summer and temperature the whole year round. The $RE_{\text{median}}$ in the Tropics showed a clear annual cycle with a superimposed semiannual signal. Normally $RE_{\text{median}}$ was highest in January. For GPH at the higher levels, this peak was shifted to February. A secondary maximum was reached in July. The annual cycle of the $RE_{\text{median}}$ for the Tropics is larger than for the Northern Hemisphere.

Whereas the reconstructions for Z850 and Z700 were reasonable to good from 1880 through the 1950s, $RE_{\text{median}}$ remained low for GPH in the higher levels and for all temperature levels before the mid 1930s. The inclusion of upper-air data after 1935 raised the skill of the reconstruction model, especially for GPH during summer. The $RE_{\text{median}}$ increased most for the highest levels for both GPH and temperature. In the Z100 level (not shown), $RE_{\text{median}}$ values of 0.85 were reached in the 1950s. Although the annual cycle of the $RE_{\text{median}}$ was reduced by introducing upper-air predictors, very low skill can be observed even in the 1950s in April to May or August to September for some years. In the Southern Hemisphere the skill was very low before the 1950s. The only exceptions were the Z850 and Z700 levels, which showed an acceptable $RE_{\text{median}}$ of 0.2 to 0.4. Although the few upper-air predictors raised the skill considerably, the $RE_{\text{median}}$ remained on a low level even in the 1950s. Surprisingly, the maximum of the $RE_{\text{median}}$ was reached in January.

To get an estimate of the maximum expected skill, the $RE_{\text{median}}$ was calculated for a comparison of the two reanalyses in the period 1957 to 1971. For the sake of comparability with the results obtained from the reconstructions, the reanalyses anomalies were interpolated to the equal area grid described for the predictand. For the calculation of $RE$ (see equation 2.1), $x_{\text{obs}}$ was defined as ERA-40 and $x_{\text{rec}}$ as NCEP/NCAR. The results are shown in Fig. 2.4 (right). In the Northern Hemisphere both reanalyses show excellent agreement for GPH and temperature at all levels. While in the Tropics the skill for the higher levels and GPH was still good, it dropped considerably for temperature on all levels and for GPH on the lower levels, especially from May to August, probably due to small scale processes not resolved in the reanalyses, like convection. In the Southern Hemisphere skill was low except for GPH from October to March and for 500 hPa temperature.

While the $RE_{\text{median}}$ is a good measure for the overall fit of the model, the spatial details deserve more attention. Even if the field median RE value was low, some regions still showed good skill, depending on the season and available station network. Figs. 2.5 and 2.6 show fields of $RE_{\text{median}}$ values calculated over the period 1900 to 1904 and 1940 to 1944, respectively. The earlier period represents a time when no upper-air predictors were available, whereas in the later period the station network was relatively well developed. As already seen from the analysis of the $RE_{\text{median}}$ time series, RE values for GPH were clearly better than for temperature, and lower levels were better reproduced than higher levels (except for the Tropics). This fact also appeared in the RE patterns. For the Northern Hemisphere, independent of whether upper-air data were available or not, RE is high over North America, Europe and East Asia/Japan. Poor reconstructions were found over parts of Central and Northeast Asia, North Africa, parts of the Atlantic and Pacific Subtropics and the North Polar Region. Upper-air data increased the skill in general and especially over
Fig. 2.5 Maps showing fields of $\text{RE}_{\text{median}}$ for the period January 1900 to December 1904 for GPH and temperature on the levels 700, 300, and 100 hPa.

Fig. 2.6 As in Fig. 2.5 but for the period January 1940 to December 1944.
central Asia and northwestern North America. In the Tropics skill was high over the Pacific except for the T100 level. Lower skill was found over the Atlantic, Central Africa and Central America for some levels.

The T100 level showed exceptionally low RE values, particularly spanning from the Indian Ocean over Africa to Central America. Inclusion of upper-air data generally increased the RE. For the Southern Hemisphere, although RE_median time series implicated low skill, good skill was found over the southeastern part of Australia.

In Figs. 2.7 and 2.8 the Z300 RE fields for January and July 1904 and 1944, respectively, are shown. While in 1904 the skill remained high year-round for temperature and GPH over central North America, the northern Atlantic, Western Europe, the Tropical Pacific, the Indian Ocean, and parts of Australia, some regions showed much lower RE values in July.

Fig. 2.7 Maps showing RE fields for GPH and temperature on the 300 hPa level for January and July 1904.
Most striking are the low values over the Northwestern Pacific, parts of Asia, Central Africa extending to the Eastern Atlantic, and the whole Antarctic. Because of the inclusion of upper-air data, the skill in 1944 was generally better year-round. Some regions showed an increase in the RE values above average, especially in July. For example, the SSV exhibited skill for northwestern North America and the central and northern parts of Australia in 1944, whereas there was no skill at all in 1904.

To separate the influence of including upper-air predictors from the effect of a changing surface station network, sensitivity experiments with only surface predictors were performed. Fig. 2.9 shows RE fields for Z700, T700, Z300, and T300 for the Northern Hemisphere with (upper panels) and without (lower panels) upper-air data. While the inclusion of upper-air data increased the RE values for GPH on a global scale, the effect on temperature was more regional and is best visible over Northern America and Northern Europe.
**Fig. 2.9** RE fields for GPH and temperature for January 1944 in the Northern Hemisphere (NH).  
**a)** With all available upper-air predictors included. Blue triangles in the left panels: all available upper-air predictors for the specified month. Blue triangles in the right panels: upper-air predictors available only above the 700 hPa level. **b)** Only surface predictors.

### 2.4.2 Validation with historical upper-air data and independent reconstructions

In addition to the SSV, the reconstructions were compared with independent historical upper-air data (i.e., data not used for the reconstruction), as this is the only way to validate the final reconstruction. RE values were calculated in the same way as the SSV. For comparison, the data were pooled for each level. RE values were between 0.57 and 0.81 for GPH and between 0.33 and 0.55 for temperature. The corresponding correlations were between 0.75 and 0.81 for GPH and between 0.48 and 0.75 for temperature. Surprisingly, RE and correlation were highest for Z100 (although n is only 15). RE values based on historical data were clearly higher than the annually averaged $RE_{median}$ time series from the SSV from the same time period. It must be noted that the validation data were from regions with a dense station network and therefore with a high reconstruction skill in the SSV. When taking into account the spatial distribution of the validation stations, comparable results are found for the historical data and the SSV. Figs. 2.10 a) – f) show scatter plots of reconstructed anomalies versus historical data. The plots show a good overall agreement, although the variability is underestimated (due to the least squares fitting). The comparison
confirmed the result from the SSV. The reconstruction was generally better for GPH than for temperature, for which the correlation drops off at 300 hPa.

Figs. 2.10 g) – k) show time series of monthly anomalies for selected periods and two validation sites. The error bars show the uncertainty in the observations of ± 1 °C for both levels and ± 30 gpm and ± 20 gpm for the 500 hPa and 700 hPa levels, respectively. For the reconstructions the 95% confidence interval was determined from the SSV. The overall agreement is excellent. Extremes are well represented, although slightly underestimated due to our reconstruction approach. Data and reconstructions are mostly within each other’s confidence intervals. For Oakland in 1942 a cold temperature and low pressure bias appears. Since the number of predictors is large in these years, it is unlikely that such a bias is real. We therefore suspect that this is a remaining data problem.

Unfortunately the data coverage in the Southern Hemisphere and parts of the Tropics was very low. Therefore no independent upper-air validation data with the required quality were available in these regions. However, it has already been shown in the SSV results that poor skill is to be expected for the Southern Hemisphere. Through the recovery of additional historical upper-air data, reconstructions and the corresponding validation will improve in the future.

The reconstruction agreed well with independent reconstructions from Schmutz et al. (2001) and Brönnimann and Luterbacher (2004). All levels for both reconstructions showed correlations above 0.8 for the pooled grid cells for each level. For the reconstruction by Schmutz et al. (2001) the correlation slightly dropped from 0.85 in the 700 hPa level to 0.81 in the 300 hPa level. Since Schmutz et al. (2001) do not include any upper-air data, this was expected. The reconstructions reproduce the 1940-42 anomalies at upper-levels (related to El Niño) shown by Brönnimann et al. (2004).

Comparisons were also performed with the historical reanalysis of Compo et al. (2009). An excellent agreement was found for the 1940s El Niño case and for most of the features during the “Dust Bowl” droughts of the 1930s. Some of the comparisons are presented in Compo et al. (2009).

2.4.3 Validation with historical total ozone data

The results of the validation of the reconstructions against total ozone are shown in Fig. 2.11. To calculate the correlations, the reconstructions were merged with ERA-40 and the mean seasonal cycle from November 1978 to October 1994 (before the gap in the TOMS sensor) was subtracted from the all series. Then correlations were calculated separately for the pre-1957 and post-1958 period. Note that the correlations with total ozone show a characteristic profile: Correlations increase in strength from the surface to the upper troposphere. They reach a maximum at about 200 hPa. Above that level, the correlation with temperature changes sign.
Fig. 2.10 a) through f) Observed and reconstructed anomalies of GPH and temperature at 500, 300, and 100 hPa for Ellendale (green), Lindenberg (blue) and Oakland (red). g) through k) Time series of observed (brown) and reconstructed (blue) anomalies of geopotential height g) and temperature h) at 700 hPa at Lindenberg for 1905 to 1917 and 500 hPa at Oakland for the period 1938 to 1945 (j and k, respectively). Error bars give the assumed uncertainty of the observations and the 95% confidence intervals for the reconstructions. Anomalies are with respect to 1961-1990.
Fig. 2.11 Profiles of the coefficient of correlation between total ozone and local upper-level variables for six sites. Grey lines are for temperature, black lines for GPH. Solid lines represent the period 1957-2001, dashed lines represent the reconstruction period. The area of insignificant correlation in the reconstruction period is shaded.

In general, the correlation profiles from the reconstructions (dashed lines) agreed well with those from the reanalysis period. Deviations can have had three causes: a) the low number of historical total ozone data (to account for this, the region of insignificant correlation for the historical period is shaded), b) a low skill of the reconstructions, and c) inaccurate total ozone data. Three of the total ozone series have been homogenized and can be considered to have a higher quality: Arosa, Oxford, and Tromsø. In fact, the agreement between the correlation profiles was excellent for these sites (note that lower correlations were expected for Tromsø; see solid lines). Good agreement was also found for New York. Results were particularly interesting for Canberra (1929-1932) and Shanghai (1932-1942), which represents periods and regions for which no upper-air data was included in the reconstructions (though the historical total ozone series are clearly of lower quality). Both sites showed significant correlations in the troposphere, confirming that the reconstructions have skill even in these cases.
2.5 An analysis of reconstructed fields

In this section an analysis of selected fields is presented in order to point to possible applications of the reconstructions. Anomaly fields (with respect to the 1961-1990 mean annual cycle) are shown for winter (DJF) 1925-26, summer (JJA) 1936 and winter (DJF) 1936-37. This selection was made because surface temperature and station temperature anomalies for different heights over the US for the same months were already presented in a previous study by Ewen et al. (2008b, Figs. 9-11), allowing a direct comparison. During the summer of 1936, in the middle of the decade of the “Dust Bowl” drought, high precipitation deficits and surface temperature anomalies occurred over the central US. The winters 1925-26 and 1936-37 represent extreme positive and negative values of a PNA index, respectively (see also Ewen et al. 2008a). Fig. 2.12 shows the reconstructed fields of GPH and temperature for the levels 700, 300, and 100 hPa in the Northern Hemisphere. Shading denotes areas with poor skill (RE < 0.2) based on the SSV results.

