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Climate of the Past
Discussions

The 1600 Huaynaputina Eruption as Possible Trigger for Persistent Cooling in the North Atlantic Region

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Abstract. Paleoclimate reconstructions identify a period of exceptional summer and winter cooling in the North Atlantic region following the eruption of the tropical volcano Huaynaputina (Peru) in 1600 CE. Numerical climate simulations indicate a possible eruption-induced mechanism for the persistent cooling in a slowdown of the North Atlantic subpolar gyre (SPG) and consequent ocean-atmosphere feedbacks. To examine the possibility of such an eruption-induced cooling mechanism, this study compares simulations with and without volcanic forcing and an SPG shift to reconstructions from annual proxies in natural archives and historical written records as well as contemporary historical observations of relevant climate and environmental conditions. These reconstructions and observations demonstrate patterns of cooling and sea ice expansion consistent with, but not necessarily indicative of, an eruption trigger for the proposed SPG slowdown mechanism. The results point to possible improvements in future model-data comparison studies utilizing historical written records. Moreover, we consider historical societal impacts and adaptations associated with the reconstructed climatic and environmental anomalies.

1 Introduction

This article draws on high-resolution climate proxies and direct historical observations to examine whether the Huaynaputina eruption in 1600 CE was the trigger for persistent summer and winter cooling in the North Atlantic region during the early 17th century. Huaynaputina is a stratovolcano in southern Peru (4,850m asl; 16.61°S, 70.85°W) and belongs to the Central Volcanic Zone of the Andean Volcanic Belt originating in the subduction of the oceanic Nazca plate beneath the continental South American plate (Global Volcanism Project, 2021). The VEI 6 1600 Huaynaputina eruption was the largest eruption in the Andes in historical times and the largest source of dust in the annually resolved Quelccaya ice-core record (Thompson et al.,

https://doi.org/10.5194/cp-2021-82
Preprint. Discussion started: 29 July 2021

Climate of the Past Discussions

2013). According to contemporary written accounts, tremors began to be felt in the region in mid-February 1600. A large Plinian eruption took place on February 19, followed by a low, heavy ashfall that blackened the sky for tens of kilometers around. Pyroclastic flows continued through February 26, and a phase of vulcanian eruption continued to eject pumice and ash through approximately the beginning of March, with a total ejection of ~13-14 km^3 of tephra reaching up to 400 km from the volcano (Jara et al., 2000; Adams et al., 2001; Thouret et al., 2002; Prival et al., 2019). Pyroclastic flows, lahars, meltwater flooding, and earthquakes caused widespread damage and loss of life in the surrounding region (De Silva et al., 2000).

The first years of the 17th century stand out as some of the coldest of the past two millennia in multiple Northern Hemisphere (NH) summer temperature reconstructions, particularly those based on tree-ring density (Schneider et al., 2015; Stoffel et al., 2015; Guillet et al., 2017). Several reconstructions rank 1601 as the coldest summer and/or 1600-09 as the decade with the coldest NH summer temperatures in at least the past 420 years (D'Arrigo et al., 2006; D'Arrigo et al., 2009). Moreover, 1601 stands out as an extreme cold and/or dry summer in numerous regional climate reconstructions around the North Atlantic, including those for Quebec (Gennaretti et al., 2014) and Scandinavia (McCarroll et al., 2013). This abrupt and exceptional cooling has been attributed to the direct radiative response to the enhancement of the stratospheric aerosol layers by the Huaynaputina eruption (De Silva and Zielinksi, 1998; Briffa et al., 1998; Verosub and Lippman, 2008). Previous studies have linked this eruption to a volcanic dust veil and reported dimming of sunlight (Lamb, 1970). For example, a Russian chronicle records persistent dark skies in spring 1601 (Akiander, 1849), pointing to the persistence of aerosols in the stratosphere more than a year after the eruption. Sulfate depositions dated to ca. 1600 in ice cores at both poles and the tropics have also been attributed to the Huaynaputina eruption (De Silva et al., 1998; Thompson et al., 2013).

Despite the large body of paleoclimatic and historical sources, the connection between the Huaynaputina eruption and concomitant climatic anomalies poses unresolved research issues. Ice core records with robust sampling indicate that the Huaynaputina eruption injected fewer sulfates into the stratosphere than other Little Ice Age (LIA) eruptions with a comparable cooling effect, such as the 1453 Kuwae or the 1815 Tambora eruptions, and that these sulfates were asymmetrically distributed between the Northern and Southern Hemispheres (Sigl et al., 2015; Toohey and Sigl, 2017). European historical records and temperature reconstructions utilizing written records of weather observations and climate proxies—the "archives of societies" (Brönnimann et al., 2018)—indicate that there was more exceptional winter cooling than summer cooling in the years following the eruption, particularly in Central and Northern Europe (Luterbacher et al., 2004; Pfister et al., 2018). Most importantly, the observed summer and especially winter cooling persisted longer than would be expected from the direct radiative forcing by transient stratospheric volcanic aerosols alone (Stoffel et al., 2015; Toohey et al., 2019).

Recent work in numerical climate modeling, supported by paleocenographic, temperature, and atmospheric circulation reconstructions, has suggested a possible mechanism for this amplified and persistent cooling, particularly in and around the North Atlantic (Moreno-Chamarro et al., 2017a). The mechanism features sea-ice expansion and reduced ocean heat losses in the Nordic and Barents seas driven by a reduction in the northward heat transport by the North Atlantic subpolar gyre (SPG). In this mechanism, feedbacks between North Atlantic cooling and sea-ice buildup, on the one hand, and sub-Arctic atmospheric pressure anomalies, on the other, lead to an increased frequency of atmospheric blocking conditions over Europe, in turn

https://doi.org/10.5194/cp-2021-82
Preprint. Discussion started: 29 July 2021

Climate of the Past
Discussions

driving extreme winter cooling over the continent and particularly over Scandinavia and Central Europe. This mechanism does not entail significant changes in the dominant mode of large-scale atmospheric variability over the North Atlantic, namely the North Atlantic Oscillation (NAO), which is usually involved in post-eruption interannual and decadal climate variability (e.g., Zanchettin et al., 2013).

