Revisiting the observed climate surface response to large volcanic eruptions

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Abstract. In light of the range in presently available observational, reanalysis and model data, we revisit the surface climate response to large tropical volcanic eruptions from the end of the 19th century until present. We focus on the dynamical driven response of the North Atlantic Oscillation (NAO) and the radiative driven tropical temperature response. Using ten different reanalysis products and the Hadley Centre Sea Level Pressure observational dataset (HadSLP2) we confirm a positive tendency in the phase of the NAO during boreal winters following large volcanic eruptions, although conclude that it is not as clear cut as the current literature suggests. Especially during poorly observed periods where higher uncertainties produce a less robust signal. The phase of the NAO leads to a dynamically driven warm anomaly over Northern Europe in winter. The general cooling of the surface temperature due to reduced incoming shortwave radiation is therefore disturbed by dynamical impacts. In the tropics, where less dynamical driven influences are present, we confirm a predominant cooling after most but not after all eruptions. The observation, reanalysis and model datasets agree well with the in strength of the tropical response. But due to the high uncertainty of the model response, no clear tropical cooling signal was found.

1 Introduction

Understanding of the atmospheric naturally-forced variability is a key issue to estimate the human induced contribution to the recent climate change. Large volcanic eruptions can have an impact on the climate over several years (Robock, 2000). The injected aerosols into the lower stratosphere influence the radiation balance, resulting in a cooling of the tropical surface temperature (Humphreys, 1913, 1940) and a heating in the tropical lower stratosphere (Labitzke and McCormick, 1982; Parker and Brownscombe, 1983). For the eruption of Mt. Pinatubo in June 1991, which was the strongest tropical eruption in the satellite era, the lower tropical stratosphere was warmed up to 3 K (Mitchell et al., 2014; Fujiwara et al., 2015). The cooling signal on the surface is less pronounced and therefore more difficult to separate from other internal and external climate variability. To perform this separation Mitchell et al. (2014) and Fujiwara et al. (2015) used multiple linear regression with nine reanalysis datasets (ERA-Interim, ERA-40, JRA-25, JRA-55, MERRA, NCEP-R1, NCEP-CFSR, NCEP-R2 and NOAA-20CR). A good agreement between most reanalysis datasets for the temperature response in the stratosphere was found. Uncertainties which arise from variability of the differences between the reanalysis models are currently being assessed in the Stratospheric and tropospheric Processes And their Role in Climate - Model inter comparison Project (SPARC-RIP) (Fujiwara and Jackson,

2013). The agreement between the reanalysis datasets in the troposphere is strong but no clear tropospheric cooling could be found, taking into account all large tropical eruptions after 1960 (Fujiwara et al., 2015). A factor which contributes to the uncertainty of the tropospheric cooling is a winter warming signal over Europe. The stratospheric polar vortex in boreal winter is strengthened due to weaker wave flux from the troposphere into the stratosphere and ozone losses in the high latitutes of

5   the stratosphere (Stenchikov et al., 2002). This strengthening of the polar vortex can lead to descending wind anomalies into the troposphere and intensify the westerly winds near the surface (Baldwin and Dunkerton, 1999). With a strengthening of the polar vortex comes a shift of the North Atlantic Oscillation (NAO) towards a positive state, leading to anomalous advection of warm air from the Atlantic to Europe (Thompson and Wallace, 2001). On a global scale the dynamically driven warming weakens the radiative driven cooling after large volcanic eruptions, hence the interplay between the two is important to con-

10   sider.

While studies have looked at the volcanic response on the NAO and tropical surface before, here we investigates both in conjunction, assessing the variability of the NAO and the general temperature response following volcanic eruptions. We test the robustness of previous studies by including observations and all available reanalysis products to examine the uncertainty of the radiative and dynamical response to volcanic eruptions. Our results are also informed by the latest coupled climate models. The

15   analysis of the uncertainty in the reanalysis products will contribute to the SPARC-RIP report, comparing the representation of various atmospheric modes in reanalysis.

The remainder of this paper is as follows. Datasets and methods used in this study are described in section 2, followed by the results in section 3 and the discussion and summery in section 4.

2   Data and analysis method

20   In this study we use near surface monthly mean Temperature of Air at the Surface (TAS) and sea level pressure data from ten available reanalyses (Table 1). Furthermore we analyse the TAS fields of the CMIP5 historical model experiments provided by the World Climate Research Programme (Taylor et al., 2012).

For comparison of the temperature we use the Met Office Hadley Centre HadCRUT4 dataset, available with 100 ensemble members covering the period 1850 until present (Morice et al., 2012). This product has a global coverage with a resolution of

25   $5°x5°$ but it includes missing data depending on the observational data base. The provided data are anomalies referring to the mean climatology between 1961-1990 based on the relatively small number of missing data during this period.

To investigate the dynamical impact of the volcanic eruptions during boreal winter we make use of sea level pressure observations from the Hadley Centre Sea Level Pressure dataset (HadSLP2) (Allan and Ansell, 2006). This product does not contain missing data because of an applied interpolation procedure, which can cause uncertainties, specially in less covered

30   regions like the Arctic, Antarctic or deserts. The spatial resolution of $5°x5°$ is equal to the temperature product but the data includes the period 1850 until 2004. However, an updated version is available using NCEP/NCAR reanalysis fields from 2005 until present, named HadSLP2r (Kalnay et al., 1996). The mean values for both dataset are homogeneous but the variance is higher in HadSLP2r. Nevertheless we consider this adding of the dataset as justified since we use the updated period only for

the calculation of the climatology, significance testing and for the Empirical Orthogonal Function (EOF) analysis in order to calculate the NAO time series (Thompson and Wallace, 1998).

2.1 Reanalysis Data

All reanalysis datasets span at least the period from 1979 until 2012 except two ECMWF products: ERA-20C (20th century reanalysis product) (Poli et al., 2013), which ends in December 2010 and ERA-40 (Uppala et al., 2005), which ends in August 2002. To be able to compare all reanalysis over the same temporal region, we extend the ERA-20C and ERA-40 datasets until 2012 with data of the ERA-Interim reanalysis. ERA-Interim provides some advantages to the ERA-40 version including an improved data assimilation and a better representation of the stratospheric circulation (Dee et al., 2011). Since we use this added data just indirectly for calculations of climatology, anomaly fields and the EOF, the differences which would arise by using a full period consideration are negligible.

Both reanalysis products of the Japan Meteorological Agency (JMA) are used for the study. JRA-25 ends in 2004 but the data until present is available from the JMA Climate Data Assimilation System (JCDAS) with the same system as JRA-25 (Onogi et al., 2007). The subsequent JMA product is called JRA-55 and covers a longer period beginning from 1958. Several improvements have made in comparison to the previous product such as a significant reduction of the large temperature bias in the lower stratosphere by using a new radiation scheme (Ebita et al., 2011).

The MERRA reanalysis obtained from the National Aeronautics and Space Administration (NASA) is focused on the correct simulation of the hydrological cycle and is the only reanalysis used which does not represent the analysis field with spectral coefficients (Rienecker et al., 2011).

The first reanalysis project was operated by the National Centers for Environmental Prediction (NCEP), called NCEP-R1 (Kalnay et al., 1996) and has been updated with the NCEP-R2 product (Kanamitsu et al., 2002). A more sophisticated and recent reanalysis product of NCEP is the Climate Forecast System Reanalysis (NCEP-CFSR) (Saha et al., 2010).