On the 300 hPa level in both winters, a clear PNA-like pattern appears with the two north-south oriented centers over the North Pacific and the two northwest-southeast oriented centers over the United States. In the winter 1925-26 the centers over the US were slightly rotated counterclockwise, whereas in 1936-37 the northern center over North America extended into the Atlantic. Closer to the surface, the PNA signal becomes weaker but the pattern remained the same. Besides the strong PNA-like pattern over the US, a prominent positive anomaly over Scandinavia is visible. This feature was potentially connected to increased transport of warm air into the Arctic region, seen in the lower troposphere (see e.g. Grant et al. 2008). On the 100 hPa level, a weak (1925-26) or strong (1936-37) negative anomaly over the Pole is observable, pointing to a strong polar vortex. The GPH feature in the Arctic is consistent with the cold temperatures in the Arctic stratosphere in the same years (1925-26 and 1936-37). Although the reconstruction approach was designed to capture the large scale (global) features, regional features like the temperature gradient in the lower troposphere (see Ewen et al. 2008b) over the US from warm in the southeast to cold in the northwest in the winter 1936-37 were well captured. Even the weakened gradient with height depicted by the single stations was present in the reconstructions. The same gradient in observations and reconstructions but with opposite sign was apparent in 1925-26.

In the summer of 1936 a positive GPH anomaly on the 300 hPa level over the US was present. Mid-tropospheric ridging has already been linked to later Midwest droughts by Namias (1982), but was never shown to have actually occurred during the 1930s. Because this feature extended into the lower troposphere, a dynamical cause is likely. In the case of a mainly thermally driven feature, negative anomaly would be expected near the surface. The GPH anomaly is embedded into a wave train with further positive nodes over the Aleutian Islands, Northern Europe and Eastern Asia, and negative nodes in the eastern North Pacific, northwestern North Atlantic and Russia. The surface temperature anomaly over the US in the summer of 1936 (see Ewen et al., 2008b) extended into the middle troposphere. At the same time, warmer than normal temperatures were observed over the
Arctic region in the lower troposphere, possibly as a consequence of the positive GPH anomaly over Northern Europe which was mentioned above.

![Reconstructed anomaly fields of 700, 300, and 100 hPa GPH and temperature for winter (DJF) 1925-26, summer (JJA) 1936 and winter (DJF) 1936-37. Light shading denotes areas with RE < 0.2. Anomalies are with respect to the 1961-1990 period.](image)

**Fig. 2.12** Reconstructed anomaly fields of 700, 300, and 100 hPa GPH and temperature for winter (DJF) 1925-26, summer (JJA) 1936 and winter (DJF) 1936-37. Light shading denotes areas with RE < 0.2. Anomalies are with respect to the 1961-1990 period.

### 2.6 Conclusion

Global monthly mean fields of temperature and GPH up to 100 hPa have been reconstructed for the period 1880 to 1957. For the reconstruction a statistical model was set up in the calibration period (1957-2002) linking the predictand and the predictor dataset. The derived model and coefficients were applied to historical predictor data in the reconstruction period. As predictand the ERA-40 reanalysis was used. The predictor consisted of surface data like gridded SLP and surface station temperature and upper-air measurements taken by radiosondes, kites, aircraft, and pilot balloons from the historical period before 1958. A total of 15 394 upper-air predictor variables were used. To factor in
the lowered data quality in the reconstruction period the predictor data in the calibration period was perturbed with normally distributed noise.

The quality of the reconstruction was checked based on a statistical split-sample approach, independent historical upper-air measurements, historical total ozone data, and independent reconstructions. Validation experiments revealed a good overall quality. However, there was a large spread in the skill of the reconstructions. When working with the reconstructions, the seasonal and spatial variability of the skill has to be considered.

The best reconstructions were found for GPH and the Northern Hemisphere in the winter season, and lower levels were better reproduced than higher levels. The results for the Tropics and the Southern Hemisphere should be interpreted with care. Over the ocean the model often showed poor skill. It is important that the reconstructed fields are used only after due consideration of the corresponding RE fields.

The application of the reconstructed fields to selected months showed that the reconstructions are suitable for studying the large-scale circulation. Through the recovery of historical upper-air data the skill of the reconstructions could be improved in the future, especially in the Tropics and in the Southern Hemisphere.

2.7 Acknowledgements

This work was supported by the Swiss National Science Foundation, Project “Past climate variability from an upper-level perspective”. We wish to thank all data providers, especially Roy Jenne (NCAR) and Tom Ross (NOAA/NCDC) as well as MétéoFrance for providing pilot balloon data. Wolfgang Adam (DWD, German Weather Service) provided the historical data from Lindenberg that was used for the validation.
3 The 1930s “Dust Bowl” drought in the US

In the following paper the “Dust Bowl” drought in the US Great Plains is analyzed. The author is third author of the paper and contributed to the work by providing the reconstructed upper-level fields, analyzing various climate variables and reviewing and editing the manuscript. An additional appendix (not included in the submitted version) provides a more detailed view of the upper-level circulation during the “Dust Bowl” period.

**Exceptional Atmospheric Circulation during the “Dust Bowl”**

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**Abstract**

The three-dimensional, regional and large-scale atmospheric circulation during the “Dust Bowl” is analyzed based on newly available historical upper-air data and reconstructed upper-level fields. The Great Plains Low Level Jet, transporting moisture into the region, was weakened on its eastern side, shallower, and penetrated less far north than during wet years. Nocturnal convection was likely suppressed by increased stability. Strong mid-tropospheric ridging was found over the Great Plains, and upper-tropospheric flow anomalies extended from the North Pacific across North America to the Atlantic. These findings provide a dynamical view of the “Dust Bowl” droughts, some aspects of which are distinct from other droughts. It is demonstrated that this is important for assessing predictive capabilities of current modeling systems.
3.1 Introduction

The midwestern United States is a region repeatedly affected by droughts, the most famous being the “Dust Bowl” droughts of the 1930s, with devastating social and economic consequences (Worster, 1979). The mechanisms behind this event are still not fully understood. Based on model simulations, oceanic forcing has been suggested as a trigger (Seager et al., 2005, 2008; Seager, 2007; Woodhouse and Overpeck, 2000; McCabe et al., 2004; Schubert et al., 2004ab, Cook et al., 2008), amplified by land-atmosphere interactions (Seager et al., 2005, 2008; Seager, 2007; Woodhouse and Overpeck, 2000; Schubert et al., 2004ab) or atmospheric dust (Cook et al., 2008). However, due to lack of data, comparisons with observations were previously limited to the Earth’s surface (mostly precipitation), which is insufficient to conclusively address effects of remote forcings. Here we analyze the 3-dimensional atmospheric circulation during the “Dust Bowl” based on historical observations, reconstructions, and climate model simulations.

The forcings responsible for the “Dust Bowl” droughts, as found in model simulations, are expected to proceed through changes in atmospheric circulation. Therefore, causes are best identified by evaluating simulations with respect to the response of the atmospheric circulation to the imposed forcing. In the following we focus on three specific circulation features which based on the literature are expected to be critical to understanding the “Dust Bowl” droughts: the Great Plains Low Level Jet (GPLLJ), mid-tropospheric ridging, and the flow in the upper troposphere.

The GPLLJ is responsible for most of the moisture advection from the Gulf of Mexico, the Caribbean Sea, and the North Atlantic into the Great Plains region (Rasmusson, 1967; Helfand and Schubert, 1995; Higgins et al., 1997). The GPLLJ also affects low-level convergence and nocturnal convective systems (Higgins et al., 1997). Interannual changes in the GPLLJ can be caused by a displacement or a change in strength of the climatological pressure centers, which in turn have been linked to sea-surface temperature (SST) anomalies in the tropical Pacific, to the North Atlantic Oscillation, and to the Asian Monsoon (Rodwell and Hoskins, 2001; Weaver and Nigam, 2008).

Persistent mid-tropospheric ridging has been found to be a dominant meteorological feature during Great Plains droughts in the 1950s and later (when upper-air data are available) (Namias, 1982; Chang and Wallace, 1987). Accompanying subsidence may suppress convection and thus reduce precipitation, which is predominantly convective in the Midwest during the warm season. Various factors such as large-scale oceanic forcing, regional land-atmosphere interaction, and atmospheric dust have been suggested to be involved in reinforcing the “Great Plains ridge”, as suggested by Namias (1982) in his visionary but observationally not well supported work.

The flow in the upper troposphere also plays an important role. Tropical oceanic forcing likely results in changes in the Hadley circulation and upper tropospheric wave trains penetrating into the extratropics. The jet streams provide a waveguide for quasi-stationary waves and affect the propagation of rain-bearing disturbances from the Pacific into the
continent in winter and spring. This might be important because soil moisture in spring has been suggested to affect summertime precipitation in the Great Plains (Oglesby, 1991; van der Schrier et al., 2007) and could also be of interest with respect to predictability.

3.2 Data and Methods

The historical upper-air data analyzed here include wind profiles obtained with pilot balloons (from the NOAA Climate Database Modernization Project, National Climatic Data Center) and temperature and pressure profiles from routine aircraft observations and (starting in 1938) radiosondes (Brönnimann, 2003b; Ewen et al. 2008b). Due to changes in the observing systems in the mid-1940s, we contrast data from the drought period (defined as 1931-1939) with data from adjacent years using the same observing system. The years 1941-1944 were wet in the Great Plains, with a precipitation anomaly almost exactly opposite to that in 1931-1939. The year 1940, which was relatively dry in the Great Plains, is considered a transitional year.

In addition to observations we also present statistical reconstructions of upper-level geopotential height (GPH) fields that are based on a large amount of global historical upper-air data, supplemented with land station temperature and sea-level pressure (SLP) data (Griesser et al., 2008a). We express all fields as anomalies with respect to 1921-1950 to avoid effects of long-term trends. The ERA-40 reanalysis (Uppala et al., 2005) is used as a counterpart in the more recent past.

3.3 Results and discussion

3.3.1 Surface fields

Observed precipitation (GHCNv2) anomalies (Vose et al., 1992) from April to August (AMJJA) for 1931-1939 are shown in Fig. 3.1 together with corresponding SST anomalies (HadISST2) (Rayner et al., 2006), and SLP (HadSLP2) (Allan and Ansell, 2006). Precipitation was strongly reduced in the Great Plains region (but arguably increased over the Gulf of Mexico). As is well known from other studies, the North Pacific was anomalously cold while the North Atlantic was warm. The two dominating high pressure systems, the North Pacific and Azores-Bermuda highs, show slight westward and northeastward shifts, respectively. Without wind data, however, no conclusions can be drawn about atmospheric circulation changes.

3.3.2 The Great Plains Low-Level Jet

Because the GPLLLJ changes during the course of a night (Fig. 3.2 central panel), we show only late night (3:00-6:00 local time) measurements (≥18 profiles must be available per month). Figure 2 shows the averaged AMJJA meridional wind profiles for each year for selected sites. Despite the limited vertical resolution, the low-level jet structure is clearly
visible. During the drought period (red), the meridional component of the GPLLJ was generally weaker than during the early 1940s (blue). However, this does not hold for the core of the jet (e.g., Del Rio, TX). The difference was largest above the altitude of the jet maximum (e.g., Oklahoma City, OK, Wichita, KS) and east of the jet core (e.g., St. Louis, MO). Qualitatively similar results were also found for the early night (21:00-24:00) and for the daytime (but with a weaker jet; results not shown) and hence can be considered robust.

**Fig. 3.1** AMJJA averages of precipitation and SST anomalies (top) as well as SLP and SLP anomalies (middle, hPa) for 1931-1939. Anomalies are with respect to 1921-1950. Bottom: Same as top for dry minus wet years in ERA-40 data.
Wind fields at 1000 and 2000 m asl (Fig. 3.3) further confirm that the main signal is not the weakening of the average jet structure but a change in orientation (stronger westerly component) and northward extent. In this respect the “Dust Bowl” differs from other droughts (e.g., Lyon and Dole, 1995). In fact, an analysis of nocturnal (12 UTC) wind fields for dry minus wet summers in ERA-40 (defined as AMJJA precipitation anomalies in the Great Plains, as in Schubert et al. (2004b), outside ±8 mm/mon) clearly shows a weakening of the GPLLJ in its core region (Fig. 3.3, top right). Other aspects of the “Dust Bowl” are similar to recent droughts, e.g., that the signal is largest above the jet maximum.
The 1930s “Dust Bowl” drought in the US

**Fig. 3.3** Nocturnal winds at two levels averaged for different seasons for drought years (red arrows) and wet years (blue arrows; white arrows indicate the difference). The left two columns show pilot balloon data from 1931-1939 and 1941-1944. The right two columns show a corresponding analysis using ERA-40 data, with difference vectors multiplied by 5. Difference vectors whose $u$ or $v$ component was statistically significant (heteroskedastic t-test, $\alpha=0.05$) are shown in black.