However, despite the agreement between SPG weakening and seasonally asymmetric European cooling in simulations and reconstructions, aspects of the proposed mechanism remain unclear. Ensemble sensitivity simulations with the same imposed forcing and different initial conditions have called into question whether volcanic forcing was necessary to trigger the SPG shift, and they have revealed substantial uncertainties about the initial anomaly in Arctic freshwater export that triggers the SPG-centered feedback loop (Moreno-Chamarro et al., 2017b). Previous studies have drawn on only a few proxies in natural archives, mostly at decadal or lower resolution, to determine the magnitude and timing of shifts in ocean sea surface temperature and sea-ice extent characteristic of the SPG slowdown (Moreno-Chamarro et al., 2017a,b).

The question of whether the Huaynaputina eruption triggered persistent cooling has implications for human history as well. Past research has associated climatic anomalies—particularly cool, wet summers—with harvest failures, price spikes, excess mortality, and civil unrest across Europe during the 1590s and early 17th century (Clark, 1985; Parker, 2013; Parker, 2018; Degroot, 2018). These impacts included severe subsistence crises in Scandinavia (Utterström, 1955; Huhtamaa and Helama, 2017) and Russia (Dunning, 2001) during exceptionally cold years. By constraining the timing of summer and winter cooling and the role of the Huaynaputina eruption therein we can help determine the relative roles of climatic versus non-climatic causes, volcanic forcing versus intrinsic variability of the climate system, and short-term anomalies versus longer-term climatic changes in those societal impacts. Moreover, by analyzing climatic changes in both summer and winter at high temporal and spatial resolution, as well as environmental changes such as sea ice duration and extent, this study may help specify the nature of societal vulnerabilities and possibilities for adaptation ca.1600 CE.

Therefore, this study applies new high-resolution data to determine whether the 1600 Huaynaputina eruption was the trigger for persistent cooling in the early 17th century. It compares paleoclimate simulations with and without volcanic forcing and an SPG shift to reconstructions from annual proxies in natural archives and historical written records as well as direct historical observations of relevant climate and environmental conditions. We seek to establish whether the years following the eruption saw an abrupt shift in mean conditions of sea-ice extent and/or winter temperature that are characteristic traits of the SPG mechanism and whether particular climatic conditions at the time of the Huaynaputina eruption determined activation (or not) of the SPG slowdown mechanism. Thus, this study examines the potential for combining natural archives and written historical records to test model-derived mechanisms during the pre-instrumental period, and thus to further the integration of climate modeling, paleoclimatology, and historical climatology for a better understanding of past climate variability and its human dimensions.

https://doi.org/10.5194/cp-2021-82
Preprint. Discussion started: 29 July 2021

2. A Mechanism for Persistent Cooling ca. 1600 CE and Current Paleoclimate Evidence

Volcanic eruptions such as Huaynaputina influence climate by ejecting chemically and microphysically active gases and solid particles into the atmosphere (Robock, 2000; Timmreck, 2012). During strong eruptions, the volcanic column penetrates the lower stratosphere, where sulfur-containing gases turn into aerosol particles that enhance the existing stratospheric aerosol layer. Volcanic aerosols then typically persist in the stratosphere for up to two years (e.g., Timmreck, 2012; Zanchettin et al., 2012). Aerosol particles influence the Earth's radiative balance both by scattering incoming solar radiation back to space, which cools the troposphere and the surface, and by absorbing long-wave solar and terrestrial radiation, which locally warms the stratosphere.

Although these radiative effects are short-lived, volcanic eruptions may influence the decadal and longer evolution of regional climates in two ways (e.g., Zanchettin, 2017). First, close successions of small or moderate eruptions—so-called volcanic clusters—may lead to prolonged periods of surface cooling associated with the persistence of an enhanced stratospheric aerosol layer. Second, dynamical responses may initiate decadal or multidecadal feedback loops encompassing large-scale atmospheric circulation, sea ice, and oceanic thermohaline circulation (e.g., Ottera et al., 2010; Zanchettin et al., 2012).

Moreno-Chamarro et al. (2017b) have investigated such a scenario of eruption-induced persistent cooling for the period around 1600 CE. According to paleoclimate model simulations, the late 16th-century cluster of volcanic eruptions culminating in Huaynaputina produced a cumulative global-average negative radiative flux of about -5 Wm^{-2} at the top of the atmosphere (Fig. 1a; see also Moreno-Chamarro et al., 2017b). This volcanic cluster includes the 1585 eruption of Colima, Mexico (VEI 4), a 1591 unidentified eruption, and the 1595 eruption of Nevado del Ruiz, Colombia (VEI 4) observed by contemporaries (Thouret et al., 1990; Simón, 1892, pp. 127-129). Each of these eruptions injected sulfates into the stratosphere (Sigl et al., 2013) and had significant short-term cooling effects on NH summer temperatures (Sigl et al., 2015).

As shown by paleoclimate simulations of the last millennium, this cluster of eruptions could have contributed to the onset and persistence of LIA-like climate anomalies (Helama et al., 2021) for two or more centuries due to a sudden increase in the freshwater export from the Arctic that pushed the SPG into a stable weak state (Moreno Chamarro et al., 2017a), as illustrated in **Figure 1**. More specifically, volcanic cooling abruptly increased Arctic sea ice and the associated freshwater export (Figure 1b), which induced freshening of the upper subpolar North Atlantic (Figure 1c); the freshening decreased upper ocean density in the Labrador Sea and thereby weakened the baroclinic zonal density gradient (Figure 1d) that controls the SPG strength (e.g., Born and Stocker, 2014). Once in a weak mode, the SPG induced long-lasting anomalies in the surrounding subpolar and polar ocean, affecting the oceanic transports of heat and salt as well as the sea ice, which helped sustain the weak SPG mode and the associated anomalous climate state (Figure 1e). The existence of bistability in the SPG in different climate models (Born et al., 2013) supports the possibility of such a regime shift of the subpolar North Atlantic Ocean.