Most products include a period beginning from the late 20th century, which means that just the last two large tropical eruptions of Mt. Pinatubo and El Chichón can be considered for the analysis. Due to the large number of individual ensemble members and the long time series beginning in 1851, the NOAA-CIRES 20th Century Reanalysis Version 2c (NOAA-20CR) is the main product used in this study (Compo et al., 2011). The assimilation scheme of the NOAA-20CR reanalysis product uses a ensemble Kalman Filter in streams of 5 years. Each stream has 56 members.

The ERA-20C product starts in 1900 and includes most of the period which we investigated. The assimilation of these long datasets only include surface observation data, in contrast to the other products assimilating also satellite and radio-sonde measurements. The reanalysis datasets are generally in good agreement with surface observation data, specially for sea level pressure and near surface temperature data, used in this study (Simmons et al., 2004; Makshtas et al., 2007; Lindsay et al., 2014). NCEP-R1 and JRA-25 show differences in the sea level pressure field over Greenland and MERRA generally over mountain areas likely because of distinguish surface pressure reduction methods (Lindsay et al., 2014). Since we use just anomaly fields for our calculations, this should not affect the results significantly.

2.2 CMIP5 models

The model data are obtained from the historic simulations of the CMIP5 models which include simulations with just volcanic forcing (Table 2). All models except the coupled-physical model of the Geophysical Fluid Dynamics Laboratory (GFDL-CM3) start from the beginning of 1850 and have at least three ensemble members. The advantages of the GFDL-CM3 model is a

5   sophisticated interaction scheme between aerosols and clouds and a focus on coupling between the troposphere and stratosphere (Donner et al., 2011). The Community Climate System Model 4 (CCSM4) and the Community Earth System Model version 1 with Community Atmospheric Model version 5 (CESM1-CAM5) models use the aerosol optical depths description of Ammann et al. (2007). All other models use the updated version of the Sato et al. (1993) description. CESM1-CAM5 includes the direct and indirect effects of aerosols (Meehl et al., 2013), while CCSM4 just provides the direct effects. These models show a good

10   reproduction of the ENSO due to an improved deep convection scheme in the atmosphere component (Gent et al., 2011). The ModelE2 version of the NASA Goddard Institute for Space Studies (GISS-E2) provides four different simulations with just volcanic forcing (Schmidt et al., 2014). They differ by using distinguished ocean models and whether the models include interactive chemistry and parametrization of indirect aerosol effects. GISS-E2-R uses the Russell ocean model (Hansen et al., 2007) and GISS-E2-H uses the Hybrid Coordinate Ocean Model (HYCOM) (Sun and Bleck, 2006). Both realizations are

15   available in a version with non-interactive chemistry (NINT), comparable to the prior CMIP3 simulation, but with a tuned aerosol indirect effect following Hansen et al. (2005) and a version with Tracers of Chemistry, Aerosols and their Direct and Indirect effects (TCADI) including interactive chemistry and a parametrization of the first indirect aerosol effects (Menon et al., 2010).

2.3 Choice of volcanoes

20   Not all studies which concentrate on the large-scale impact of volcanic eruptions use the same criteria for the choice of which eruption should be considered for a composite analysis. The volcanic explosivity index (VEI) introduced by Newhall and Self (1982) is a very frequently used measurement for the strength of the eruption (Robock, 2000). The calculation of the index is restricted to volcanic measurements but does account for the height of the eruption column. In Figure 1 all volcanic eruptions since the 19th century with a VEI-index of at least 5 are shown. Additionally volcanoes are included which in other studies are

25   considered to have an impact on the climate but just reached a VEI-index of 4. The size of the triangle of each eruption indicate the respective VEI-index. The colour shows the phase of the El Niño-Southern Oscillation (ENSO) in the first winter after the eruption. The last three tropical eruptions of Mt. Agung in 1963, El Chichón in 1982 and Mt. Pinatubo in 1991 were followed by an El Niño event, suggesting a possible connection between large tropical volcanic eruptions and ENSO (Adams et al., 2003). Since most studies assume that the events are coincidental, we remove the effects of ENSO with a linear regression (see

30   Methods).

Climate models represent volcanic eruption by an increase of the atmospheric aerosols due to the ejected material. Most models use the updated version of the so called Sato-Index (Sato et al., 1993). This index shows the aerosol optical depth (AOD) at wavelength 550 nm and is available as a zonal mean with global coverage and a meridional resolution of around 8°. In Figure 2

the tropical (30°S-30°N) AOD is plotted. As expected from the chosen region the values of low latitude eruptions are generally more pronounced than the extratropical eruptions. The tropical region is characterised by rising air in the stratosphere which lift the aerosols into higher levels. The residual stratospheric meridional circulation transports the aerosols to high latitudes (Trepte and Hitchman, 1992). A volcanic eruption in higher latitudes is expected to have less influence on the climate system

5 because the downward flow in the stratospheric extratropics avoids rising aerosols in higher levels. Nevertheless some studies could show that also extratropical volcanic eruptions can have a significant large scale impact on climate but usually just on the hemisphere where the eruption took place (Graf and Timmreck, 2001; Oman et al., 2005). Since we focus on both the particular impact of the eruption on the NAO and the global temperature response, we consider the AOD in the tropical middle and upper stratosphere. In Figure 2 we draw a threshold at 0.05 nm in order to distinguish which volcanoes likely have a strong

10 impact on the global climate. Therefore we choose the eruption of Krakatau in August 1883, Santa María in October 1902, Mt. Agung in March 1963, El Chichón in April 1982 and Mt. Pinatubo in June 1991 for our analysis.

2.4 Methods

For the calculation of the NAO we use the leading EOF between 0°-70°W and 35°N-80°N during the period 1979-2012 (Thompson and Wallace, 1998; Baldwin et al., 2008). We consider the SLP anomalies of the two winters following the erup-

15 tion as volcanic influenced, except in the case of Santa María which erupted so late during the year that the full influence on the winter circulation is unlikely. The anomalies are calculated with respect to the mean for the years 1979-2012, excluding the following two years after the eruptions of El Chichón and Mt. Pinatubo. The analysis of the temperature fields will be compared to the HadCRUT4 dataset, consisting of anomalies relative to the 1961-1990 reference period (Morice et al., 2012). Since the HadCRUT4 dataset contains missing data, we just use the gridpoints and timesteps of the reanalysis products with

20 non-missing values in the observational record. This ensures that all datasets can be compared directly. As some reanalysis products start in 1979 we just consider the five reanalysis datasets which include this reference period (ERA-20C, ERA-40, JRA-55, NCEP-R1, and NOAA-20CR).

After removing the mean seasonal cycle we subtract the data with a 10 year running mean to remove any further trend. To make sure that the running mean is not influenced by the considered year, we average over the 60 months before the year and

25 the 60 months after the considered year. Both, temperature and NAO response can be influenced significantly by the occurence of ENSO phanomena (Lehner et al., 2016). To remove the effects of ENSO we calculate a monthly linear regression on the Niño 3.4 SST index. This seasonal dependent ENSO pattern is then subtracted from the data. The significance of the resulting anomaly field is calculated with a Monte Carlo test assuming independence between the volcanic eruption events.

3 Results

3.1 Pressure and NAO response

To analyse the NAO response to large volcanic eruptions we use surface pressure data of all available reanalysis products. Since MERRA, ERA-Interim, JRA25, NCEP-R2 and NCEP-CFSR start in 1979 with the beginning of continuous satellite observations we use the years 1979-2012 as our reference period. Two tropical eruptions had a significant impact on the climate system during this period, El Chichón in 1982 and Mt. Pinatubo in 1991. Figure 3 shows the mean SLP anomaly in the first two post volcanic winters after El Chichón and Mt. Pinatubo over the extratropical Northern Hemisphere in observation and for the multi-reanalysis mean.