The GPLLJ change between the “Dust Bowl” and the early 1940s shows weak intra-seasonal dependence, with a slightly stronger signal in summer than in spring. In contrast, in the ERA-40 analysis the largest change is found in spring. Furthermore, almost only westerly wind anomalies (compared to 1941-1944) were found during the “Dust Bowl” years, but easterly anomalies in the ERA-40 analysis in the northern regions in spring and
in the southern regions in summer. The differences between the two analyses are largest in summer.

In summary, wind observations show that during the “Dust Bowl” years, the GPLLJ was weaker on its eastward side, shallower, and penetrated less far north in comparison to the early 1940s. Humidity data are not available for the 1930s. However, other studies show that specific humidity, on an interannual scale is less important than wind speed for moisture transport through the GPLLJ (Wang et al. 2008). This suggests weakened moisture transport induced by the weakened GPLLJ during the “Dust Bowl” droughts.

3.3.3 Thermal structure and mid-tropospheric ridging

Since warm season precipitation is primarily convective, we analyzed temperature profiles from nocturnal aircraft and radiosonde ascents from Omaha, NE, the only continuous record in the region, in order to assess the thermal structure of the atmosphere. Observation times changed in several steps from 10 to 4 UTC. We adjusted the data to 6 UTC based on interpolated 6-hourly climatologies from ERA-interim (1989-2005), however, results should be interpreted with care. The two hot and dry summers 1934 and 1936 clearly stand out (Fig. 3.2 top right) with a maximum warming at 1.5 km asl.

Because surface data could not be adjusted, we analyzed daily minimum temperatures at nearby Tekamah. In contrast to the daily maxima, the minima were only slightly higher in 1934 and 1936 (and even lower in the 1931-1939 average) than in 1941-1944. Even though daily minima cannot substitute nocturnal temperature, the contrast to the situation at 1.5 km asl is very large and implies stronger stability in the lowest kilometer and possibly suppressed convection during the night, likely accentuated by decreased low-level humidity (note that stability was decreased above 1.5 km asl).

Large-scale subsidence might have contributed to the thermal structure. The reconstructed 500 hPa GPH fields (Fig. 3.4) show ridging in 1931-1939, with the largest positive anomalies northeast of the Great Plains and additional centers in the Gulf of Alaska and over the Atlantic. Radiative effects of atmospheric dust or feedbacks involving the land surface could have contributed to the ridging and the vertical thermal structure over the Great Plains, but need to be assessed by detailed comparisons with targeted model simulations.
3.3.4 Large-scale upper-tropospheric flow

Our reconstructed 200 hPa GPH fields indicate persistent changes in the upper tropospheric mean flow (Fig. 3.4). The averaged winter field (Nov-Mar preceding the analyzed summer seasons) shows a zonally structured, positive GPH anomaly stretching from the North Pacific into the North Atlantic (with three centers) and a strong negative...
anomaly over northwestern Canada. The effect of tropical Pacific forcing on the extratropics is strongest in winter and is likely reflected in this pattern. In spring, positive anomalies off the coast of California might indicate weaker Pacific jets and possibly fewer disturbances that would otherwise bring in moisture from the Pacific (note that neither reconstructions nor observations allow addressing jet streams and disturbances directly). In summer, a positive GPH anomaly centre is located over the continent. GPH gradients suggest a poleward migration of the polar front jet over the Great Plains, which is consistent with (albeit sparse) upper tropospheric wind observations from the northern plains.

Note that in all seasons, the largest GPH anomaly is found over northeastern Europe. Though an influence on the Great Plains is unlikely, this anomaly might have played a role for the concurrent extremely warm years in the European Arctic.

3.3.5 Comparison with model simulations

The identified characteristic flow features can now be used to assess SST-forced climate model simulations. In addition to published studies, we also refer to our own simulations with the SOCOL model (Schraner et al., 2008; Fischer et al., 2008b; see Auxiliary Material) for illustrative purposes.

Many models, including SOCOL (Fig. 3.5), reproduce the drought in the Midwest in the 1930s in the ensemble mean when forced with observed SSTs and also show increased precipitation over the Caribbean. However, many also produce strong precipitation deficits in Northern Mexico, which was not observed (Seager et al., 2005, 2008; Schubert et al., 2004b; Cook et al., 2008). SOCOL does this in each ensemble member (Fig. 3.5), which could point to a misrepresentation of the regional circulation.

The large-scale circulation response is reproduced reasonably well in the models. 200 hPa GPH patterns in published studies are in good agreement with our reconstructions (Seager et al., 2005; see also Schubert et al., 2004a), thus lending credibility to the models. SOCOL has the band of positive GPH anomalies across the Pacific North American sector too far south in the ensemble mean (Fig. 3.7, left), but the member with the strongest response in Great Plains precipitation also shows the best agreement in 200 hPa GPH (Fig. 3.7, right).

Most studies do not show modeled wind fields, but results from Seager et al. (2005) and SOCOL (Fig. 3.6) agree better with the ERA-40 analysis than with the “Dust Bowl” as they show easterly wind anomalies and suggest a weakening of the jet core. A limitation in reproducing regional circulation changes evidently might affect the simulated spatial pattern of precipitation.
Fig. 3.5 The “Dust Bowl” drought as simulated in the SOCOL model. Anomalies (with respect to 1921-1950) of warm season (Apr-Aug) precipitation in the 9 ensemble members and in the ensemble mean. Hatching in the ensemble mean plot denotes areas where less than 7 out of 9 ensemble members agree on the sign of the anomaly. Ensemble member 4 shows the best correspondence with observed Great Plains precipitation. Note that the color scale is the same as in Fig. 3.1.
Fig. 3.6 Winds fields in SOCOL (daily averages) at two levels in the lower troposphere averaged for different seasons for 1931-1944 (red arrows) and 1921-1950 (black arrows). Difference vectors are plotted in white and are multiplied by 5. The left two columns show the ensemble mean, the right two columns show ensemble member 4.
Fig. 3.7 Anomalies (with respect to 1921-1950) of 200 hPa GPH in SOCOL for different seasons from 1931-1939. Note that the color scale is the same as in Fig. 3.4.
3.4 Discussion and conclusions

The historical upper-air data and reconstructions provide a dynamical view of the “Dust Bowl” droughts (a conceptual depiction is shown in Fig. 3.8), which is necessary to understand its causes. The trigger undoubtedly was oceanic forcing. Only SST-forced model simulations reproduce the drought tendency in the Great Plains as well as the large-scale flow in the upper-troposphere. This suggests that there is predictability (to the extent to which SSTs are predictable) in the likelihood of “Dust Bowl”-like droughts occurring in North America.

![Conceptual depiction of the “Dust Bowl” and its relation to atmospheric circulation.](image)

**Fig. 3.8** Conceptual depiction of the “Dust Bowl” and its relation to atmospheric circulation. Thick coloured boxes denote forcings, the precipitation response is coloured brown, grey boxes denote the new observation-based evidence presented in this work.

What is the relative role of Pacific and Atlantic SSTs? Pacific SSTs affect the upper-level circulation year-round and the GPLLJ in spring and summer while Atlantic SSTs have their largest influence in summer and fall and also affect the GPLLJ (e.g., Schubert et al., 2004b; Wang et al., 2008; Weaver and Nigam, 2008). The upper-level anomalies and the unanimous weakening in the GPLLJ core in spring could therefore reflect Pacific influences, while the summer signal during the “Dust Bowl” (which is distinct from more recent droughts) could point to an additional Atlantic influence. In fact, Caribbean SST anomalies were warmer during the “Dust Bowl” compared to other droughts (Fig. 3.1). The causes for the mismatch in the spatial pattern of precipitation and wind anomalies between the “Dust Bowl”, recent droughts, and model simulations need to be further studied.
The results presented in this paper show that the historical data and reconstructions can serve as a benchmark for climate model evaluation. They allow a refined, process-based assessment of climate models ranging from regional to large-scale circulation responses.

3.5 Acknowledgements

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3.6 Supplementary Material

3.6.1 Simulations with the SOCOL model

In the paper we compare our historical data and reconstructions with published model studies. While the published figures and text allow comparisons in a general sense, the output of these simulations is in many cases not readily available for a more detailed comparison. For illustrative purposes, we therefore also show results from simulations which we performed ourselves and for which we have access to the full output in order to reproduce exactly the same figures as in the paper. Though these simulations were not performed specifically for the analyses of „Dust Bowl“ droughts (which explains the model and set-up used), they nevertheless allow interesting insights.

The simulations were performed with the Chemistry-Climate Model SOCOL (Schraner et al., 2008). We used an ensemble of nine simulations performed in an “all forcings” set-up for the period 1901-1999, i.e., the model was constrained with monthly varying SSTs, sea ice, land-surface conditions, stratospheric aerosols, solar variability, surface concentrations of greenhouse gases and ozone depleting substances, and emissions of short lived species (Fischer et al., 2008b). During the 1930s and the 1921-1950 period (which is used as a reference), there were no strong volcanic eruptions and SSTs are expected to be by far the most important boundary condition. In comparison with other studies, our simulations can be addressed as GOGA (global ocean, global atmosphere) type experiments.

SOCOL is a combination of the middle atmosphere version of ECHAM4 (Manzini and McFarlane, 1998) coupled to the chemistry-transport model MEZON (Egorova et al., 2003). It is a spectral model with T30 horizontal truncation and 39 vertical levels with a model top at 0.01 hPa. SOCOL has been extensively well validated and compared with other models both in the realm of Chemistry-Climate Models and Atmospheric General Circulation Models. In particular, SOCOL was involved in the intercomparison within the framework of the CLIVAR “Climate of the 20th Century (C20C)” Project (Folland et al., 2002). It has shown reasonable performances against selected 20th century climate events (Scaife et al., 2008), as well as monsoon variability (Kucharski et al., 2008; Zhou et al., 2008). SOCOL also showed a good performance of the large-scale circulation when
compared to observations and reconstructions (Brönnimann et al., 2009) and, in particular, a realistic response to El Niño/Southern Oscillation (Fischer et al. 2008a).

### 3.6.2 Results

Figures 3.5 to 3.7 correspond to Figs. 3.1, 3.3, and 3.4 in the paper. We use the same analyses and show the results with the same color scales to allow a one-to-one comparison. The only exception is the wind (Fig. 3.6), which in SOCOL is a daily average and not 03:00-06:00 local time as the observations. The Great Plains Low Level Jet is mainly a nocturnal phenomenon and is much stronger during the night; hence we use a different scale for the wind vector. Also, we compare the winds with the 1921-1950 climatology and not with the 1941-1944 wet period as in the observations. This is due to the fact that the 1941-1944 wet period was rather weak in the model (as compared to the observation where this was a strong event) and hence the difference signal would be small and noisy. Using 1921-1950 gives a more robust signature (though qualitatively very similar as when using 1941-1944) and provides a better consistency with the other figures (Figs. 3.5 and 3.7).

For precipitation we show all ensemble members as well as the ensemble mean. For the other figures we then show only the ensemble mean as well as the ensemble member that is closest to observations with respect to Great Plains precipitation in Fig. 3.5.

In Figure 3.8 we show a conceptual depiction of the anomalous atmospheric dynamics that was responsible for the “Dust Bowl” drought. We mark the observation-based components (boxes) that are new in this work. They allow more targeted model simulations which are necessary to evaluate the arrows in the figures.

### 3.7 Appendix

In the presented paper the drought period 1931 to 1939 in the US Great Plains was compared to the subsequent wet period 1941 to 1944. Additionally the results were compared with climate model runs and with the ERA-40 reanalysis in the recent decades. Since the analysis was based on seasonal and annual means for the period 1931 to 1939 it mainly describes the circulation variability on a decadal scale. However, the 1930s reveal a clear interannual variability. Years with large precipitation deficits were followed by years with wetter conditions.