https://doi.org/10.5194/cp-2021-82
Preprint. Discussion started: 29 July 2021

Climate of the Past Discussions

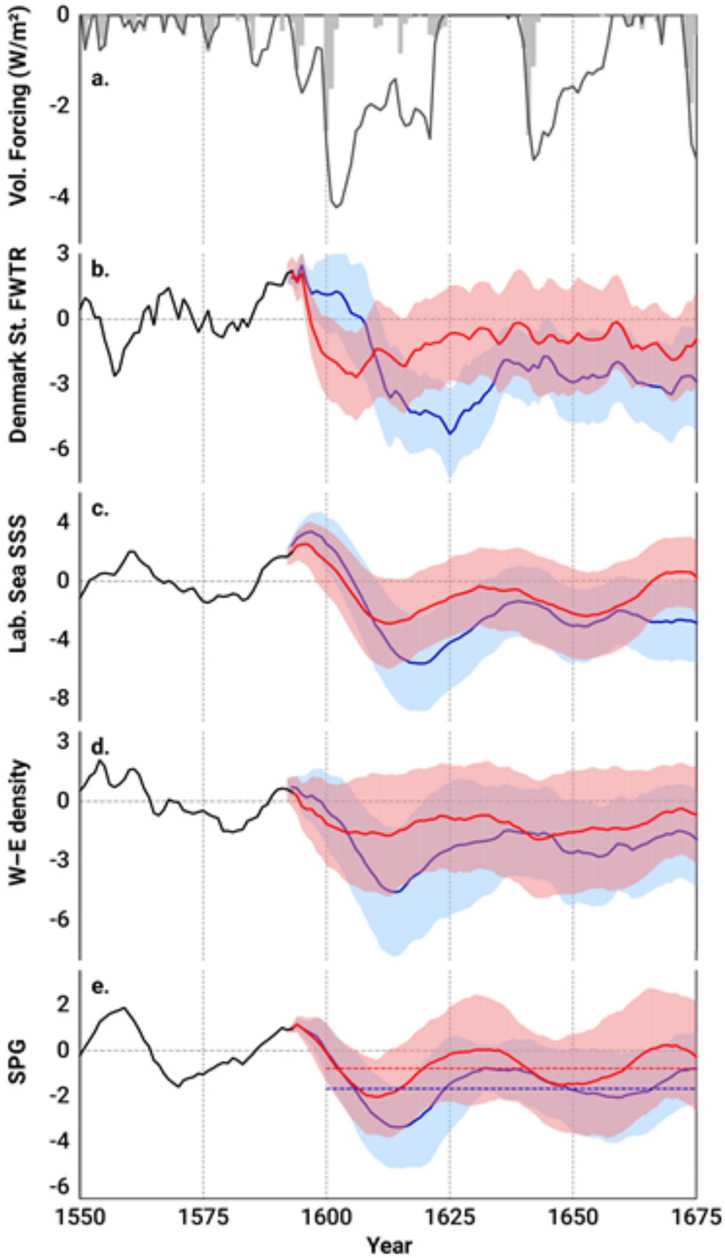


Figure 1: Evolution of North Atlantic climate around the 1600 Huaynaputina eruption from ensemble simulations with MPI-ESM-P (after Moreno-Chamarro et al., 2017a). Blue: realizations with an SPG shift around 1600; red: realizations with no SPG shift. Continuous lines: means of realizations; shadings: standard deviation across realizations; dashed lines: post-1600 temporal averages showing the change in regime in the SPG. From top: (a) Accumulated global top-of-the-atmosphere radiative volcanic forcing (line) and the contribution of each individual eruption (bars), both in W/m²; (b) Denmark Strait freshwater transport (FWTR), as the sum of the liquid and the sea-ice contributions (a larger negative value reflects a stronger southward export); (c) Labrador Sea sea-surface salinity (SSS); (d) West–east gradient in upper-ocean densities in the subpolar North Atlantic; and (e) SPG strength. (b–e) Time series normalized with respect to the pre-eruption period 1550–1590.