The response pattern is captured well in all reanalysis products. Since the assimilation of surface pressure data is essential for reanalysis products, the difference between the individual products and the observations is expected to be small (Kalnay et al., 1996). Over the Arctic region a higher level of observational uncertainty is apparent due to the decreased number of assimilated measurements than in the middle latitudes. Low pressure anomalies over the Arctic region are observed, bordered by a positive signal with the centre over Europe. Differences between the reanalysis products are weak but can be seen mainly above mountain areas, e.g. Rocky Mountains (Supplementary Information, Figures S1,S2), likely due to different pressure reduction techniques (Lindsay et al., 2014). Also the NOAA-20CR reanalysis product is in good agreement with other reanalysis datasets. NOAA-20CR is found to capture well the stratospheric temperature response to volcanic eruptions but with a slightly lower amplitude in comparison to other reanalysis products (Mitchell et al., 2014; Fujiwara et al., 2015). At the surface the NOAA-20CR does not show major differences of the pressure response after the eruption of Mt. Pinatubo and El Chichón (Figures S1,S2).

In comparison to an expected positive NAO, in the first winter the high SLP anomaly is shifted towards Central Europe and therefore the negative centre is shifted northwards (Hurrell and Deser, 2009; Hurrell, 1995). Fujiwara et al. (2015) showed that the eruption of Mt. Pinatubo influences the stratospheric temperature and circulation stronger than the eruption of El Chichón. Therefore it is expected that the surface pressure signal is dominated by the response to the eruption of Mt. Pinatubo. Probably due to the east phase of the quasi-biennial oscillation (QBO) in the first winter after the eruption of Mt. Pinatubo the volcanic signal on the SLP was weakened (Stenchikov et al., 2002). In contrast, the response of the second winter (Figure 3 b,d) shows a more NAO like pattern over the North Atlantic with significant positive anomalies in the region of the Azores. A sample of only two single events is not a robust data basis to make general conclusions. Hence, we use the NOAA-20CR product which agrees well with the other reanalysis products, considering the SLP response after the eruption of Mt. Pinatubo and El Chichón (Figures S1,S2) and contains a longer period between 1851-2012. During this period five major tropical eruptions took place: Krakatau, Santa María, Mt. Agung, El Chichón and Mt. Pinatubo. The mean SLP response to these five eruptions is shown in Figure 4. The pattern in the North Atlantic region is similar to the mean over the last two eruptions but the pressure anomaly gradient is smaller, with a predominantly significant response pattern in the first winter. In general, the average pressure anomaly field indicates a shift towards a positive NAO in the first winter after the eruption. A significant mean NAO signal could be found by Christiansen (2008) and Driscoll et al. (2012), taking into account tropical and extratropical eruptions. We

calculated the monthly winter NAO index of just the tropical eruptions and show the index individually for every eruption in Figure 5. The observation data is shown by crosses and all reanalysis data is shown by blue ranges. The orange ranges indicate the NAO index of the 56 NOAA-20CR ensemble members (95% ensemble spread).

The spread of the reanalysis products is small showing high agreement between the individual reanalysis products. The observations agree well with the reanalysis data but in some cases the reanalysis NAO response can differ from the observational signal (e.g. in December and January after the eruption of El Chichón). Especially with higher uncertainty of the observation data in the post satellite era, the differences of the NAO index between the individual members of NOAA-20CR are increased, indicated by a wider spread of the error bar. Therefore some ensemble members can differ with the corresponding observations by strength and even sign of the NAO phase (e.g. Santa María in December).

Consistent with Jones et al. (2003) there is no evidence for a positive NAO shift due to the volcanic eruptions in particular months and in the second winter after the eruption (Figures S3). The response in the mean winter NAO is not clear. In the first winter after the eruptions of El Chichón and Mt. Pinatubo a strong positive NAO phase could be found. Relative to the distribution shown with gray bars, a moderate positive NAO was present following the eruption of Krakatau. For the eruption of Santa María we do not use the first winter after the eruption because the volcano erupted so late during the year that an influence of the injected aerosols to the stratospheric circulation is unlikely. In the winter 1903/04 a NAO-index around zero was found. This agrees with the results of Christiansen (2008) and Driscoll et al. (2012). In the winter directly after the eruption (1902/03) a strong positive NAO was found (Christiansen, 2008). In the winter after the eruption of Mt. Agung we found a negative NAO. Most of the aerosols after the eruption of Mt. Agung were concentrated in the Southern Hemisphere (Figure 2), which reduces the impact on the boreal stratosphere in winter. Therefore we conclude that we do not find a significant positive NAO response to volcanic eruptions with taking just the strongest five tropical eruptions from the end of the 19th century until present. We confirm that the NAO generally shifts towards a positive state in Figure 4 and in Figure 5d) but exceptions like Mt. Agung or Fernandina are found (Figure S4). Responsible for these exceptions can be disturbances of the polar vortex which descend downwards and influence the NAO over a short time scale of some weeks (Baldwin and Dunkerton, 1999). The positive NAO response to volcanic eruptions leads to a positive temperature anomaly over Northern Europe in winter (Robock and Mao, 1992, 1995; Fischer et al., 2007). This warming disagrees with the radiative driven cooling of the troposphere following the eruptions. Therefore just dynamically driven effects like circulation changes could explain this response. It is important to understand these dynamical driven effects in order to understand the total volcanic signal. The studies of Stenchikov et al. (2006) and Driscoll et al. (2012) could show that general circulation models are generally not able to reproduce this secondary effects and hence it is questionable if they reproduce the temperature response well. Only a subset of models show an associated warming over Northern Eurasia but much weaker than the observations (Driscoll et al., 2012; Gillett and Fyfe, 2013).

3.2 Temperature response

Figure 6 shows the TAS anomaly composites over the last three large volcanic eruptions of the reanalysis products in comparison to the observation data. The temperature anomalies are calculated over the whole first or second year after the eruption. We note here that, due to the observational anomalies respective to the period 1961-1990, we only consider the five reanalysis

products which include this period. Influences due to ENSO and temperature trends are removed. We find a significant warming over Northern Europe which is consistent with the expected winter warming over this region. The warming over Siberia and the cooling in the Middle East suggests a positive NAO response following volcanic eruptions (Thompson and Wallace, 1998). The cooling over Alaska and North Canada and the warming over Southeastern United States and North Pacific suggest a negative Pacific/North American teleconnection pattern (PNA). This is consistent to a stronger polar vortex in boreal winter due to less wave propagation into the stratosphere (Garfinkel and Hartmann, 2008). The reanalysis mean shows a general cooling over sea areas, which is significant over the tropical Pacific and Southern Indian Ocean. The HadCRUT4 dataset shows a similar temperature anomaly pattern in high latitudes of the Northern Hemsiphere. The cooling over the oceans is less pronounced in the observation dataset. The second year after the eruptions shows cooling over most continental and ocean areas with exceptions over Northern Siberia and the Eastern Pacific.

To assess a larger sample of eruption events we expand the considered period and include the early eruptions of Krakatau and Santa María. During the period after this eruptions much less observation data is available, especially over continental areas. In Figure 7a,b we averaged over at least four of five years after the volcanic eruptions. The results are similar to those of Figure 6. To be able to evaluate the effect of the volcanic eruptions over less observed areas we calculated the mean temperature response for all five eruptions with the NOAA-20CR product without missing data consideration (Figure 7c,d). Also the consideration of five independent eruptions confirms the significant warming over Northern Europe due to the shift of the NAO towards a positive state. The strongest cooling signals are found in areas with sparse observations. Therefore, these signals highly depends on the reanalysis configuration. In the tropics a general cooling is found over land and sea areas with exception over the Eastern Pacific in the second year after the eruptions. This would confirm that the cooling after large volcanic eruptions is strongest in the tropics (Driscoll et al., 2012). For this reason we focus in the following on the TAS signal in the tropical region (30°N-30°S).