In this section the focus is on “wet” and “dry” years within the “Dust Bowl” period (1931-1939), although the wet seasons are only slightly wetter than the 20th century average. Four years are analyzed in detailed. 1934 and 1936 were the driest years during the “Dust Bowl” period in the Great Plains and 1935 and 1938 the wettest. For each year the period April through August and November through March (preceding winter) are examined.

Fig. 3.9 and 3.10 depict anomaly fields of different variables for the wet years 1935 and 1938. The precipitation surplus occurred mainly in the period April to August in both years.
1935 and 1938 exhibit above normal temperatures over the central US from November to March, accompanied by a positive geopotential height anomaly (GPH) in the 500 hPa level over the entire US extending into the 200 hPa level. In summer this positive anomaly is shifted to the north-west, right off the cost of Alaska with a negative counterpart in the northern Pacific. Below normal temperatures are dominating over the US during summer. Sea surface temperatures (SSTs) show a year-round warm anomaly in the North Atlantic.
Beside the similarities between these two wet years distinct differences are apparent. From April to August in 1935 the positive precipitation anomaly stretches from the Gulf of Mexico into the US approximately along the low level jet core region. In 1938 during the same period the largest surplus is observed in the north-eastern part of the US while in the low level jet entrance region a precipitation deficit is observed. As summer precipitation is closely linked to the strength and position of the low level jet (Higgins et al., 1997) we expect a weaker low level jet in 1935 than in 1938.

**Fig. 3.10** As Fig. 3.9 but for November 1937 through March 1938 and April 1938 through August 1938.
Fig. 3.11 and 3.12 show anomaly fields for various variables for the dry years 1934 and 1936. Strong negative precipitation anomalies are noticeably from April through August over most of the Central US accompanied by above normal temperatures in the same region. In the northern Atlantic a year-round warm anomaly is visible similar as in the wet years 1935 and 1938.

From November through March colder than normal temperatures prevailed in 1936 over the US whereas the winter 1934 was warmer than normal. The warm anomaly in 1934 is
centred below a positive GPH anomaly in the middle to upper troposphere. In the SLP field a corresponding high pressure anomaly is apparent although the low pressure anomaly over Greenland is relatively stronger. In winter 1936 positive anomalies appear over Greenland and negative anomalies over the north-eastern US and Central Canada. 1934 and 1936 show positive GPH anomalies from April to August over the US with additional centres in the northern Pacific, over the central North Atlantic and over northern Europe. The positive GPH anomaly over the US is strongest in the 200 hPa level.

Fig. 3.12 As Fig. 3.9 but for November 1935 through March 1936 and April 1936 through August 1936.
Analyzing the 500 hPa GPH monthly mean fields in more detail a “blocking” over the Central US in July is apparent for both dry years (Fig. 3.13). The precursor of the blocking is already visible in June over the eastern subtropical Atlantic, propagating upstream. The largest precipitation deficits in the US Great Plains occur together with the strongest blocking (not shown).

**Fig. 3.13** GPH fields on 500 hPa for 1934 and 1936. Presented are May, June, July and August.
While GPH anomalies on the 500 hPa level in April to August show distinct differences among wet and dry years (positive anomalies over the Central US in dry years) the patterns in the 200 hPa level show much smaller differences. All years exhibit positive anomalies in the northern Pacific, over the US, over the central North Atlantic and northern Europe. These centres mainly vary in strength. Striking is the in phase variation of the pressure centres over Europe and the US.

Since oceanic forcing is assumed to play an important role by maintaining drought conditions over the US the observed SSTs are of special interest. In the North Atlantic a warm anomaly in the SSTs is apparent year-round in all examined years. This is in accordance with the findings by Sutton and Hodson (2005) which linked a warm Atlantic Multidecadal Oscillation (AMO) phase with reduced precipitation over the southern US. Additionally the 200 hPa GPH anomaly pattern for the analyzed years is in fairly agreement with the results by Schubert et al. (2004a). They used a climate model forced with SSTs showing that the tropical Pacific and Atlantic have a significant influence upon drought conditions in the US. That points to a potential predictability of droughts in the US.
Atmospheric Circulation Variability related to Droughts in Central Europe since 1890

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Abstract

Drought events occurring in the period 1890 to 2002 affecting Western Europe, especially Switzerland, are analyzed. The drought events are identified using two drought indices, one of which includes information of winter precipitation. Thresholds for drought occurrence were derived yielding 14 and 16 drought events, respectively. Surface and upper-level fields are then analyzed for these droughts. The drought events are investigated addressing the event-non-event differences, inter-event variability and intra-event variability. The drought event 2003 is not included in the event-non-event, inter-event and intra-event analysis but is shown for comparison.

As expected, the analysis reveals strong precipitation, temperature and pressure anomalies accompanying drought events. Below normal precipitation over Western Europe goes together with above normal temperature during summer. A strong positive anomaly in the SLP field over North-western Europe persists from winter through late summer. This anomaly is stronger in winter and significant when taking the winter precipitation into account. No significant winter signal is retained using summer precipitation only. In the upper-level fields a ridge with its centre over north-western to central Europe is visible in
late summer whereas in early summer a low pressure anomaly over the Atlantic is dominating.

Apparent variability between and within the drought events is evident. From December through March the drought events differ in the location of the upper-level ridge and the strength of the subtropical jet stream whereas in the period from June to September mainly the strength of the ridge is varying. This fact is depicted by the four in detail analyzed drought events (1911, 1947, 1972, and 2003) as well. Although all drought events depict a similar pattern from July through September with a positive GPH anomaly in the 500 and 200 hPa level over north-western Europe the precedent evolution is very different. In 1911 and 2003 a positive GPH anomaly is already apparent in the period from December through March with the centre shifted over the Atlantic in 1911, whereas in 1947 the positive GPH anomaly in summer is replaced by a negative anomaly in winter. In contrast in 1972 a positive GPH anomaly in winter was centred over Scandinavia.

The only significant precursors to drought summers are low SSTs in the tropical oceans, most notably the Indian Ocean, pointing to a connection between drought events and oceanic forcing.

4.1 Introduction

Droughts are among the world’s costliest natural disasters. They affect more people than any other natural disaster (Wilhite, 2000). The consequences caused by droughts are widespread and encompass not only agricultural effects but also effects on human health, transport and energy sector. With a better understanding of past droughts we can detect and potentially predict future drought events in an early state of development and possibly mitigate impacts.

As droughts are rare and extreme events, long time series are required for their analysis. Up to now this was only possible for surface variables like precipitation (RR), surface air temperature (SAT), sea level pressure (SLP), and sea surface temperature (SST) (e.g. Della-Marta et al., 2007; Sutton and Hodson, 2005) or products based on these variables like drought indices (e.g Dai et al., 1998) which are available back to the 19th century or even earlier. Addressing upper-air data would be important to address the mechanisms leading to reduced precipitation. However, such studies are restricted to the period covered by reanalysis data sets (ERA-40: Uppala et al., 2005; NCEP-NCAR: Kistler et al., 2001) and consider either a single drought event in the most recent decades (e.g. Black et al., 2004; Black and Sutton, 2006; Fischer et al., 2007b; Schär et al., 2004) or are confined to the second half of the 20th century (e.g. Cassou et al., 2005). With the inclusion of reconstructed upper-level fields back to 1890 (Griesser et al., 2009) we can more than double the period of available upper-air data and hence the number of drought events and assess the circulation variability on a hemispherical scale related to drought occurrence on much more detail.
The primary element of a drought is a precipitation deficit (Lloyd-Hughes and Saunders, 2002). Precipitation is strongly linked to convective and frontal systems and therefore to the meso- and large-scale atmospheric circulation. The precipitation deficit leads to a negative deviation of the water balance compared to the climatological mean. To measure the severity of a drought, the degree of deficiency, its duration and the affected area has to be considered. Apart from precipitation, other variables (e.g. temperature, wind, humidity, evapotranspiration and soil moisture) play a role. For instance decreased evapotranspiration due to an increasing soil moisture deficit can produce a surface heat low and therefore alter the midtropospheric circulation (Fischer et al., 2007b). In some regions this effect might become more important in a future climate (Seneviratne et al., 2006).

Western Europe is a region regularly but not permanently affected by droughts. Over the last century there have been several drought events or even drought-periods (sequences of drought years). Most notably, in the mid 1940s to the early 1950s, there was a period with prolonged droughts (Dai et al., 1998). Summer droughts often coincide with heat waves as it was the case in 2003 (see Black et al., 2004; Black and Sutton, 2006; Fischer et al., 2007b; Schär and Jendritzky, 2004; Schär et al., 2004). Many studies consider heatwaves rather than droughts. To some extend, the findings are also valid for droughts. However, as droughts are longer lasting than heat waves by definition, we expect feedbacks on a different timescale.

Due to the geographical position, Europe is mainly influenced by westerly flow from the Atlantic Ocean year-round. A variety of studies relate the occurrence of droughts and heat waves in Western Europe with an atmospheric blocking situation in the extratropical circulation (e.g. Beniston and Diaz, 2004; Black et al., 2004; Carril et al., 2007; Della-Marta et al., 2007; Fink et al, 2004; Fischer et al., 2007b, Schär et al., 2004). The blocking alters the horizontal advection and has effects on various variables. More clear sky conditions increase the downward shortwave radiation and increase temperature. Additionally the track of the cyclones is shifted to the north and therefore the precipitation over Europe reduced. The anomalous circulation pattern (block) typically remains nearly stationary or moves westward and persists for about a weak. Blocking situations are most frequent during June (Carril et al., 2007). Depending on the analyzed drought events and applied methods the pressure anomaly is centred over central Western Europe (Fischer et al., 2007b) or north-western Europe (Cassou et al., 2005; Folland et al., 2008) or even northern Europe (Della-Marta et al., 2007). Since we are working with monthly data we can not directly address blockings but rather “ridges”.

Some studies found a link between Atlantic SST patterns and the occurrence of droughts or blockings over Europe (e.g. Della-Marta et al., 2007; Cassou et al., 2005; Sutton and Hodson, 2005). Therefore predictability of droughts is expected in that extent as SSTs are predictable. The studies by Della-Marta et al. (2007) and Sutton and Hodson (2005) show a link between the Atlantic SST pattern called Atlantic Multidecadal Oscillation (AMO) and European summer temperature and precipitation. In the work by Folland et al. (2008) the AMO is linked to the Summer North-Atlantic Oscillation (SNAO). The SNAO is defined...
on SLP fields as the first Empirical Orthogonal Function (EOF) in the Extratropical North Atlantic-European region during summer time. The SNAO is a nearly monopole pattern with the strongest anomaly between UK and Scandinavia.

Beside influences from the Atlantic Black and Sutton (2006) assume connections between the Indian Ocean, Pacific and drought events. Descending air, associated with a rossby wave response to the monsoon-heating and the westward propagation has a large influence on summer droughts, especially in summer 2003.

The SST anomalies are linked with the European region via the upper-level circulation. A detailed examination of upper-level fields related to drought events can help to understand how an oceanic signal is transferred to the continents and therefore to Europe.

Apart from the large-scale circulation anomalies regional effects such as land-atmosphere interactions can have a very important role maintaining a drought. Evapotranspiration is an indirect heat storage and therefore can damp the temperature increase. Zampieri et al. (2006) found considerable connections between summer droughts in Europe and precipitation deficits over Southern Europe in the preceding winter.

In this work we will focus on summer drought events in Europe in the period 1890 to 2002 on a monthly to seasonal scale which affected the alpine region, especially Switzerland. The macro-scale circulation variability during drought events is analyzed based on a variety of gridded surface and upper-level variables. At the ground we consider SST, SAT, RR and SLP data. In the upper-levels we focus on geopotential height (GPH) fields. Particular attention is paid to the inter- and intra-event variability of drought events. Composites of drought years contrasted to non-drought years are analyzed to identify dominant circulation modes. Since large variability between and within the drought events is expected, the inter- and intra-event variability is examined using a principal component analysis (PCA). The first two empirical orthogonal functions are retained in the PCA describing either stronger/weaker anomalies or a shift of the anomaly centres. Drought events are defined using two different drought indices (see section 4.2.3).