https://doi.org/10.5194/cp-2021-82
Preprint. Discussion started: 29 July 2021

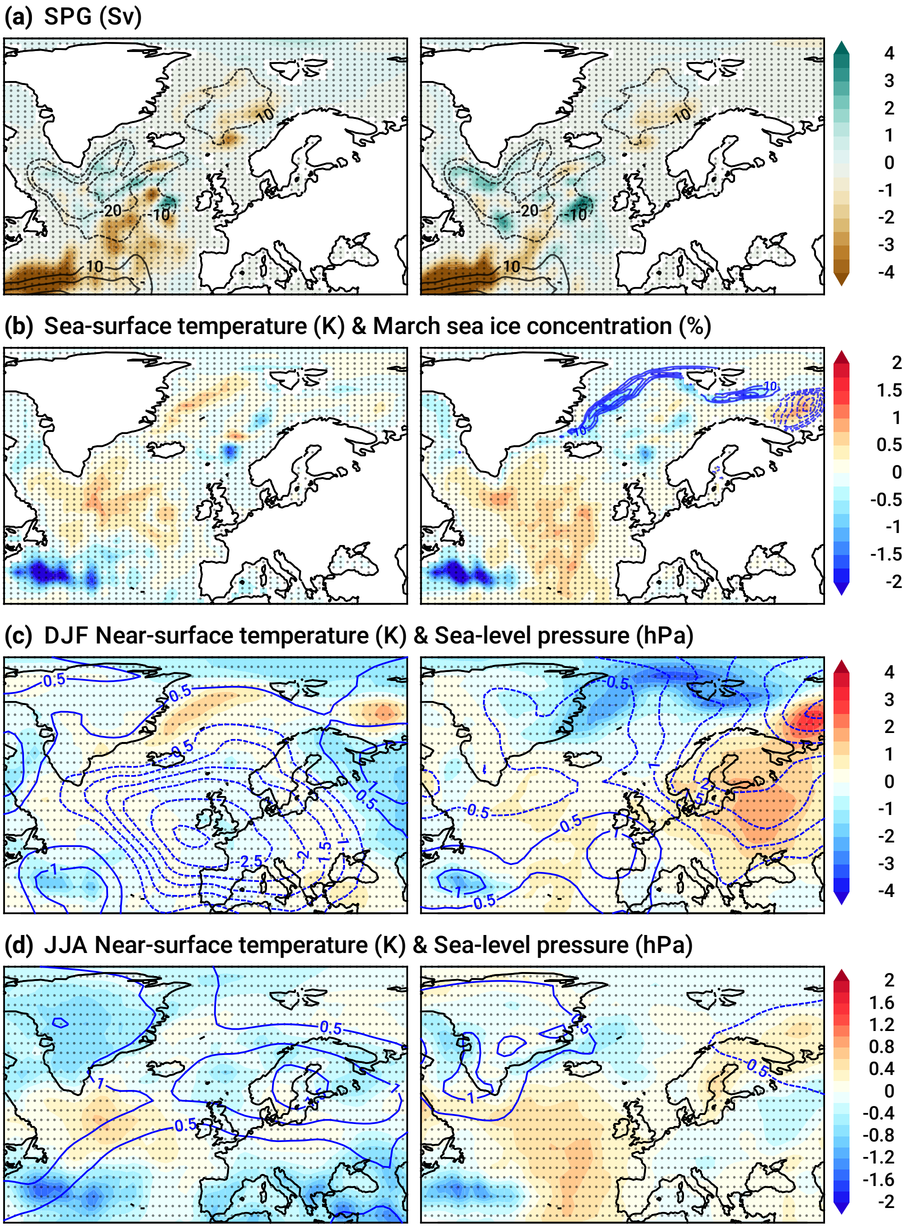
Moreno Chamarro et al. (2017b) found consistencies at multidecadal scales between simulations with a weakened SPG and reconstructed changes in several geophysical variables of the North Atlantic after ca. 1600 CE. The study did not conclude that the late 16th-century volcanic cluster was necessary for the SPG shift, which was instead mainly attributed to intrinsic variability of the simulated climate system. Sensitivity simulations of the period 1593-1650 with no volcanic forcing yielded SPG shifts similar to those in the volcanically forced simulations. Nevertheless, simulations with eruptions produced SPG shifts more frequently than those without: Of the 8 total members of the SPG-shift ensemble, 6 had volcanic forcing. Even in volcanically forced simulations, the interplay between the volcanic forcing and ongoing intrinsic climate variability determined how the eruptions shaped the SPG response. This finding agrees with the general consideration that the climatic response to eruptions depends on the background climate state (Zhong et al., 2011; Zanchettin et al., 2012, 2013; Pausata et al., 2015, 2016; Moreno-Chamarro et al., 2017a; Gagné et al., 2017).

The dependency of the response on background conditions poses a barrier to attribution of the SPG shift around 1600 CE. Two possible ways to overcome this barrier are (1) high-resolution reconstructions that better constrain the climate state evolution through the initial phases of the SPG slowdown, and (2) identification among the ensemble sensitivity simulations of a local anomaly that preceded the enhanced Arctic freshwater release and thus determined the SPG shift. For this study, a search for such an "initial seed" was performed in model variables such as sea-surface temperature, sea ice, and oceanic circulation in the sensitivity simulations; however, no evident initial conditions were identified. The precise conditions for the onset of the mechanism may be related either to stochastic phenomena or to a combination of anomalous states in different climate components, or else they may be undetectable within the limits of the small ensemble size investigated by Moreno-Chamarro et al. (2017b).

Although an initial seed was not identified, the simulations with the SPG slowdown showed short-term and long-term climatic and environmental changes with distinguishing traits, which may constrain the combinations of external forcing and climate state compatible with the SPG shift. In 1601 and 1602, these simulations show a widespread summer cooling, especially around the Mediterranean Sea, Greenland, and the eastern North Atlantic, as shown in **Figure 2**. This cooling mainly reflects the direct radiative response to the volcanic forcing, as this is still included in 6 out of the 8 members of the SPG-shift ensemble. The summer cooling is, by contrast, absent in the no-shift ensemble, which comprises mainly simulations without volcanic forcing (8 out of 12). In the first years of the 1600s, the two ensembles show minor differences in oceanic variables such as the barotropic stream function and winter sea-surface temperature in the North Atlantic, which are weakly impacted by the volcanic forcing in the short term. Larger differences in these variables between the ensembles emerge over the following decades, particularly after the 1610s, in association with the SPG slowdown, as shown in **Figure 3**. The weakening in the barotropic circulation in the subpolar North Atlantic, which represents the SPG slowdown, is accompanied by cooling and expanded sea ice in the Nordic Seas, in response to reduced northward oceanic heat transport. Colder conditions extend into the lower atmosphere especially in winter and induce an anomalous blocking-like structure over northern Europe and associated anomalous easterlies over central Europe. In contrast to winter, the SPG slowdown has a weaker impact on the summer climate over Europe.

https://doi.org/10.5194/cp-2021-82
Preprint. Discussion started: 29 July 2021

Climate of the Past
Discussions



175 **Figure 2: Short-term response (anomalies in 1601-1602 wrt 1550-1590) in the MPI-ESM-P ensemble with (left) and without (right) SPG shift. Anomalies are in (a) the annual barotropic streamfunction (in Sv; shading), with contours for the 1550-1590 mean; (b) annual sea-surface temperature (in K; shading) and March sea ice concentration (in percentage, contours every 5 %); and in near-surface (2 m) air surface temperature (in K; shading) and sea-level pressure (in hPa; contours), in (c) winter (DJF) and (d) summer (JJA). Stippling masks statistically non-significant anomalies at the 5% level based on a Student's t test.**