Figure 8 shows the mean anomalies of the tropical temperature in the first and second year after the eruptions, relative to the climatology of 1961-1990. The reanalysis and observation data are generally in good agreement, except in the first year after the eruption of Krakatau and Mt. Pinatubo, with a separation of the reanalysis anomaly spread and the mean temperature reponse in the observational record. This shows that the spread of the reanalysis products does not account for the whole range of uncertainty present in the observations. A general cooling of the tropical TAS response is not always visible in the observation and reanalysis data. After the eruption of Mt. Agung negative and positive anomalies in the first winter after the eruption are found. The seond year after the eruption of El Chichón was characterized by a warming of the tropics. The study of Fujiwara et al. (2015) did not show a clear temperature signal in the tropical troposphere after the eruption of El Chichón. The eruption of Mt. Pinatubo was the strongest in the 20th century and caused a cooling around 0.1 to 0.2 K, according to the observation and reanalysis data. In the first year after the eruption of Krakatau the cooling was just weak, regarding the TAS anomaly distribution, shown with gray bars. We generally find a cooling of the tropical temperature following large volcanic eruptions. The global mean reponse is less robust due to the impact of dynamical warming especially over Europe and the high uncertainty range of the NOAA-20C dataset after the eruption of Krakatau and Santa Maía (Figure S6).

The historical simulations of the CMIP5 models involve far higher range of TAS responses than the observations. Most simu-

lations with just volcanic forcing show a negative temperature signal. But due to the large spread of simulated TAS response just the eruptions of Mt. Pinatubo shows a significant tropical cooling. For consistency we calculated the anomalies relative to the same period as the observation and reanalysis data. This period (1961-1990) was highly influenced by volcanic activity. By calculating anomalies relative to a non-volcanic influenced period (e.g. 1931-1960) we found that all simulations with just volcanic forcing show a negative TAS response after both years of all five eruptions (not shown). The mean TAS response to the eruptions of the historical simulations including also anthropogenic and other natural forcing agrees well with the simulations with just volcanic forcing.

4 Conclusion and Discussion

In this study we investigated the uncertainty in surface climate response to strong volcanic eruptions. The most up-to-date available reanalysis products, GCMs and the newest observation datasets are used to best evaluate the radiative driven tropical temperature response and the dynamical driven NAO response following eruptions. Given the availability of these new datasets, it is timely to revisit the surface response to test the robustness of past studies of volcanic influences on climate. A summary of volcanic eruption intensity, and occurrence of El Niño events is present in Figure 1 from 1880-present.

The shift of the NAO towards a positive state in boreal winter due to an intensification of the polar vortex was noted in some observational studies (e.g., Shindell et al., 2004) but only some models with a good representation of the stratosphere are able to reproduce the associated winter warming over Eurasia (Kirchner et al., 1999). The CMIP5 models generally fail or underestimate the impact of the volcanic eruptions on the Northern Hemispheric circulation (Driscoll et al., 2012). This shows that the dynamical mechanism is still not fully understood (Graf et al., 2007). Conditional on the injected material into the stratosphere, we selected the strongest five tropical volcanic eruptions from the end of the 19th century until present for our analysis. They are expected to have the biggest impact on the atmosphere.

We showed that a positive NAO phase is likely to be present during the first post volcanic winter, but uncertainties still remain because not all winters following large tropical volcanic eruptions show a positive NAO (e.g. Mt. Agung and Fernandina). Also none of the particular winter months show a significant shift of the NAO after the eruptions. By taking into account all available reanalysis datasets and the HadSLP2 observation data we have seen that there is a general agreement between the datasets. It is known that the atmospheric condition after the volcanic eruption is a big source of uncertainty for the impact of eruptions on the NAO. The QBO phase (Holton and Tan, 1982; Stenchikov et al., 2004), an El Niño or La Niña event (Moron and Plaut, 2003; Manzini et al., 2006; García-Herrera et al., 2006; Calvo et al., 2009) as well as the solar variability (Lean et al., 1995; Haigh, 2002; Gray et al., 2013) can influence the NAO phase directly or indirectly by modulating the stratospheric winter circulation in the Northern Hemisphere. A more robust indicator for the strengthening of the dynamical driven influence of the volcanic eruption is the characteristic winter warming over Northern Europe (Shindell et al., 2004; Fischer et al., 2007). By averaging over the whole first year after the eruption we still could find a significant positive signal in Northern Europe due to the winter warming. This means that the general decrease of the surface temperature due to the injected aerosols is overwhelmed by this dynamical effect at mid-latitudes. This contributes to the fact that the strength of the radiative driven cooling is still not

completely determined. The less dynamical influenced tropical response to volcanic eruptions show a general cooling mainly over ocean areas. Nevertheless, considerung the first two years after the eruption of Mt. Agung and El Chichón no clear tropical cooling in both reanalysis and observation data is found. The model reponse to the eruptions shows a large uncertainty spread, suggesting a high dependence of the climate response on the atmospheric state during and after the eruption.