4.2 Data and Methods

4.2.1 Data

We use the HadCrut3v data set (Brohan et al., 2006) that combines land and marine temperature anomalies fields on a 5° by 5° grid. The anomalies are variance adjusted. As the data set contains missing values only these grid points are retained with at least 90 % available data for the analyzed period. For precipitation (RR) we base on the GHCN V2 data set (Vose et al., 1992), containing monthly values on a 5° by 5° grid. The sea level pressure (SLP) data set HadSLP2 was developed by Allan and Ansell (2006) and is available on a 5° by 5° grid. For the upper-level geopotential height (GPH) anomalies we combine information from two sources. For the period 1957 to 2002 we are using the ERA-40 reanalysis (Uppala et al., 2005), whereas before 1957 we base on the
reconstructions by Griesser et al. (2009). The reanalysis and the reconstructions are available on a 2.5° by 2.5° grid. We are using the 200 hPa and 500 hPa fields for the extratropical Northern Hemisphere. Based on the reduction of error measure (RE) (see Griesser et al, 2009) the reconstruction are considered to be good over Europe, North America and the northern Atlantic, whereas the results over East Asia and the Northern Pacific have to be interpret with care, especially the 200 hPa level. For all data sets anomalies are calculated using the 1961-1990 annual cycle.

4.2.2 Drought indices

To measure the strength or severity of a drought (meteorological, agricultural, hydrological or socioeconomic), many different drought indices have been developed. Depending on the number and selection of variables, they vary in their complexity and effectiveness in describing the different types of droughts.

In our work we use the self-calibrating Palmer Drought Severity Index (scPDSI) (see Van der Schrier et al., 2007) and an adapted version of an agricultural drought index developed by Calanca (2007) (hereafter: CDI) for defining drought events.

The scPDSI is based on the probably most widely used drought index, the Palmer Drought Severity Index (PDSI) (Palmer, 1985). The PDSI was originally developed for the US and was generalized by Wells et al. (2004) and renamed to scPDSI. Van der Schrier et al. (2007) finally applied the index to the European Alpine Region. The scPDSI is based on a simple water-balance model and incorporates precipitation from the antecedent twelve months and moisture supply and demand at the surface. For a comprehensive overview of the calculation procedures the reader is referred to the specific literature (e.g. Wells et al., 2004). For the European Alpine Region, the scPDSI is calculated using the interpolated fields of monthly precipitation and temperature observations published by Efthymiadis et al. (2006). Thus, van der Schrier et al. (2007) developed a scPDSI dataset, which spans the period 1800-2003 and is on a resolution of 10 minutes longitude by 10 minutes latitude, ranging from 4° to 19° east and from 43° to 49° north.

Calanca (2007) has shown that a drought index related to the average soil water availability during the vegetation period of crops is suitable for identifying and gauging agricultural droughts. In Calanca (2007) the drought index was evaluated based on the probabilistic approach to the soil water storage problem developed by Rodriguez-Iturbe et al. (1999). This approach can further be simplified if the evapotranspiration efficiency, that is the ratio of actual to potential evapotranspiration, is used as a proxy for the water stress effects on crop growth (Doorenbos and Kassam, 1979). Budyko (1974) has shown that at the regional and seasonal scale the following holds approximately true:

\[
\frac{ET_{act}}{ET_{pot}} \approx \tanh \left( \frac{Pr_{ec}}{ET_{pot}} \right)
\]
where $ET_{act}$ and $ET_{pot}$ are the seasonal mean (April through September) actual and potential evapotranspiration rates and $Prec$ is the seasonal mean precipitation rate. A corresponding drought index can therefore be defined as:

$$CDI = 1 - \frac{ET_{act}}{ET_{pot}} \approx 1 - \tanh\left(\frac{Prec}{ET_{pot}}\right)$$

For the present application, the drought index was evaluated using homogenized monthly precipitation and temperature data for Bern-Liebefeld ($07^\circ25'\ E$, $46^\circ56\ N$, 565 m.a.s.l.), Zürich SMA ($08^\circ34'\ E$, $47^\circ23'\ N$, 556 m.a.s.l.) and Basel-Binningen ($07^\circ35'\ E$, $47^\circ33'\ N$, 316 m.a.s.l.) (Begert et al., 2005). An estimate of the seasonal mean potential evapotranspiration rate was obtained from the monthly temperature using the Thornthwaite (1948) formula.

The scPDSI and CDI differ with respect to the underlying data and the integrated time period. Whereas the scPDSI takes for each month the precipitation of the antecedent twelve months into account, the CDI considers only the average April to September conditions. Since the two indices are based on different data sets the homogeneity of the underlying data is checked using the Standarized Precipitation Index (SPI) (McKee et al., 1995). SPI is easy to adapt to longer and shorter integration periods and therefore can mimic the behaviour of the CDI and scPDSI index. The SPI is analyzed using an integration period of 12 (SPI$_{12}$) or 6 (SPI$_{6}$) months and is applied to the homogenised station series used for the CDI and to the gridded precipitation data set by Efthymiadis et al. (2006) used for the calculation of the scPDSI.

### 4.2.3 Defining drought events

As drought event we consider a year during which drought conditions prevailed most of the time from April through September or during which severe drought condition occurred for at least one month in the same period. Additionally the drought events should match as close as possible the drought record (with the drought years: 1893, 1904, 1911, 1921, 1934, 1945, 1947, 1949, 1950, 1952, 1959, 1976, and 1983) inferred from historical documents by Schorer (1992 and 2005), Pfister (1999) and Pfister and Rutishauser (2000) for Switzerland.

In terms of the scPDSI and based on our definition of a drought event we consider a year as a drought event if the scPDSI averaged over Switzerland ($6^\circ\ E$ to $10^\circ\ E$ and $46.5^\circ\ N$ to $48.5^\circ\ N$) reaches at least in one month from April to September moderately to extremely dry conditions (scPDSI values $<-2$) or if the whole summer half year experiences dry conditions ($<-1$). In this way we identified 14 drought events based on the scPDSI, namely 1893, 1894, 1895, 1909, 1921, 1929, 1934, 1943, 1944, 1947, 1949, 1950, 1972, and 1976. For the monthly scPDSI values used for the calculations see Fig. 4.1.
Fig. 4.1 Time series of the area mean scPDSI of van den Schrier et al. (2007) (red), the area mean SPI12 from the gridded precipitation data of Efthymiadis et al. (2006) (green), the SPI6 for the month of September from the gridded precipitation data of Efthymiadis et al. (2006) (blue), and the drought index CDI (bars). Years in which the area mean scPDSI satisfied the drought criteria defined in the text (either all monthly values between Apr. and Sep. < -1 or at least one of the values between Apr. and Sep. < -2) are indicated with arrows. To match the scale and sign of the scPDSI and SPI, the CDI was multiplied by -10.
Since the simplified CDI index only provides an approximate metrics of drought, an empirical threshold is introduced to provide similar results as those obtained by Calanca (2007):

$$ CDI \approx \max \left\{ 0, 0.7 - \tanh \left( \frac{\text{Prec}}{\text{ET}_{\text{pot}}} \right) \right\} $$

and the arithmetic mean of the station indices was taken as a representative value for the study region. With the CDI we identified totally 16 drought events, namely 1893, 1895, 1904, 1906, 1911, 1919, 1921, 1929, 1943, 1947, 1949, 1952, 1959, 1962, 1976, and 1992. The CDI values for the drought events are presented in Fig. 4.1.

The drought event 2003 is not included in our analysis since it was characterized by an extreme temperature anomaly accompanied by a strong blocking (Beniston and Diaz, 2004; Fischer et al., 2007; Schär et al., 2004) but only a moderate precipitation deficit. Schär et al. (2004) stated that an event like 2003 will occur only every 46'000 years. If including this single event in the study it is likely that it would dominate the whole analysis.

The drought events are analyzed in terms of the late winter (DJFM), early summer (AMJ) and late summer (JAS) seasonal means. To characterize drought events we partitioned the variability of the analyzed fields into three components: event-non-event difference, inter-event variability and intra-event variability. Inter-event variability is defined in terms of the leading modes of variability for a given season within the sample of drought years. These were determined using a principal component analysis after subtracting the sample mean and standardizing. Intra-event variability was defined on a monthly time scale. From each monthly anomaly pertaining to a drought event we subtracted the mean anomaly from April to September from this specific event. Again the pooled data from all events were analyzed using a PCA. The significance of the difference between event and non-event is estimated using the Wilcoxon Rank-Sum Test for two non-group samples.

### 4.3 Homogeneity analysis

Results of the homogeneity analysis are depicted in Figs. 4.1 and 4.2. As seen in Fig. 4.2, there is a close correspondence between the regional SPI12 estimated from the monthly precipitation at the three stations, and the SPI12 obtained from the gridded precipitation field of Efthymiadis et al. (2006). Based on this finding we conclude that differences in the number of drought years identified by the scPDSI and the CDI are not the results of systematic differences in the underlying data. In addition we note in Fig. 4.1 that the drought index CDI reflects the course of the SPI6 for the month of September, which is a measure of the precipitation anomaly for the months April to September. Differences in the drought sequences inferred from the scPDSI and the CDI are thus effectively caused by the different time scales implicit in the indices (12 months for the former, 6 for the latter). Concerning the severity of drought, the CDI registered the one of 1947 as the most severe.
In fact while the agricultural production only partially suffered from other droughts, an almost complete loss was recorded in 1947 (Schorer, 1992; Pfister, 1999; Pfister and Rutishauser, 2000). The six drought events identified by the scPDSI and not by the CDI are all cases where scPDSI showed very low values already in the early summer (April to May) and therefore can be rather classified as “spring” droughts. The four drought events contained only in the CDI sample are cases with very strong precipitation deficits (not shown) during 1-3 months following a near average period. In these cases the scPDSI with its long memory effect is not able to follow the excursion.

Fig. 4.2 Comparison of the stations based SPI12 with the area mean SPI12 obtained for the region 6-10 E and 46.5-48.5 N from the gridded precipitation data by Efthymiadis et al. (2006). The station based SPI12 was computed as the arithmetic average of the corresponding SPI12 at Bern, Basel and Zurich. The data span the period 1864-2003.

4.4 Results

In Fig. 4.3 event-non-event differences for precipitation are presented. During drought events precipitation is reduced by definition. Generally the largest precipitation deficits are observed over France and therefore west of the region where the drought indices are defined. The strongest anomalies extend into UK and Germany. At the same time the Mediterranean Region experience anomalous wet conditions, especially the south-east which is in agreement with other studies (e.g. Della-Marta et al., 2007). Over Scandinavia a precipitation surplus is registered as well.
Distinct differences are apparent among the drought events identified by the two drought indices. Drought events based on scPDSI (hereafter: scPDSI drought events) reveal a significant precipitation deficit during the winter season which decreases and becomes insignificant in the summer season. For the drought events defined by the CDI (hereafter: CDI drought events) the precipitation deficit is largest in the late summer.

Drought events, independent from the drought index, go together with colder than normal temperatures in Western Europe during winter (Fig. 4.4) and warmer than normal temperatures during summer time. The core of the warm anomaly is centred over France, where the largest precipitation deficit is observed. Noteworthy we detect significant below normal SSTs in the tropical and extra-tropical Atlantic like Della-Marta et al. (2007),
especially from April through September. Depending on the selected drought events the anomalies are strongest in the Caribbean Sea or near the Equator. Beside the SST anomalies apparent in the Atlantic below normal SSTs are measured in the Indian Ocean.

![Fig. 4.4](image-url) As Fig. 4.3 but for temperature.

The SLP event-non-event differences (Fig. 4.5) exhibit a weak to strong positive pressure anomaly centred over north-western Europe similar as the SNAO pattern described by Folland et al. (2008). ScPDSI drought events show a significant positive SLP anomaly during winter time whereas SLP anomalies related to the CDI drought events are strongest in the early summer. For CDI drought events the centre of the strong positive anomaly shifts slightly to the north-west from the early to the late summer. The negative anomaly over the northern Atlantic described by Cassou et al. (2005) is a dominant feature from April through June but completely disappears in the late summer.