https://doi.org/10.5194/cp-2021-82
Preprint. Discussion started: 29 July 2021

Climate of the Past
Discussions
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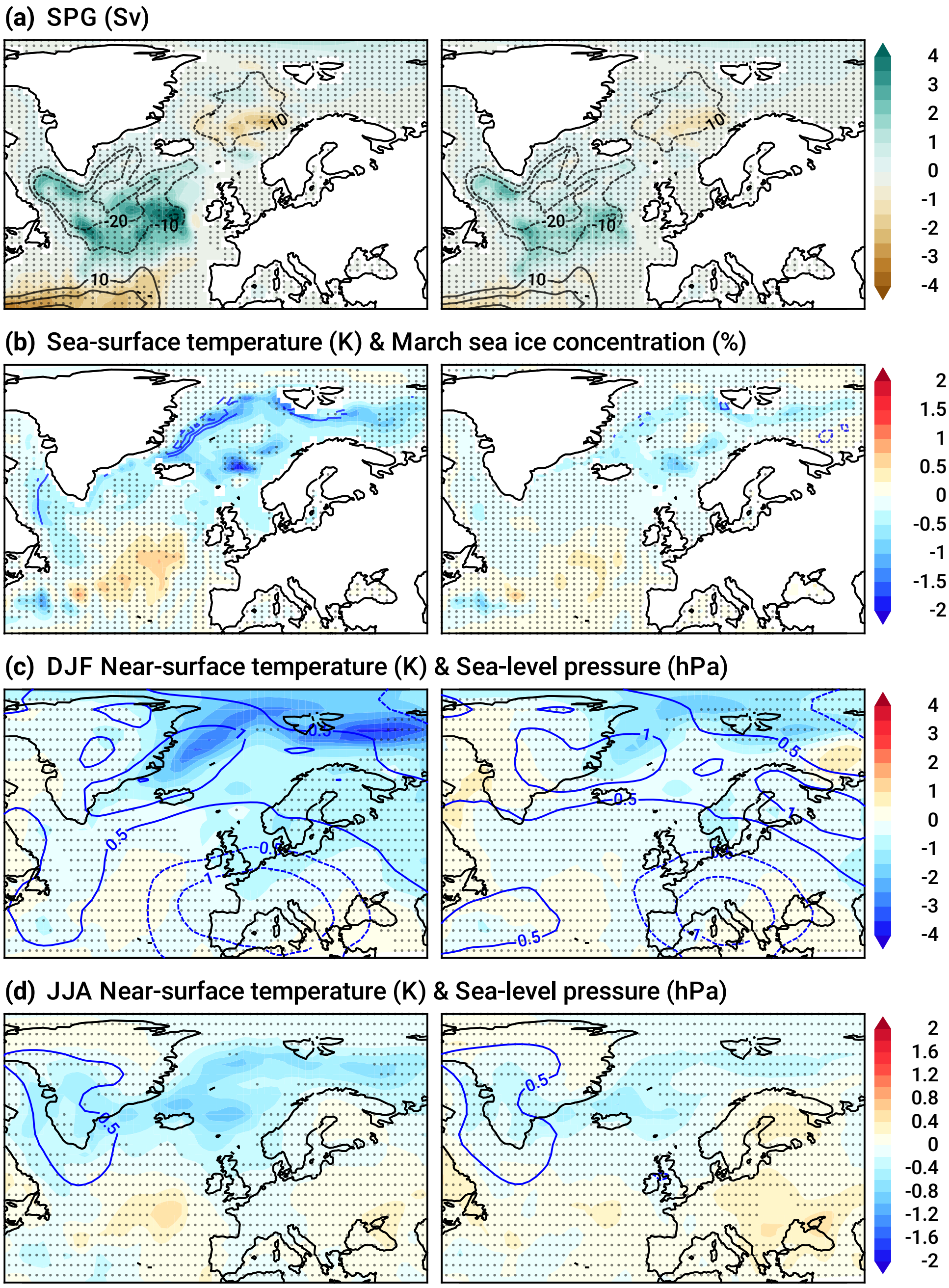


180

Figure 3: Long-term response (anomalies in 1601-1630 wrt 1550-1590) in the MPI-ESM-P ensemble with (left) and without (right) SPG shift, as in Figure 2.

https://doi.org/10.5194/cp-2021-82
Preprint. Discussion started: 29 July 2021

3. Annually Resolved Climatic and Environmental Reconstructions

Previous research on the proposed mechanism relied mainly on reconstructions from marine archives with decadal to centennial resolution to determine whether historical changes in North Atlantic sea ice extent and sea-surface temperature were compatible with a SPG shift (Moreno-Chamarro et al., 2017a,b). Although these reconstructions do indeed suggest that colder conditions and expanded sea ice prevailed over the first half of the 17th century in the North Atlantic, they lack the temporal resolution to study whether the SPG slowdown occurred through a regime shift triggered by the 1600 Huaynaputina eruption. Similarly, past studies did not distinguish the different short-term and long-term impacts in the simulations as described above. Therefore, in this study we first re-examine two high-resolution reconstructions of relevant climatic and environmental conditions: a network of tree-ring width and maximum density measurements as a proxy for summer temperatures, and the annual opening dates of ports around the Baltic as a proxy for winter sea ice and winter temperatures. We then present information from historical records containing direct observations of relevant conditions: wind directions from the North Sea coast and observations of sea-ice extent from Iceland and from ship voyages in the North Atlantic (see **Figure 4** for locations). Although still sparse, such information provides additional constraints on the timing of climate and environmental changes in and around the North Atlantic, and hence it can be used to examine whether the post-Huaynaputina eruption period marked a shift in mean conditions for variables sensitive to the SPG slowdown.