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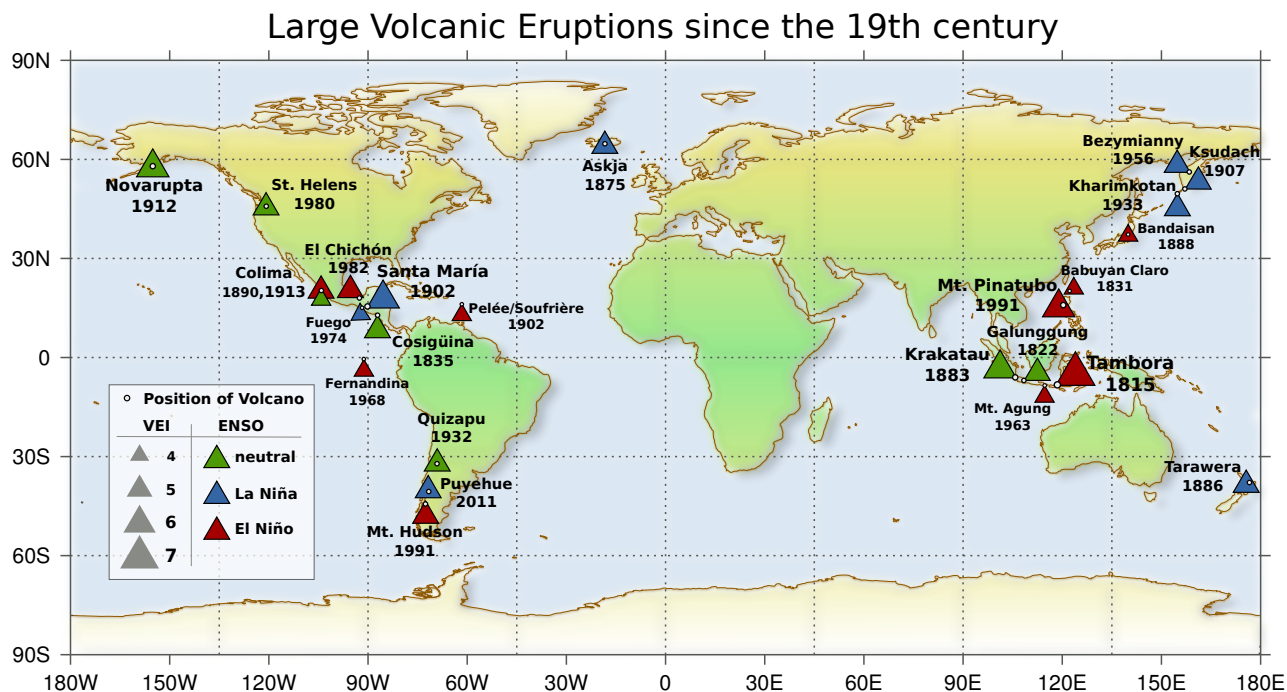
Figure 1. Large volcanic eruptions since the 19th century. The size of the triangle indicates the volcanic explosivity index (VEI) (Newhall and Self, 1982), from Simkin and Siebert (1994) and the Smithsonian Institution (www.volcano.si.edu). The filled colour shows the El Niño-Southern Oscillation (ENSO) event in the first winter after the eruption from Cook et al. (2009) until 1870 and from Wolter and Timlin (2011) from 1871 until present. In the case of Fuego, Santa María, Babuyan Claro and Galunggung the second winter after the eruption is considered for the definition of the ENSO phase.
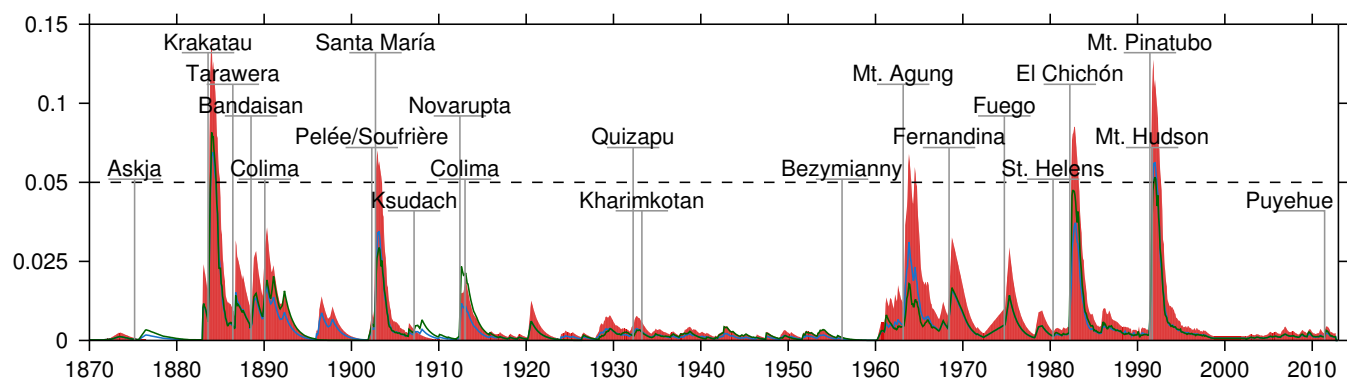
Figure 2. Tropical averaged Stratospheric Aerosol Optical Depth (AOD) at 550nm for the altitude region 20-25 km from Sato et al. (1993), updated dataset from data.giss.nasa.gov/modelforce/strataer/. All volcanic eruptions from Figure 1 for the period 1870 until present are included. They are indicated with a gray line corresponding to the eruption time and the name of the volcano. The green line is the mean AOD for the Northern Hemisphere and the blue line shows the mean AOD for the Southern Hemisphere.
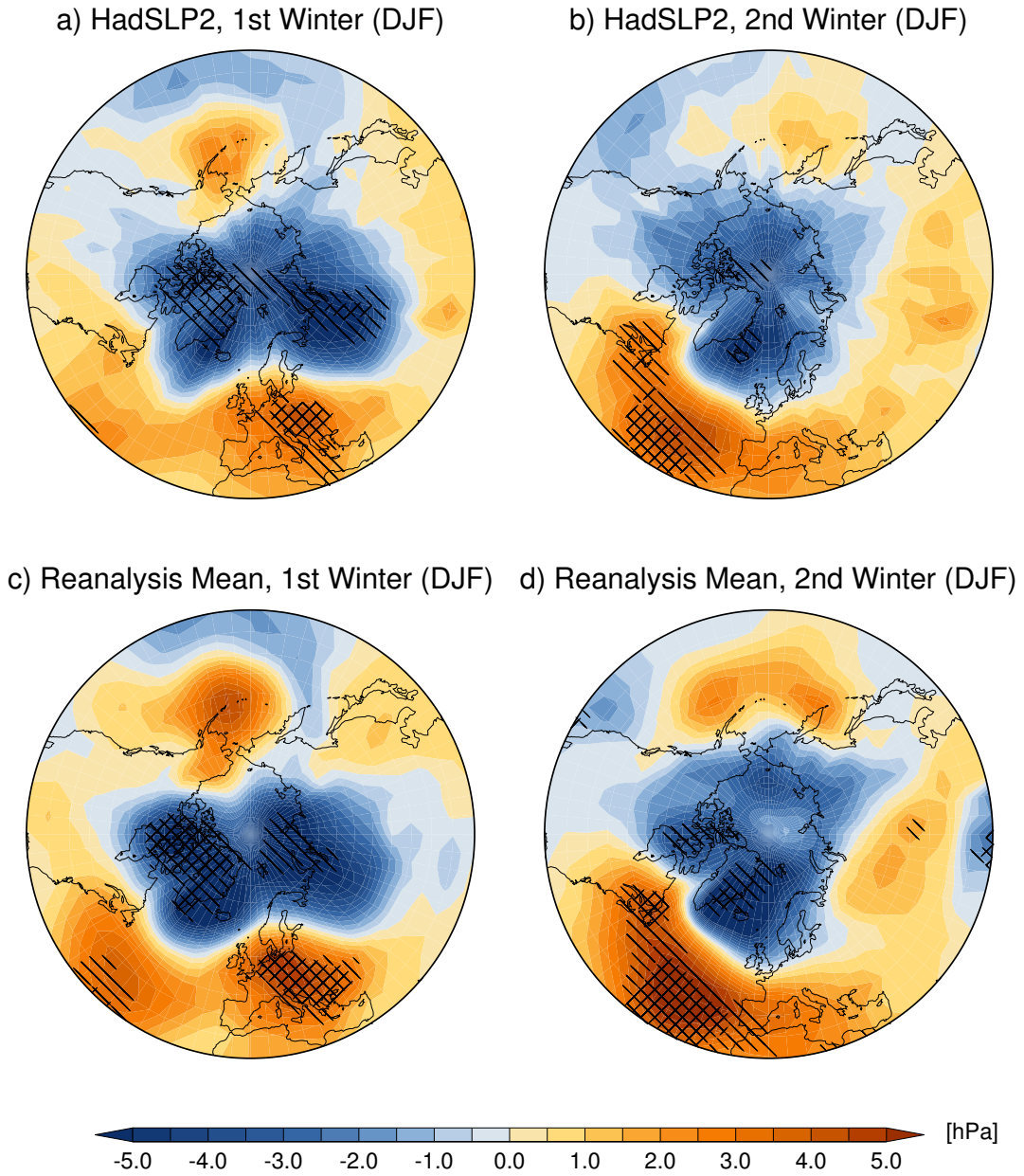
Figure 3. Observed and multi-reanalysis mean SLP anomalies (hPa) in the Northern Hemisphere in the first two post volcanic winters (DJF) averaged over the two volcanic eruptions of Mt. Pinatubo (1991) and El Chichón (1982). Anomalies are calculated with respect to the mean for the years 1979-2012, excluding the following two