The 500 hPa GPH pattern (Fig. 4.6) is nearly congruent to the SLP anomalies. The positive anomaly near the surface over north-western Europe extends into the troposphere. In strength and position the positive anomaly is comparable with the “blocking” situation described by Cassou et al. (2005). Except for the scPDSI drought events during winter the positive anomaly becomes less significant and weakens compared to the negative anomaly.
over southern Greenland. For CDI drought events and late summer a strong negative anomaly over northern Scandinavia is visible.

**Fig. 4.5** As Fig. 4.3 but for sea level pressure (SLP).

In the 200 hPa level (Fig. 4.7) the Northern Hemisphere in winter is dominated by a large negative anomaly over Canada accompanied by weaker positive anomalies over the northern Pacific and over north-western Europe although not significant for CDI drought events. For scPDSI drought events this anomalies become weaker and insignificant in summer whereas for CDI drought events these anomalies strengthen.

Inter-event and intra-event variability of the drought events reveal that the single drought events differ little in the degree of the precipitation deficit over Switzerland but show a similar spatial pattern as the event-non-event difference (Fig. 4.8), especially in winter and early summer. The main difference therefore is that between strong and weak events. In the late summer some drought events spread to the north-west (UK). Within the year (Fig. 4.8:
Intra-event) the dominant mode of variability is a wetting over north-western Europe for the scPDSI drought events (winter drier as summer) or a drying in the same regions for the CDI drought events (summer drier as winter).

**Fig. 4.6** As Fig. 4.3 but for geopotential height on the 500 hPa level.

The SLP and 500 hPa inter-event variability (Fig. 4.9 & 4.10) depicts shifts of the pressure centres. In winter and early summer the first inter-event EOF reveals a north-south displacement while the second EOF exhibits a west-east relocation. In the late summer the drought events differ mainly in the strength of the positive pressure centre over North-western Europe. The intra-event variability (Fig. 4.9 & 4.10, bottom) is dominated by a relocation of the pressure centres. The second intra-event EOF reveals a pattern similar as the Atlantic low apparent in the event-non-event difference.
During winter and early summer all drought events show nearly the same pattern over Europe in the 200 hPa level (Fig. 4.11), though the anomalies vary in strength. The first inter-event EOF for scPDSI and CDI drought events show a clear centre over north-western Europe revealing a stronger or weaker positive pressure anomaly. Furthermore the first inter-event EOF reveals a structure along the sub-tropical jet pointing to a stronger or weaker jet or even a shift of the jet axis. From April through June the displacement of the positive anomaly over Europe is dominant and the structure potentially related to the sub-tropical jet completely disappeared. In the Intra-event variability the relocation of the positive pressure anomaly is the leading mode.

As clear differences appear among the scPDSI and CDI drought events we have a closer look at three distinct drought events. We select 1911, 1947 and 1972 as example cases. 1911 is a drought event based on CDI but not based on scPDSI. The year 1947 probably was the most severe drought observed in Switzerland in the last about 100 years and therefore represents an extreme case. 1972 is only a drought year if we use the scPDSI. For comparison the year 2003 which was not included in our analysis is presented as well.

Fig. 4.12 shows the anomaly fields for the year 1911. As in the event-non-event composites for CDI drought events the precipitation deficit in 1911 over Europe is strongest in the late summer and only moderate in the preceding winter and early summer. Clear above normal temperatures are observed for the late summer. Generally SLP fields depict above normal pressure over Western Europe with the centre in the winter over the Atlantic and in the late summer over north-western Europe. In the upper-levels a strong positive anomaly over Europe in the late summer is visible going together with a low
pressure anomaly over the Atlantic. In the winter this positive anomaly stretches over the Atlantic.

Fig. 4.8 First empirical orthogonal function (EOF) for the inter-event (top six panels) and intra-event variability of the drought events. The drought events are defined either based on the scPDSI (left) or CDI (right). The intra-event variability is calculated on a seasonal scale. The EOF’s are multiplied by the pertaining standard deviations.
Fig. 4.9 As Fig. 4.8 but for sea level pressure (SLP). In contrast to Fig. 4.8 the first two EOF’s are given.
Fig. 4.10 As Fig. 4.8 but for geopotential height (GPH) on the 500 hPa level. In contrast to Fig. 4.8 the first two EOF’s are given.
Fig. 4.11 As Fig. 4.8 but for geopotential height (GPH) on the 200 hPa level. In contrast to Fig. 4.8 the first two EOF’s are given.
Fig. 4.12 The drought event 1911 resolved on a seasonal scale. From left to right: winter, early summer and late summer. All fields show anomalies. From top to bottom: geopotential height (GPH) 200 hPa, GPH 500 hPa, sea level pressure (SLP), temperature and precipitation. The anomalies refer to the period 1961 to 1990.

In the winter 1947 (Fig. 4.13) the negative precipitation anomaly is located over France and stretches north into Scandinavia. The negative precipitation anomaly extends in the early summer over whole Europe. From the early to late summer Western Europe experiences above normal temperatures. The SLP field is dominated by a strong positive anomaly over Scandinavia during winter and an opposite anomaly over the Atlantic. The pattern changes completely in summer, except the anomaly over the Atlantic. The strong positive anomaly is replaced by a weaker but broader centre over Central Europe shifting to north-western Europe in late summer. The 200 and 500 hPa level show a comparable pattern as the SLP field.

1972 is characterized by a strong precipitation deficit in winter over Western Europe and by moderate to strong deficits over Central Europe. These anomalies go together with a positive pressure anomaly over Scandinavia apparent in the SLP, 200 hPa and 500 hPa field. During early summer normal conditions prevailed with respect to precipitation and
temperature. In late summer a positive pressure anomaly builds up over north-western Europe. This pressure anomaly becomes weaker in the upper troposphere. In the higher troposphere a positive pressure anomaly over Scandinavia is dominating.

Fig. 4.13 As Fig. 4.12 but for the drought event in the year 1947.

The year 2003 was not included in the analysis but is shown for comparison (Fig. 4.15). December 2002 through March 2003 show only moderate precipitation deficits with its centre south of the Alps. Whereas the negative precipitation anomaly from April through June was strongest in the regions adjacent to the Mediterranean Sea and parts of Eastern Europe, the anomaly shifted north-west in the late summer with its centre over Germany. The precipitation anomaly in July through September was accompanied by an extreme positive temperature anomaly centred over France. The circulation was dominated by a mid-tropospheric block apparent from December through August. Since the circulation considerably changed in September the blocking is not well represented in the July to September mean.
4.5 Discussion

Significant circulation anomalies are found related to drought events in Western Europe. In late summer a ridge in the 500 hPa level was dominating apparent in the 200 hPa and SLP field as well. From April through June a low pressure system over the Atlantic is visible. Cassou et al. (2005) and Black and Sutton (2006) linked these pressure anomalies and therefore circulation anomalies with anomalous SSTs in the Atlantic and the Indian Ocean, respectively. Our analysis supports both studies as we found significant SST anomalies in the tropical and sub-tropical Atlantic and in the Indian Ocean. The signal is strongest in the sub-tropical Atlantic from April through June. If the scPDSI index was used for the drought event detection and therefore an index taking into account the precipitation from the antecedent twelve months a strong SST signal was found in the Indian Ocean from December through March. A second weaker anomaly is apparent in the Caribbean Sea and west of Africa. Using the CDI index, integrating April to September precipitation, reveals only a weak significant SST anomaly in the Indian Ocean in the preceding winter. Therefore we assume that the winter signal is transferred into the summer by a soil moisture memory effect and not by an “atmospheric memory” effect (quasi-stationary
European Droughts

circulation anomalies). The inter-event variability confirms this assumption since between the single events more variability is visible from December through March than from July through September, with respect to the location of the pressure centres. The detailed analyzed drought events (1911, 1947, 1972 and 2003) support these findings since they show distinct different pressure patterns in winter but similar patterns in late summer.

**Fig. 4.15** As Fig. 4.12 but for the drought event in the year 2003.

As we found comparable results as Cassou et al. (2005) and Black and Sutton (2006) but for a longer time period and therefore for a larger number of drought events we assume that European drought events are associated with a rossby wave response to anomalous SSTs in the Indian Ocean and the tropical and sub-tropical Atlantic. The importance of each region has to be assessed using a SST forced climate model and is beyond the scope of this paper.

### 4.6 Conclusion

Based on two drought indices (CDI and scPDSI) 14 respectively 16 summer drought events for the period 1890 to 2002 have been identified. 10 drought events are detected by
both drought indices. The difference in the detected drought events is due to a longer or shorter memory effect implicit in the drought indices. Drought events identified by the scPDSI represent a collection of summer and spring droughts whereas the CDI clearly detects summer droughts. Therefore distinct differences among the analyzed fields are apparent depending on the drought index.

The identified drought events reveal clear anomalies with respect to precipitation, temperature and pressure. A clear negative precipitation anomaly over Western Europe is apparent. If we include spring drought events negative precipitation anomalies occur in the preceding winter while analyzing only summer drought events (CDI) the winter precipitation deficit is not significant. Drought events are connected to a ridging over north-west Europe. The positive pressure anomaly is apparent throughout the troposphere. These findings are in agreement with the work by Cassou et al. (2005) with respect to the ridging and show a comparable SLP pattern as described by Folland et al. (2008) for the SNAO. Among the drought events obvious differences in the location and the strength of the pressure centre are visible. The pressure and precipitation anomalies go together with above normal temperatures during drought events. This is potentially an effect of the radiative forcing induced by more clear sky conditions. Significant colder than normal SST are observed in the tropical and sub-tropical Atlantic and in the Indian Ocean during drought events. In the preceding winters SST anomalies are much weaker and only significant if we include precipitation from the antecedent winter. The results are in accordance with Della-Marta et al. (2007) and Cassou et al. (2005) which showed a link between Atlantic SST’s and European heat waves. The analysis of single drought events (1911, 1947 and 1972) supports the general findings. 1911 and 1947 are clear summer drought events with a main precipitation deficit over Europe during summer and weaker deficits in the preceding winters. Both drought events show a positive pressure anomaly over north-western Europe, strongest in the late summer. This pressure anomaly extends throughout the troposphere. In contrast 1972 is a spring drought event with the largest precipitation anomalies during winter accompanied by a ridging over northern Europe. During summer a precipitation surplus is detected.

4.7 Acknowledgement

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Conclusions and Outlook

This thesis investigated the upper-level circulation anomalies related with droughts in Europe and the US with a special focus on the “Dust Bowl” drought over the US Great Plains. For this purpose a global upper-level dataset for temperature and geopotential height back to 1880 was statistically reconstructed based on historical upper-air data digitized in the last years and additional surface measurements. With a length of 78 years the data set more than doubled the upper-air record of a comparable resolution. It is shown that reconstructions provide useful insights into atmospheric circulation variability related with extreme climate events. For the climate modeller community the data set provides a useful tool to validate their model results for droughts. Instead of comparing modelled and observed precipitation deficits the circulation anomalies can be addressed directly by comparing oceanic forced upper-level circulation anomalies and reconstructed fields.

The reconstruction comprises several thousands upper-air and surface predictors for the period 1880 to 1957. As the data coverage is changing over time and most predictors are available over the continents in the Northern Hemisphere a complex weighting scheme was developed to retain the full information present in the data set. As predictand temperature and geopotential height fields from the ERA-40 reanalysis (Uppala et al, 2005) were used. Between predictand and predictor data a multiple linear regression model combined with a principal component analysis was fitted making optimal usage of the available data. The quality of the reconstruction was tested in detail based on a split-sample approach and with independent upper-air data in the historical period. Based on our validation results the reconstructions are considered to be good in the Northern Hemisphere with excellent skill over Europe and the US. Geopotential height is generally better reconstructed than temperature and lower levels are better reproduced than higher. Intermediate reconstructions are expected in the Tropics with increasing skill in the higher troposphere. In the Southern Hemisphere the skill is low except over Australia. This is probably due to the poor data coverage and the erroneous reanalysis. In addition the split-sample validation reveals better reconstructions during hemispheric winter. The usefulness of the reconstructed fields was demonstrated by analyzing extreme climate events over the US.