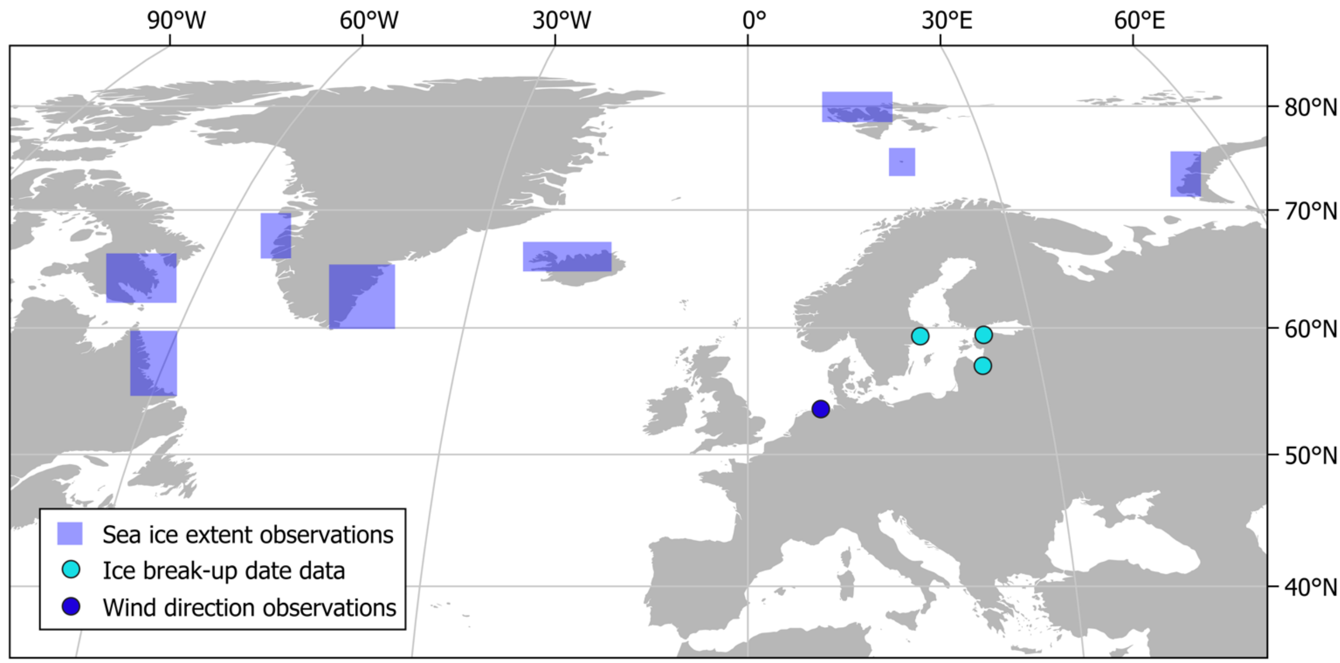


Figure 4. Location of historical observations.

3.1 Gridded Summer Temperature from a Tree-Ring Width and Density Network

https://doi.org/10.5194/cp-2021-82
Preprint. Discussion started: 29 July 2021

The reconstruction of climatic conditions and cooling induced by the 1600 Huaynaputina eruption is based on the NVOLC v2 dataset (Guillet et al., 2017, 2020) including 13 tree-ring width and 12 maximum latewood density chronologies from across the NH (see Stoffel et al., 2015; Guillet et al., 2017, for details). The characterization of regional climatic change and the illustration of spatial temperature anomalies after the 1600 Huaynaputina eruption was based on the NVOLC v2 network of tree-ring proxies and served the development of a climate field reconstruction of extratropical NH summer (JJA) temperatures as follows: In a first step, we grouped the 25 chronologies into 11 regional clusters using a correlation coefficient (r) threshold exceeding 0.3 over the period that was common to all chronologies (for details see Guillet et al., 2017, 2020). In total 3,486 grid points can thus be used for the reconstruction of the volcanic cooling induced by the Huaynaputina eruption. To examine short-term summer cooling induced by the eruption, we reconstructed the temperature anomaly in 1601-1602 with respect to the average of the period 1550-1590.

For comparison, we also employed the N-TREND (Northern Hemisphere Tree-Ring Network Development) spatial reconstruction of large-scale mean May-August temperature covering the Northern Hemisphere midlatitudes between 40° and 75˚N (Wilson et al., 2016; Anchukaitis et al., 2017). The reconstruction is based on 54 published tree-ring records and uses different parameters as proxies for temperature, including ring-width (11 records), maximum latewood density (18 records), and mixed parameters (25 records) (see Wilson et al., 2016 for details). The N-TREND version used here is version (S) detailed in Anchukaitis et al. (2017), which uses point-by-point multiple regression (Cook et al. 1994) of the tree-ring proxy records available within 1,000 to 2,000 kilometers of the center point of each 5˚ x 5˚ instrumental grid cell and a similar nesting procedure to Wilson et al. (2016). We use the average of all the grid point reconstructions for the periods where the validation reduction of error is greater than zero.

3.2 Baltic Sea Ice Concentration and Winter Temperature

In the Baltic Sea area, port towns kept records when the harbors were free from ice, marking the beginning of the sea trade season following each winter (Jevrejeva, 2001; Tarand, 1992; Tarand and Nordli, 2001). In addition to direct observations, ice break-up dates can be gained from indirect evidence such as harbor customs books, which marked the tolls paid on the date when the first ships arrived and departed after the winter (Leijonhuvfud et al., 2008; Degroot, 2018). These dates for ice break-up and the beginning of the sailing season have been identified as the proxy with the strongest reconstruction skill for Northern European winter and early spring temperatures (Hari et al., 2017); reconstructions based on this proxy can explain up to 67% of winter temperature variance (Leijonhufvud et al., 2010).

To assess short-term impacts of the 1600 Huaynaputina eruption, we compared the ice breakup data and simulations for the years 1599 and 1601, each with respect to a 1550-1598 reference period. In the model, breakup dates were defined as the first day of the year when sea-ice concentrations in the Baltic Sea (averaged between 10°E-30°E and 50°N-60°N) were lower than 1%, after smoothing with a 7-day running mean. This threshold ensures similar dates in the model and historical records, around early April. To assess whether there was a long-term shift following the Huaynaputina eruption, we looked for changepoints in the Tallinn harbour ice breakup date data for any year (with a minimum segment length of 30 years) in the

https://doi.org/10.5194/cp-2021-82
Preprint. Discussion started: 29 July 2021

period 1550-1675 using the BinSeg and PELT methods for changepoint detection (Scott and Knott, 1974; Killick et al., 2012; Killick and Eckley, 2014). The changepoint analysis could not be performed with the data from Stockholm and Riga harbors due to considerable data gaps before and after 1601.

3.3 North Sea Wind Direction

Anemometer readings are unavailable for the late 16th and early 17th centuries. While numerous voyages left information in ship logs, these do not provide a continuous series for wind direction over the North Sea until later decades of the 17th century. However, this period saw the beginnings of daily weather narratives in journals and almanacs, including some records with consistent and reliable observations of wind direction (Pfister et al., 1999; Pfister and White, 2018). Useful for the target region are the weather journals of David Fabricius, compiled in East Friesland (Germany) during 1590-1612 and reproduced and analyzed in a previous study (Lenke, 1968). These are located in flat terrain and therefore observations should reflect regional wind direction rather than local obstructions. We compared the percentage of days in each year with winds from each direction in the reconstruction and in an ensemble of simulations with and without the SPG shift.