years after the eruptions. Single diagonal lines correspond to the 90% and double diagonal lines to the 95% confidence level obtained with a Monte Carlo test of two independent samples. a) Shows the mean anomaly for the HadSLP2 observation data for the first and b) for the second winter after the eruptions. c) and d) show respectively the mean anomaly of all 10 reanalysis datasets (Table 1).
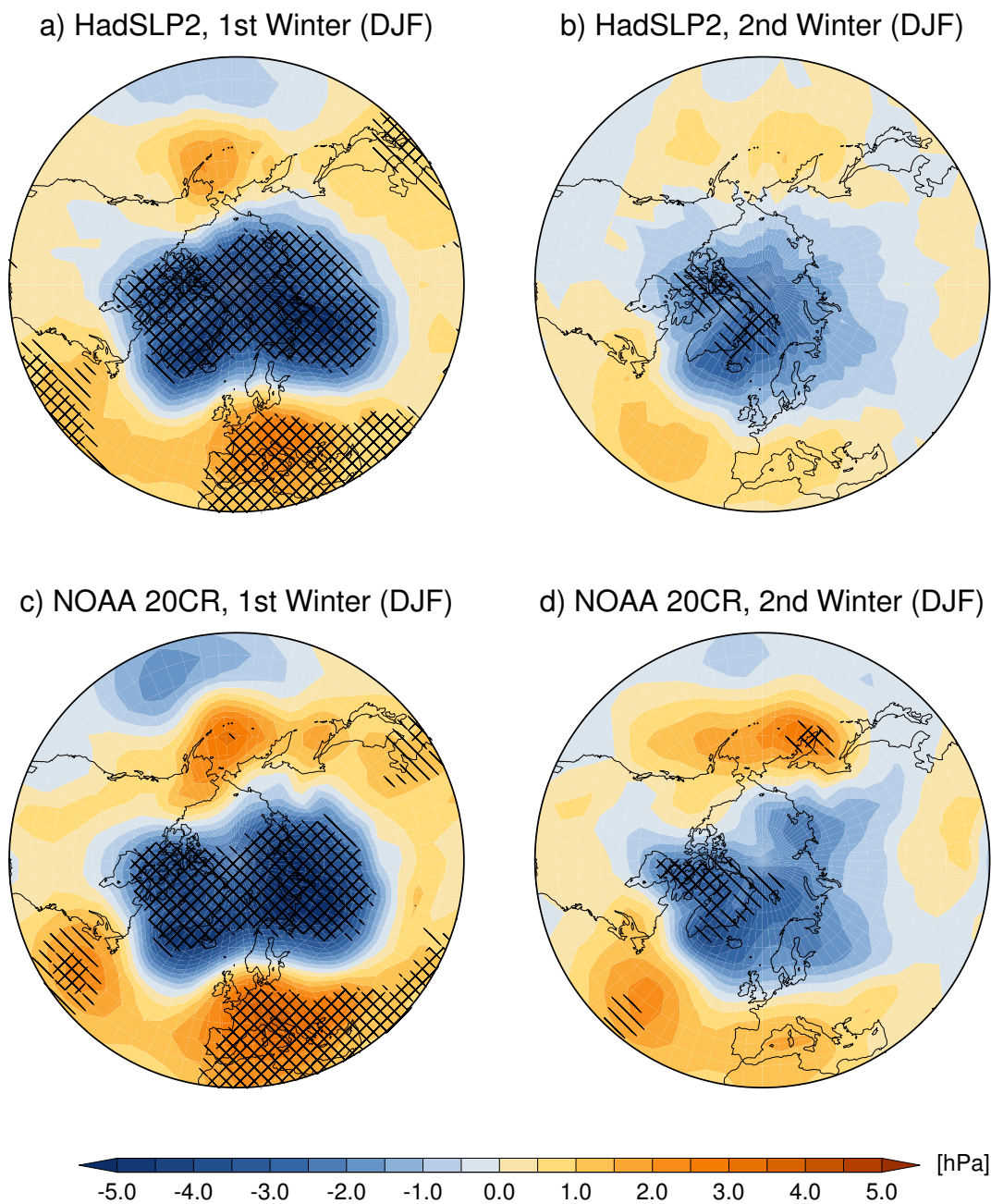
Figure 4. As Figure 3 but for just NOAA-20CR data and HadSLP2 data, averaged over all five post volcanic winters.
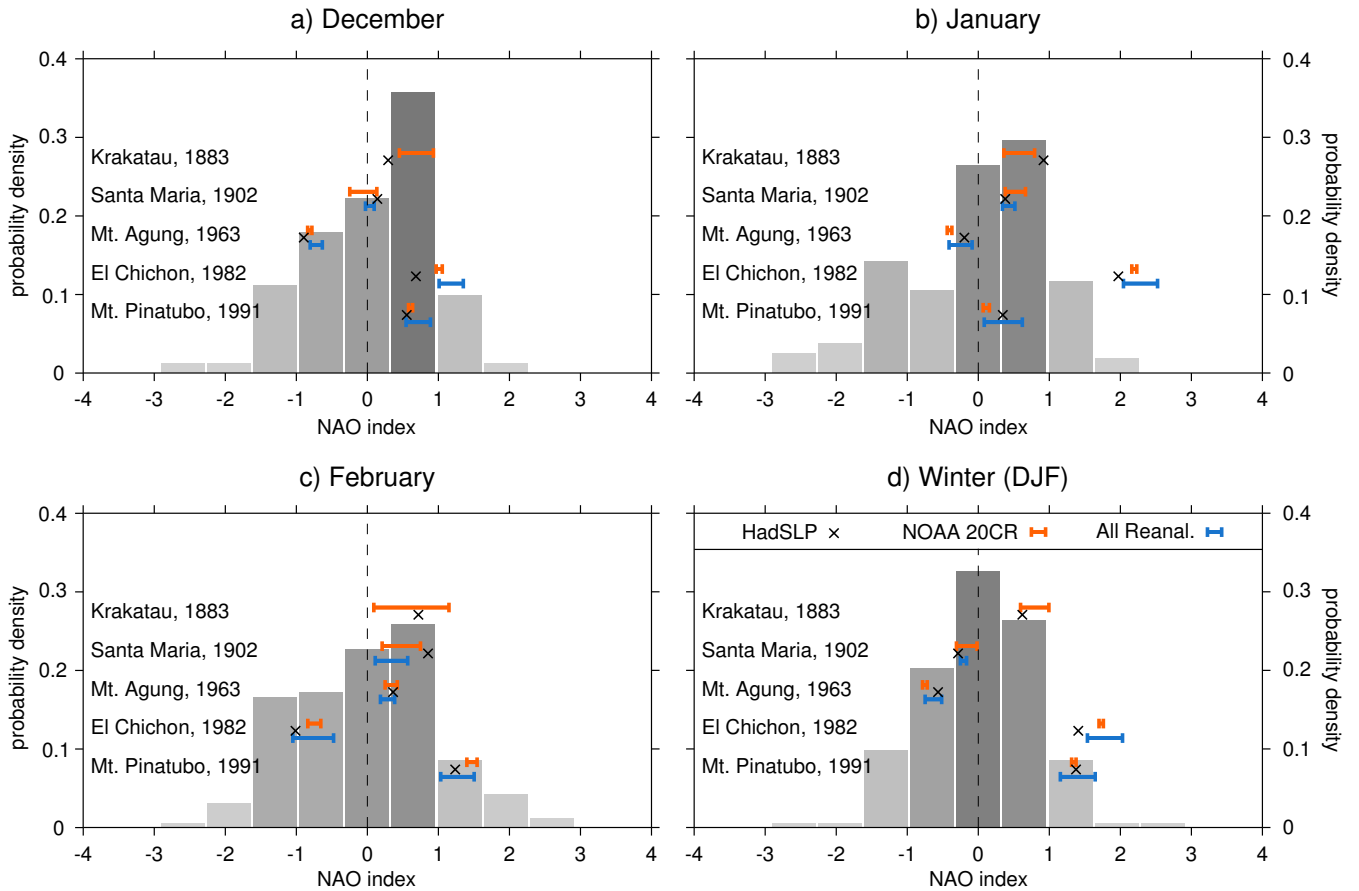
Figure 5. Monthly North Atlantic Oscillation (NAO) index of the first winter after five volcanic eruptions (a-c) and the winter mean (d), calculated with HadSLP2 observation data and reanalysis data. All data is calculated with respect to the mean for the years 1979-2012, excluding the following two years after the two eruption. The EOF is calculated over the period 1979-2012 for every product separately. The histogram shows the NAO index of the 163 years of observation data (1850-2012). Black crosses show the results of the HadSLP2 data. The blue lines show the reanalysis data spread of the NAO index for the winter after all eruptions except Krakatau. The orange lines show the 95% ensemble spread of the NAO after all five volcanic eruptions calculated with the NOAA-20CR dataset.
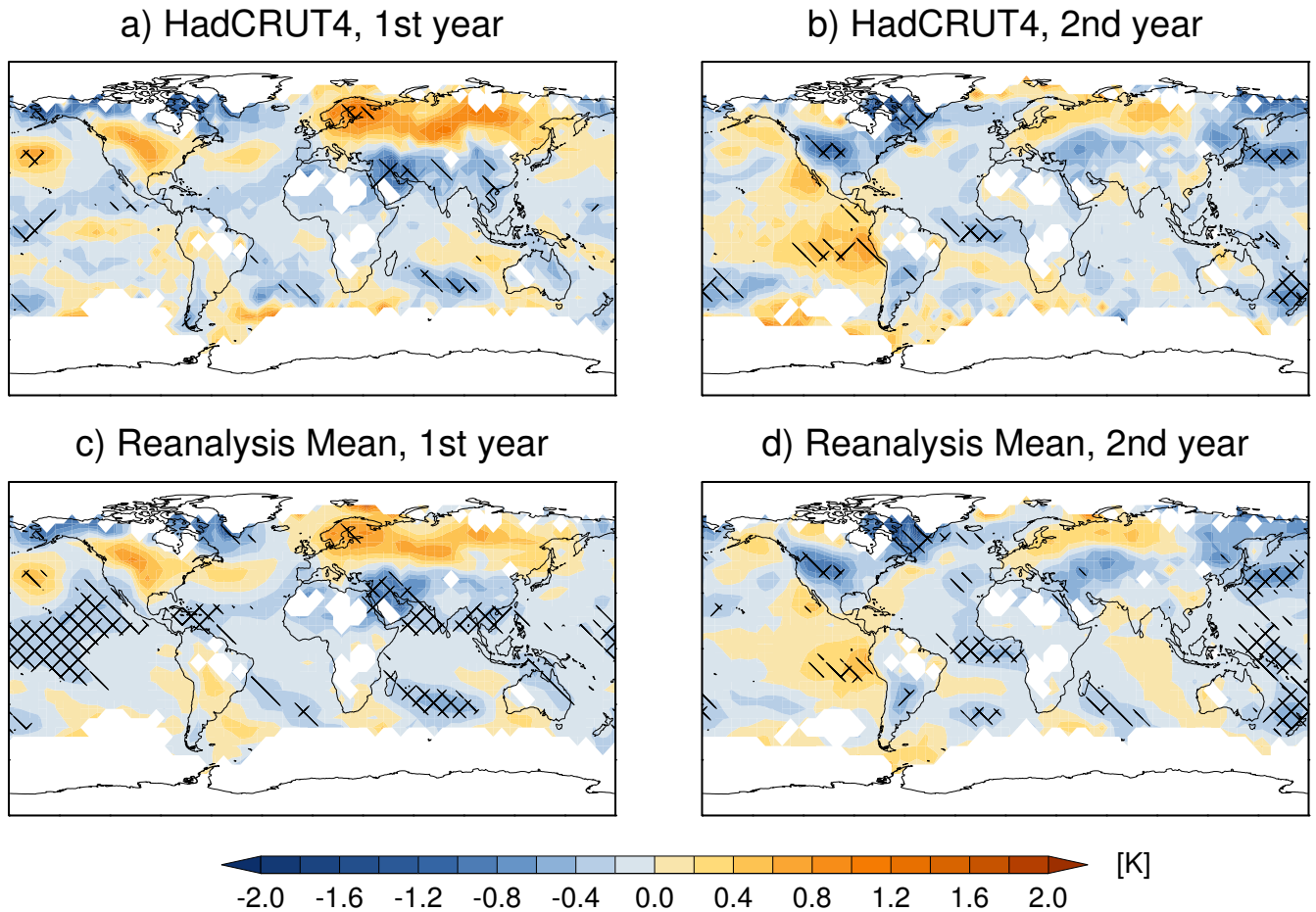