In a first application the reconstructed upper-level fields were used to examine the 1930s “Dust Bowl” drought over the US Great Plains. In the 1930s reliable reconstructions are expected as many upper-air predictors are available. To resolve the Great Plain low level jet in the boundary layer historical pilot balloon soundings were added to the analysis and the results were compared to climate model output. During the “Dust Bowl” period the
Great Plains low level jet was diverted to the east and was weaker especially in the eastern regions above the jet maximum. In the middle troposphere the weaker and diverted jet was accompanied by a ridge which suppressed convection and therefore the main cause for precipitation in the Great Plains. In the upper troposphere a zonal pattern was found stretching from the Pacific over the US into the Atlantic. In summer a positive geopotential height anomaly in the 200 hPa level is found. The position of the largest gradient is slightly shifted to the north suggesting a poleward migration of the polar-front jet.

In a second application the reconstructed fields were analyzed with a focus on European drought events in the period 1890 to 2002. Based on two different drought indices, drought events in Western Europe affecting Switzerland are defined. As the drought indices differ in the implicit memory effect, they identified different drought events. A set of 14 and 16 drought events was denoted, respectively, whereas 10 drought events pertain to both drought samples. Beside the upper-level geopotential height and temperature fields various climate variables were investigated. The different variables were analyzed with a special focus on the event-non-event, inter-event and intra-event variability. The detected drought events clearly show a precipitation deficit over Central Western Europe with the core region over France. Spring drought events show the largest precipitation anomalies in winter, whereas for summer drought events the deficit is most apparent in late summer. The drought events are accompanied by anomalous high pressure near the ground and in the middle and upper troposphere. This pressure anomaly (ridge) is mainly centred over North-Western Europe. In spring a significant low pressure anomaly over the Atlantic is visible. Significant relations were found between drought events and cold sea surface temperatures in the Tropical and Subtropical Atlantic. In winter and early summer the drought events mainly differ in the position of the ridge whereas in summer the strength of the anomaly is dominating.

Based on various studies and according to our results, drought events in Europe and the US are triggered or at least intensified by oceanic forcing. The upper-tropospheric zonal pattern, reconstructed for the 1930s “Dust Bowl” drought over the Central US, is in overall agreement with the study by Schubert et al. (2004b). Hence it is likely that the implied tropical SST forcing is true, but that it is probably modulated by an extratropical influence affecting the upper-level circulation and therefore the propagation of the tropical signal. Additionally it is not clear if the observed dust storms had significant influences on the radiation budget and therefore on the radiative forcing.

European drought events could be linked to anomalous SSTs in the tropical Atlantic, especially the Caribbean Sea, and the Indian Ocean. A weaker but clear signal was found in the Mediterranean Sea, probably an upstream effect of an Indian Ocean SST forcing or a direct effect of the droughts themselves. The oceanic influence becomes apparent in a ridge over North-Western Europe during summer and an Atlantic low pressure system in spring (April-June). Furthermore the “Atlantic low pattern” is a dominating mode in the intra-event variability and seems to play a crucial role in the development of drought events. Cassou et al. (2005) link the Atlantic low with anomalous SSTs in the Caribbean Sea and
Black and Sutton (2006) make the connection between the ridge in summer and the Indian Ocean. Therefore we have to assume that the tropical Atlantic mainly controls the early stage of a drought and that during summer the influence of the Indian Ocean dominates. As SSTs are predictable to some extent on the interseasonal to the interannual timescale there is hope that future droughts could be detected up to one year in advance. For this purpose the reconstruction method could be adapted. With lagged predictors or/and forecasted SSTs, circulation anomalies related with drought events could be identified before a drought develops.

For future reconstructions different aspects have to be considered to further improve the quality of the reconstructed fields. Basically the quality of reconstructions depends on the availability of predictors of good quality. Therefore a special effort has to be accomplished to locate additional upper-air records in data sparse regions especially in the Southern Hemisphere and in the Tropics. Additionally the predictand (ERA-40; Uppala et al., 2005) depict some deficit mainly in the Southern Hemisphere or over the Himalaya. To detect and potentially eliminate these inhomogeneities will help to make maximum profit of the few predictor data. In regions with an already very dense station network like the US or Europe the statistical model can be further improved. The current statistical model with its underlying principal component analysis is designed to capture the large scale circulation patterns. Small scale variations are not resolved by definition. The model could be divided into a large scale part capturing the global circulation patterns and a small scale part focusing on regional to local scale circulation. For the small scale part a new model has to be developed linking single predictand grid points directly with an optimal subset of predictor records in the neighbourhood. With increasing complexity of the statistical model computational power considerations become more important. The current model runs on three PCs for about three months.

For in depth evaluation, the statistical regression approach should be tested in a climate model environment. With time series of several hundred years questions could be addressed like the optimal length of a validation period or if the statistical relation depicted by the regression is stable over time. Since the reconstruction method was designed to capture continental to global scale circulation anomalies the reconstructions should not be used to address the variability at a single grid point or to investigate circulation indices defined as differences between two locations. The reconstructed fields should mainly be used to analyze the displacement or strengthening/weakening of quasi-stationary pressure anomalies. With respect to the analyzed climate extremes it would be of advantage to extend the upper-level reconstructions back in time based on surface data. As droughts are climate extremes and therefore rare events long time series are required for their analysis. To include more drought events would potentially lead to a consolidation of the findings. In the US several severe droughts occurred in the time before 1880, e.g. before the reconstructions start.
Appendix A

Climate effects of the 1883 Krakatoa eruption: Historical and present perspectives

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Abstract

The climatic impact of the Krakatoa eruption in 1883 was intensively studied by scientists at that time and had lasting effects both in volcanology and atmospheric sciences. The theoretical concept of enhanced volcanic aerosol concentration, blocking short-wave radiation and possibly cooling the Earth’s surface was first formulated after the eruption. Later studies addressed the relation between volcanic eruptions and zonal circulation in the midlatitudes. A century later, the Pinatubo eruption (in 1991) played a similar role for climate science, demonstrating the importance of stratospheric processes and their coupling with climate near the ground. Here we revisit the Krakatoa eruption from a present-day perspective. Using reconstructed upper-level circulation fields we find that the Krakatoa effects in the first winter after the eruption fit well with the currently accepted mechanism. The data suggest a strengthened polar vortex in the Arctic stratosphere, while Europe experienced a warm winter at the Earth’s surface. The second winter does not show this signal anymore, calling for a „life cycle“ view of volcanic effects. Results are important also in the context of the current debate on „geoengineering“ of the global climate system.
A.1 Introduction

Major volcanic eruptions affect climate on a large scale and represent the most prominent natural climate forcing on interannual to decadal time scales. The global cooling induced by large eruptions (Robock, 2000) might temporarily counterbalance global warming trends. In fact, geoengineering methods mimicking a permanent volcanic eruption have been suggested to counter the anthropogenic greenhouse effect (Crutzen, 2006). However, the volcanic effect is more complex than a mere global cooling. Volcanic eruptions alter the atmospheric circulation, the hydrological cycle, stratospheric chemistry, and ocean heat content in a complex way for months to years to even decades after the eruption (Robock, 2000; Gleckler et al., 2006). Studying climatic impacts of volcanic eruptions therefore forces us to look at the global climate as part of the Earth’s system (which is an important perspective also for assessing geoengineering methods) and promotes our understanding of this system. It also has a practical aspect: Should an eruption occur in the near future, climatic effects might be predicted.

The current knowledge on atmospheric effects of major eruptions is to a significant extent based on the eruption of Mount Pinatubo in June 1991. Despite the wealth of information (including satellite data) and the large number of studies, the question remains how representative the eruption was and how this compares with previous large eruptions, such as the one of Krakatoa in 1883.

Contrasting the eruptions of Mount Pinatubo 1991 and Krakatoa 1883 (the strongest eruption of the past 150 years) is appropriate because of their similarities: both volcanoes are located near the Equator and both ejected similar amounts of ashes and gases into the atmosphere. Observation-based data sets to study climate effects of the 1883 Krakatoa eruption exist, and a new upper-level data set covering this period has recently been finalised (Griesser et al., 2009). Another point makes the comparison of the two eruptions particularly interesting: Their role in our scientific understanding of volcanic effects. A large amount of contemporary literature, summarising various kinds of observations, is available for the Krakatoa eruption. For instance, colourful sunsets were observed all over the world during almost three years after the eruption (Zerefos et al., 2007). These observations and their interpretations were not only important for the history of atmospheric sciences (Symons, 1888): the works of geologists on the remaining of the Krakatoa volcanic complex built the foundations of the modern volcanology (Verbeek, 1884).

In this paper we analyse the climatic effects of the 1883 Krakatoa eruption from a historical and present day perspective. We start with a short description on the volcanic eruption, a summary of what science has learnt from Krakatoa in the 1880s and how the discussion has evolved since. Then we analyse the Krakatoa eruption in a newly available dataset of upper-level fields that were specifically reconstructed for this time period. This serves as a test whether a strengthening of the stratospheric polar vortex due to increased volcanic aerosols (as hypothesised in studies on Pinatubo) can be observed in the
reconstructions. We end with brief conclusions on open issues, which concern the life-cycle of the volcanic climate effect.

A.2 The 1883 Krakatoa eruption

In 1883, Krakatoa was an unpopulated island composed of three volcanoes, Danan, Perbuatan and Rakata. It was located in the strait of Sunda (6°6’ S, 105°25’ E), between the islands of Java and Sumatra (see Fig. A.1). Both islands belong to the arc of Sunda, where the Australian plate subducts under the lighter Eurasian plate. This region of the Earth undergoes an important tectonic activity and Indonesia experienced several major volcanic eruptions in its relatively short geological history: Toba (~71000 years ago), Tambora (1815), Krakatoa (1883) and Agung (1963) are the best-known. These eruptions all had widespread climatic effects: the Toba eruption, with an estimated decrease of global temperature of about 3-3.5°C likely affected human evolution (Ambrose, 1998). The famous “year without summer” in 1816 is associated with the Tambora eruption.

Fig. A.1 Map of the Sunda Strait after the eruption of Krakatoa (Hurlbut and Verbeek, 1887).

Three factors contribute to the large climate effect of volcanoes of the arc of Sunda. First, the volcanism associated with subduction zones is a very explosive one, allowing the ejected gases to reach the stratosphere. Secondly, the andesitic to dacitic lavas produced in the subduction zones are enriched in sulphur (in the form of SO$_2$ and H$_2$S), which in the stratosphere react to form sulphate aerosols. Thirdly, because the arc of Sunda is located near the equator, the stratospheric aerosols have a long life-time with respect to transport. In fact, the stratospheric aerosol cloud, which is zonally distributed around the whole globe within a few weeks, can spread meridionally over both hemispheres and remain in the stratosphere for 1-3 years before being transported to the troposphere and washed out.
After over 2000 years of silence, the Perbuatan volcano erupted in 1680 leading to devastation of the island’s vegetation. On the 20th of May 1883 Perbuatan showed again signs of activity: an 11 km long ash and water vapour column could be seen from Java and Sumatra. During the three following months small eruptions accompanied by earthquakes occurred from time to time. Surprisingly, this bothered neither the Indonesian nor the Dutch colonists: they even organised trips to Krakatoa Island to have a picnic beside the erupting volcano. Albeit the people remembered the destroying Tambora eruption in 1815 and realised similarities, the Perbuatan volcano was considered too small to be dangerous. Only later it was recognised that the three volcanoes were actually three volcanic cones related to the same huge magma chamber. The main eruption occurred on the 26th and the 27th of August 1883 (Fig. A.2). The lava and gases were ejected from both the Perbuatan and the Danan volcano to approximately 50 km height. The magma chamber emptied itself very quickly and collapsed, such that only half of the Rakata volcano remained. The coasts of Java and Sumatra experienced severe damages due to tsunamis and pyroclastic flows; more than 30’000 people died during the Krakatoa eruption.

![Fig. A.2](image)

*Fig. A.2* Chronology of the Krakatoa eruption during its most active phase on 26 and 27 August 1883. In the beginning the eruption was plinian, later Ignimbrit-forming (Self and Rampino, 1981). Reprinted by permission from Macmillan Publishers Ltd: Nature, © 1981.

The 1883 Krakatoa eruption was important also in terms of perception (of volcanic eruptions and of geosciences) as it was the first eruption that found a media echo all over the world. Telegraphy allowed up-to-date news coverage, and the aerosol cloud fashioned colourful sunsets and moonlight worldwide during more than two years.