3.4 North Atlantic Sea Ice Extent

Previous studies have utilized proxies in natural archives to reconstruct North Atlantic sea ice extent at multi-decadal to decadal resolution. In particular, studies of IP_{25}, a lipid biomarker produced by diatoms that inhabit the ice, indicate a significant expansion of spring sea-ice extent ca. 1600 on the Icelandic shelf (Massé et al., 2008; Cabedo-Sanz et al., 2016), the east Greenland shelf (Kolling et al., 2017), and in the Fram Strait (Müller et al., 2012). Historical observations, which are precisely dated, may indicate whether the shift occurred after the 1600 Huaynaputina eruption in a manner consistent with an eruption-triggered SPG shift.

This study examines three sources of sea-ice observations for the period. First, consistent observations of sea ice around the north shore of Iceland began during the late 16th century. These have been employed in several past studies as an indicator of annual sea-ice extent as well as local temperatures and societal impacts in Iceland (Ogilvie, 1995; Ogilvie and Jónsdóttir, 2000; Ogilvie and Jónsson, 2001; Ogilvie, 2019; Ogilvie, in press). Second, during the first two decades of the 17th century, walrus-hunting expeditions to Bjørnøya, sponsored by the English Muscovy Company, spurred the development of a large European whaling industry that centered on the bays of Spitsbergen. Because sailors in search of walruses and whales initially hunted from temporary encampments along the coast, expeditions usually arrived at their hunting grounds as soon as sea ice began to retreat from these bays (Degroot, 2020).

Third, an increasing number of European ships sailed into the North Atlantic and Arctic seas on voyages of exploration. Although this period predates systematic reconstruction of sea-ice extent from ships' logs, many of these voyages left detailed records, including descriptions of sea ice. Because written observations from the 16th and early 17th centuries lacked precise measurements or a standardized vocabulary for ice conditions, the most objective criterion for determining sea-ice extent is when and where ships changed course due to sea ice described as "impassable" or a similar term. For consistency, we identified

https://doi.org/10.5194/cp-2021-82
Preprint. Discussion started: 29 July 2021

common destinations and points at which ships did or not turn back based on sailing conditions, using only those locations that had records in the decades before and after 1600. Decisions to turn back a voyage could depend on a mix of factors, including cold, duration of voyage, type of ship, crew morale, and danger of sailing conditions. Therefore, we analyzed voyage journals, ship logs, and other underlying information to identify important factors in each case (see Supplement).

4. Results

4.1 Northern Hemisphere Summer Temperature

As found in previous tree ring-based reconstructions, the 1600 Huaynaputina eruption produced a sharp decline in NH summer temperatures during 1601-1602 (**Figure 5**). In fact, with a cooling of -1.6˚C in 1601 CE, the Huaynaputina eruption caused the most significant cooling recorded in the Northern Hemisphere reconstruction over the past 1500 years. The simulations and NVOLC v2 reconstruction show agreement in the magnitude of the average cooling over Europe (**Figure 6c**); however, there is little agreement in the spatial pattern of the anomaly, even in the closest simulation (**Figure 6a-b**). The reconstruction-simulation discrepancy is especially evident over Scandinavia and the Baltic region, where simulations yield much more cooling compared to the reconstruction.