Figure 6. Observed and multi-reanalysis mean TAS anomalies (K) averaged over the first year (a and c) and second year (b and d) after the eruption and over the three volcanic eruptions Mt. Pinatubo (1991), El Chichón (1982) and Mt. Agung (1963). a) and b) show the mean anomaly for the HadCRUT4 observations and c) and d) show respectively the mean anomaly of the five reanalysis datasets, containing this period (Table 1). Anomalies are calculated with respect to the average over the years 1961-1990, consistent to the HadCRUT4 dataset. Single diagonal lines correspond to the 90% and double diagonal lines to the 95% confidence level obtained by a Monte Carlo test of three independent samples.
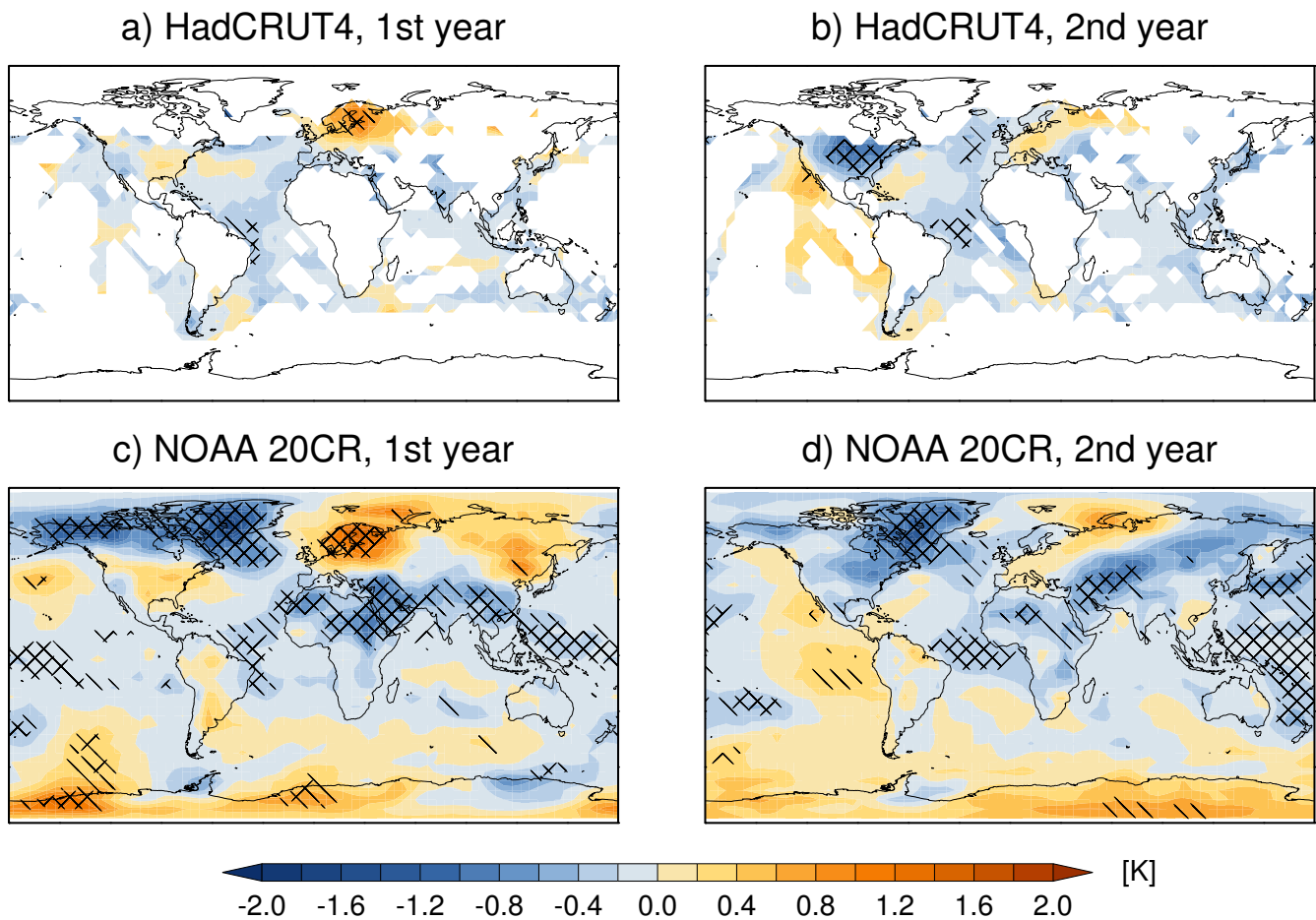
Figure 7. As Figure 6 but for just HadCRUT4 data in a-b averaged over at least four of five eruptions and NOAA-20CR data without missing data considereration, averaged over all five post volcanic years.
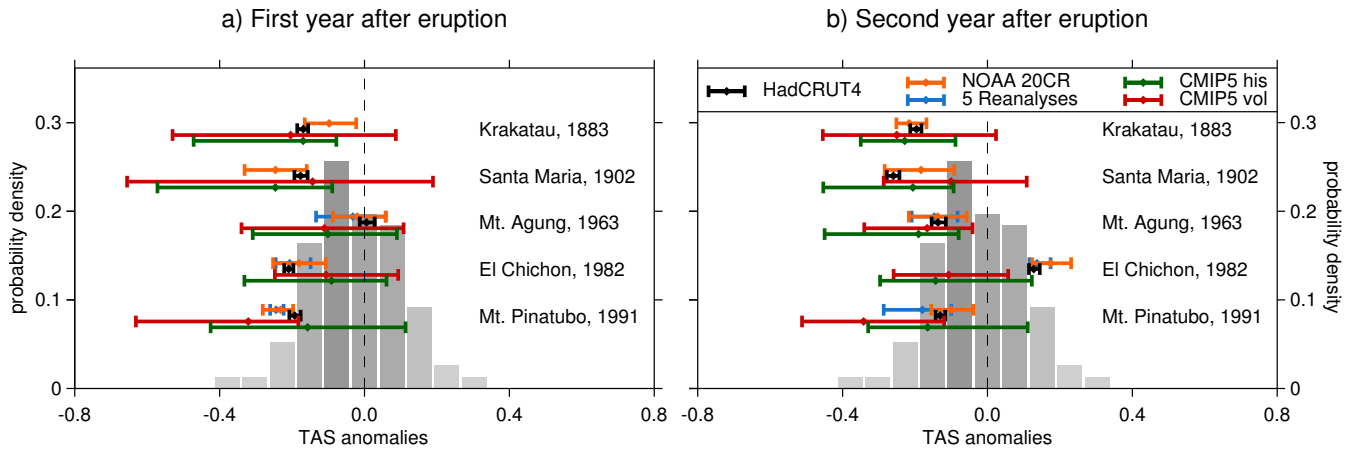
Figure 8. Tropical TAS anomalies (K) averaged over the region 30°S-30°N and over the first year (a) and second year (b) after five volcanic eruptions, calculated with HadCRUT4 observation data and reanalysis data. Anomalies are calculated with respect to the average over the years 1961-1990, consistent to the HadCRUT4 dataset. The histogram shows the anomalies of 153 years of observation data (1855-2007). The first and last five years are leaved out because of the trend removal. The blue lines show the reanalysis data spread of the TAS anomalies for the years after all eruptions except Krakatau and Santa María. The orange lines show the 90% ensemble spread of TAS after all five volcanic eruptions calculated with the NOAA-20CR dataset. The mean response is indicated by the dot between the whiskers.