### A.3 Historical literature

Rogier D. M. Verbeek was a Dutch geologist doing survey in Indonesia at the time of the Krakatoa eruption and was also the first man to dock on the remains of the Krakatoa Island a few months after it. In his first paper about the eruption (Verbeek, 1884), he was already able to give detailed scientific information on the ejected materials, and compared it with the Tambora eruption. The volume of material ejected by the Krakatoa volcano was about
20 km³, around ten times less than the Tambora eruption. The darkness caused by the ash rain persisted for three days after the onset of the Tambora eruption, in contrast to only a few hours after the Krakatoa eruption. Verbeek further wrote that the ejected gases and ashes were pushed into the upper-air currents, in 15-20 km height, where they would freeze and then travel as tiny crystals around the globe. He thought that these crystals were the cause of the colourful sunsets, since particles with the same composition as the Krakatoa’s ashes were found in the falling snow in Spain.

The eruption had a “peculiar” effect on atmospheric pressure, showing a shock wave travelling around the globe. Corresponding papers were presented in front of the Royal Society and led to the creation of a special committee to investigate causes and effects of the eruption by compiling independent observations, comments and hypothesis from scientists from all around the world (see: The Royal Society, Collection of the Month - Krakatoa Committee, January 2006, http://royalsociety.org/page.asp?id=3987). The report, published in 1888 (Symons, 1888), also discussed atmospheric effects. Around half of the observers did not attribute the observed effects (such as colourful sunsets) to the eruption; others developed inventive theories on the phenomena. Interestingly, one observer mentioned the possibility of sulphuric acid being responsible, but this hypothesis was abandoned until the 1970s. Another interesting analysis concerned the propagation of the ash cloud which circled the globe from east to west within about two weeks (see Fig. A.3). Though the stratosphere as an isothermal or inversion layer had not yet been discovered, the notion that strong (>30 m/s) upper-level winds existed began to form, which became known as “Krakatoa easterlies” (Hastenrath, 2007). From hindsight, we can interpret these winds as an easterly phase of the Quasi Biennial Oscillation (QBO). Interestingly, the report mentions that ashes travelled from east to west also after the Tambora eruption, thus pointing to an easterly phase also during this eruption.

![Fig. A.3 Observed extent of ash cloud after the Krakatoa eruption. Numbers give the dates of observations (26 August to 9 September 1883) (from Symmons, 1888).](image)

The interest in understanding the atmospheric effects of volcanic eruptions (including Krakatoa) was further boosted after the eruption of Mount Katmai in Alaska in June 1912. Here, in particular, the change in solar radiation was analysed. Abbot and Fowle (1913)
compiled solar radiation measurement and published time series of the solar radiation, sunspot numbers and temperature back to 1880 (see Fig. A.4), thus including the Krakatoa eruption. They calculated that solar radiation decreased by around 10% after the eruption (Abbot and Fowle, 1913). A few months later, Humphreys (1913) related the ice ages to volcanic eruptions in a fundamental paper. It was already known at that time that the atmosphere can be subdivided into troposphere and stratosphere and that the volcanic particles (even if their nature was still unclear) reach the stratosphere. It was also well accepted, that volcanic eruptions tend to cool the global climate because of the absorption of the direct solar radiation by the particles. Based on the measurements of Abbot and Fowle, Humphreys calculated that a global decrease of the solar constant of 10% would correspond to a global cooling of about 6.4 °C (Humphreys, 1913), which is quite an overestimation.

Fig. A.4 Solar radiation, sunspot number, and temperature from 1880 to 1910. A. Observed and smoothed annual mean noon solar radiation. B. Wolf's smoothed sun-spots numbers. C. Combined solar radiation and sun-spot numbers. D. Smoothed annual mean departures, United States maximum temperatures (15 stations). E. Smoothed annual mean departures, world temperature (47 stations) (from Abbot and Fowle, 1913).

Only later it was recognised that volcanic eruptions can also alter the atmospheric circulation. In a pioneering article on atmospheric circulation over the North Atlantic, Defant (1924) defined an index of the zonal flow (similar to the currently used North-Atlantic Oscillation index) which he compared with time series of volcanic eruptions. His analysis revealed that during the winter 1884, following the Krakatoa eruption, the atmospheric circulation was strengthened (corresponding to a positive NAO phase) and the temperatures were exceptionally warm over Europe. A few years later, Ångström (1935) summarized this in an apt statement:

"Thus a change in the solar radiation reaching the lower air layers, - it may be produced by a change of the solar constant or by a variation of the transmission of the atmosphere – must naturally be expected to produce an increase in the temperature contrast between equator and pole and in this connection also an increase in the atmospheric circulation." (Ångström, 1935)
He estimated the global cooling caused by volcanic eruptions to be of the order of 0.5-1 °C, which is more realistic than the numbers suggested by Abbot and Fowle (1913) and Humphreys (1913).

The interest in the volcanic effect on climate was revived again from the 1970s on (Lamb, 1970), in the context of historical climatology, the eruptions of Mt. Agung and later El Chichón and Pinatubo, and in the 1980s discussion about the “nuclear winter”. The Krakatao eruption was dealt with in a book by Simkin and Fiske (1983). The latest re-emergence of interest is due to geoengineering proposals such as the one by Crutzen (2006), which aim at producing a perpetual volcanic eruption by artificially injecting sulphur into the tropical stratosphere.

A.4 The current view of volcanic effects on climate

The present day knowledge of atmospheric processes related to volcanic eruptions is summarised in Robock (2000) and was significantly shaped by the eruption of Pinatubo in 1991. It is now understood that the blocking of short-wave radiation reaching the ground (leading to a cooling) makes up only part of the volcanic signal. Volcanic aerosols in the stratosphere also absorb radiation and thus heat up the stratosphere. This leads to differential heating between the (sunlit) tropical-to-midlatitude stratosphere, which after tropical eruptions is rich in aerosols, and the (dark) polar stratosphere during winter. The altered temperature gradient acts to strengthen the polar vortex in the stratosphere. Through mechanisms of downward propagation (Baldwin and Dunkerton, 2001), it is believed that the strengthened vortex also strengthens the zonal circulation in the troposphere over the North Atlantic (as found by Defant, 1924) and leads to mild winters in Europe. This was also confirmed by means of historical climate reconstructions (Fischer et al., 2007a). However, the mechanisms are still not fully clear. For instance, model studies with a simplified volcanic forcing (reducing total solar irradiance) have produced a similar climate signal over Europe (Yoshimori et al., 2005) without invoking stratospheric heating. In turn, recent model studies with more realistic forcing do not reproduce the main features very well (Stechnikov et al., 2006). One reason for this might be the apparent coincidence of El Niño events and volcanic eruptions (see also Brönnimann et al., 2007), another reason might be the lack of comparability of different volcanic eruptions. Therefore, for a better assessment of the effects, case studies of strong eruptions are necessary.

A.5 Data and methods

For the following analyses we use the sea-level pressure (SLP) data set HadSLP2 (Allan and Ansell, 2006), the surface-air temperature data set HadCRUT3v (Brohan et al., 2006) as well as reconstructions of upper-level circulation (Griesser et al., 2009). The reconstructions include temperature and geopotential height up to 100 hPa (i.e., the lower stratosphere) for the extratropical northern hemisphere back to 1880. They are based on
principal component regression and were calibrated and validated in ERA-40 reanalysis data (Uppala et al., 2005) in a similar way as described in Brönnimann and Luterbacher (2004). For the period considered here, the reconstructions are based only on data from the Earth’s surface (station temperature series and SLP fields). Hence, they represent sophisticated interpretations of surface data rather than original upper-level information. Still, they provide useful information as they indicate which state of the stratosphere is consistent with the surface data. As in Brönnimann and Luterbacher (2004), low reconstruction skill (reduction of error <0.2 determined in split sample validations) is indicated in the figures by grey shading.

We analysed winter (Dec. to Mar., labelled with the year starting in Jan.) and summer (Jun.-Sep.) of the years 1884 and 1885, i.e., one and two years after the eruption, respectively. All fields are presented as anomalies with respect to the period 1880-1909, but after removing other volcanically perturbed winters as well as strong El Niño and La Niña events according to Brönnimann et al. (2007). The remaining winters were: 1880-83, 1888, 1894-96, 1898-99, 1901-02, and 1905-09, the summers: 1880-83, 1887, 1893-95, 1897-98, 1900-01, and 1905-08. Note that due to the shortness of the reference period, no missing values were allowed, which reduces the temperature field considerably.

A.6 Results and Discussion

Figure A.5 shows the fields of SLP and 100 hPa geopotential height for the two winters following the eruption. The SLP field for the winter of 1884 exhibits a pattern resembling the positive mode of the North Atlantic Oscillation. Icelandic low and Azores high were well developed, leading to a strengthened zonal circulation over the Atlantic. Over the Pacific, in contrast, the Aleutian low was weaker than normal. The stratospheric reconstructions point to a strong polar vortex, with a shift of the vortex centre towards Greenland. Surface air temperatures (Fig. A.6) show a pronounced winter warming over Europe during the first winter and low temperatures in North America. The second winter was different in many respects. Over the North Pacific, the Aleutian low was strengthened, which could have been due to a moderate El Niño event. The SLP anomaly in the Arctic basin was similar as in the previous winter, but the pressure distribution over the North Atlantic was different, with pronounced negative anomalies over the Bay of Biscay and near-neutral conditions with respect to the North Atlantic Oscillation. Temperatures in Europe were above climatological values, but by much less than the year before, while North America was still cooler than in the reference climatology. Surface air temperature anomalies in summer show a slight cooling in both years.

The pattern found in observations and reconstructions during the first winter after the Krakatoa eruption is very similar to the one observed after the Pinatubo eruption and hence supports a volcanic winter effect via the stratosphere. The aerosols likely strengthened the meridional temperature gradient in the stratosphere, giving rise to a strong vortex, which via downward propagation might have acted on the tropospheric circulation and caused a
strengthened zonal flow over the North Atlantic and a winter warming of the European continent. Similarly, surface air temperature in summer exhibits a slight cooling, as is expected from the reduced short-wave radiation. The second year, however, shows a different pattern. We find no sign of the volcanic effect described above (although European winter temperatures remain high; see also Fischer et al., 2007a, who found significant volcanic effects also in the second winter after an eruption). There are at least two possible causes: The volcanic effect decreases in strength and other processes (i.e., variability) take over, or the volcanic effect itself changes as it undergoes a life cycle. The former might also be related to a weaker than commonly expected Krakatoa forcing. Its lavas were much poorer in sulphur (150 ppmv) than those of Tambora (380 ppmv) and Agung (800 ppmv) (Rampino and Self, 1982). The latter might be related to the stratospheric aerosol distribution, which certainly changed from the first to the second winter, or an oceanic response (including possible volcano-triggered El Niño conditions). In any case, if this is a life-cycle effect, then it is important with respect to geo-engineering projects.

Fig. A.5 Anomalies (with respect to a climatology, see text) of sea-level pressure (top) and 100 hPa geopotential height (bottom) during the winters (Dec.-Mar.) of 1884 (left) and 1885 (right). Shaded areas denote low reconstruction skill.
The major tropical volcanic eruptions Krakatoa and Pinatubo have both considerably advanced our understanding of the climate system. After the Krakatoa eruption (1883), radiative properties of volcanic aerosols, blocking short-wave radiation and possibly cooling the Earth’s surface were established. First ideas on effects on the midlatitude zonal circulation, published 40 years later, also had their seeds partly in the Krakatoa eruption. A century later, the Pinatubo eruption (in 1991) played a similar role for climate science, demonstrating the importance of stratospheric processes and their coupling with climate near the ground. Revisiting the Krakatoa eruption using newly available stratospheric reconstructions confirms that Pinatubo might be, after all, quite representative for strong tropical eruptions. This is especially true for the first year of the eruption and might help to understand past climate variability, to assess climate models, or to predict climate following future volcanic eruptions. However, further studies are necessary to more fully analyse the possible “life cycle” of volcanic effects (already hinted at in historical work, see Defant, 1924). This is particularly important with respect to geoengineering applications.

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