Table 1. Reanalysis products used in this study

Reanalysis Name	Reanalysis Resolution	Ensemble members	Time span	Volcanic aerosol changes	Reference
NASA MERRA	2/3° Lat x 1/2° Lon, L72	1	Jan 1979 - present	no	Rienecker et al. (2011)
ERA-40	T159, L60	1	Sept 1957 - Aug 2002	no	Uppala et al. (2005)
ERA-Interim	T255, L60	1	Jan 1979 - present	no	Dee et al. (2011)
ERA-20C	T159, L91	1	Jan 1900 - Dec 2010	no	Poli et al. (2013)
JRA25/JCDAS	T106, L40	1	Jan 1979 - Jan 2014	no	Onogi et al. (2007)
JRA55	T319, L60	1	Jan 1958 - present	no	Ebita et al. (2011)
NCEP-1 (R-1)	T62, L28	1	Jan 1948 - present	no	Kalnay et al. (1996)
NCEP-2 (R-2)	T62, L28	1	Jan 1979 - near present	no	Kanamitsu et al. (2002)
NCEP-CFSR / CFSv2	T382, L64	1	Jan 1979 - present	yes	Saha et al. (2010)
NOAA-CIRES 20th Century Reanalysis (v2c)	T62, L28	56	Jan 1851 - Dec 2014	yes	Compo et al. (2011)

Table 2. CMIP5 models used in this study

Model Name	Model Resolution	Ensemble members	Time span	Volcanic aerosol changes	Reference
CCSM4	0.9° Lat x 1.25° Lon, L26	3	Jan 1850 - Dec 2005	A07[a]	Gent et al. (2011)
CESM1-CAM5	0.9° Lat x 1.25° Lon, L30	3	Jan 1850 - Dec 2005	A07	Meehl et al. (2013)
CSIRO-Mk3-6-0	T63, L18	5	Jan 1850 - Dec 2012	S93[b]	Rotstayn et al. (2010)
GFDL-ESM2M	C48, L48	1	Jan 1860 - Dec 2005	S93 / S98[c]	Donner et al. (2011)
GISS-E2-H NINT	2° Lat x 2.5° Lon, L40	5	Jan 1850 - Dec 2005	S93	Schmidt et al. (2014)
GISS-E2-H TCADI	2° Lat x 2.5° Lon, L40	5	Jan 1850 - Dec 2012	S93	Schmidt et al. (2014)
GISS-E2-R NINT	2° Lat x 2.5° Lon, L40	5	Jan 1850 - Dec 2005	S93	Schmidt et al. (2014)
GISS-E2-R TCADI	2° Lat x 2.5° Lon, L40	5	Jan 1850 Dec 2012	S93	Schmidt et al. (2014)

[a] Ammann et al. (2007), [b] Sato et al. (1993), [c]Stenchikov et al. (1998).