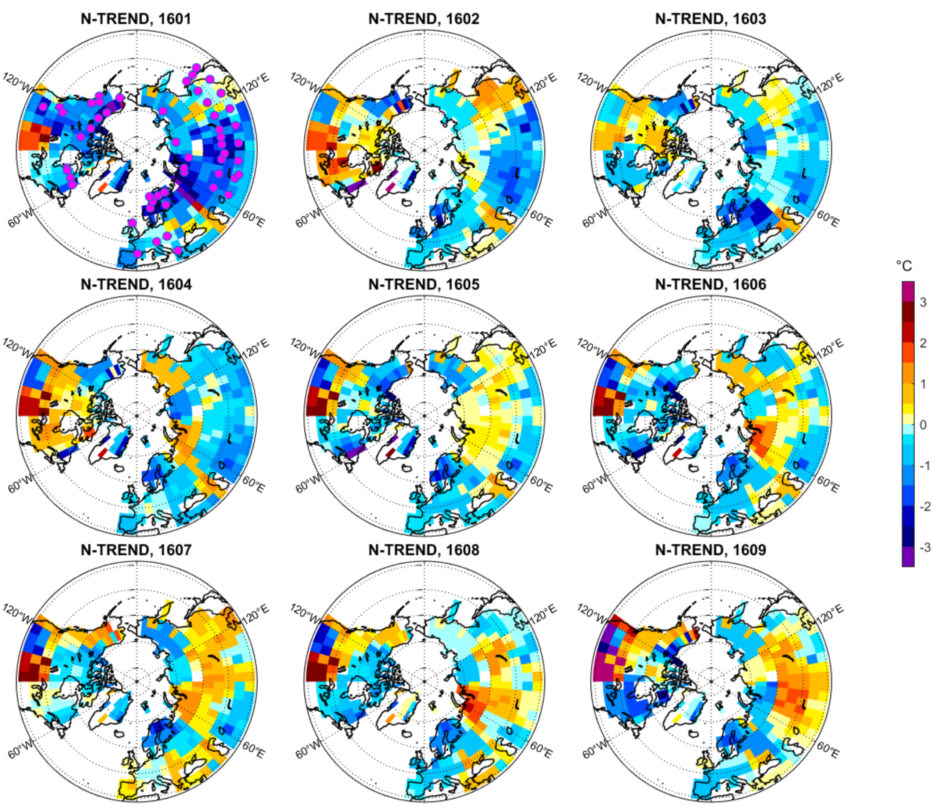
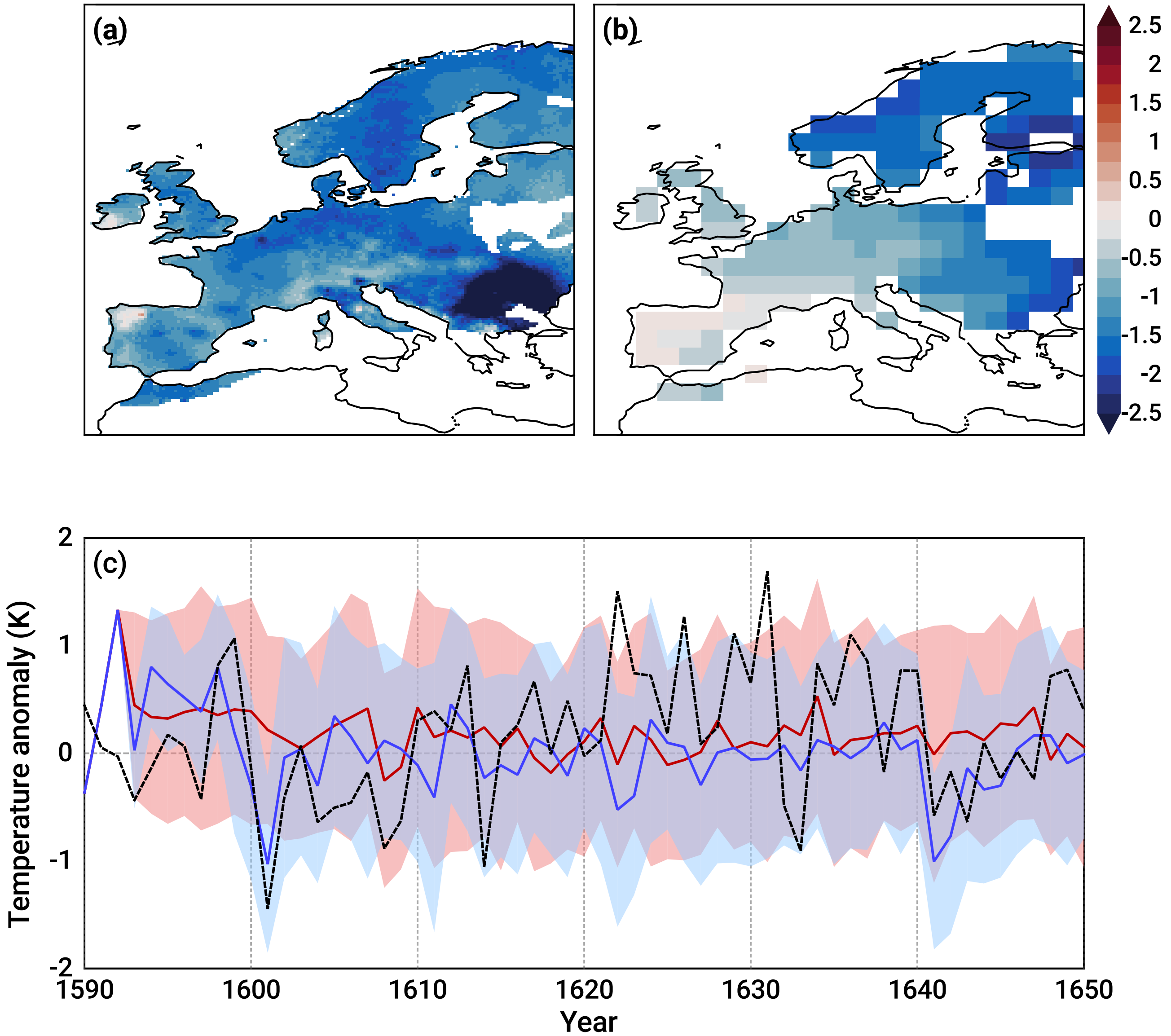


Figure 5. NTREND temperature anomaly in each year 1601-1609 with respect to the average over the period 1550-1590.

https://doi.org/10.5194/cp-2021-82
Preprint. Discussion started: 29 July 2021

Climate of the Past Discussions



285 **Figure 6: (a,b) Spatial pattern of the 1601-1602 anomaly in summer (JJA) temperature (in K; wrt 1540-1590) in (a) the NVOLC v2 reconstruction and (b) the MPI-ESM-P ensemble member r3i8, for which the spatial correlation coefficient is the largest (0.17) with the reconstruction in (a). The r3i8 is a sensitivity simulation branched in 1600 from the original simulation r3i1p1 through a slight, temporary perturbation in one atmosphere model's parameter (Moreno-Chamarro et al., 2017b). (c) Summer temperature anomaly (in K; wrt 1540-1590) of the NVOLC v2 reconstruction (black, dashed line; shifted +0.5 K) and of the ensemble with (blue) and without (red) volcanic forcing between 1593-1640 (for which thick lines are the ensemble mean and the shading is the ensemble standard deviation).**

https://doi.org/10.5194/cp-2021-82
Preprint. Discussion started: 29 July 2021

4.2 Baltic Sea-Ice Concentration and Winter Temperature

As analyzed in **Figure 7**, the ice break-up data in Riga, Stockholm, and Tallinn suggest longer "winter seasons" in 1599 and 1601 compared to the period 1550–1598, about 1–2 weeks longer in 1601 in Riga and Stockholm than in Tallinn. This may reflect the volcanic cooling over Europe after 1600. In the sensitivity simulations, however, ice break-up in the Baltic Sea in 1601 can happen at a date both earlier and later than in 1599 with and without the 1600 Huaynaputina eruption. The simulations therefore do not support a robust connection between the volcanic cooling and a later ice break-up date in these cities immediately after the eruption. This is consistent with the results described above on the simulated short-term climatic response (Figure 2), where the volcanic cooling is (on average) strongest in summer and mostly absent in winter over the region. This result also confirms that intrinsic climate variability, including climate modes such as the NAO and Arctic Oscillation (AO), plays a major role in setting winter conditions over Europe, including Baltic Sea temperatures and sea ice (Moreno-Chamarro et al., 2017b; Zanchettin et al., 2019; Chen and Hellström, 1999; Omstedt and Chen, 2001; Eriksson et al., 2007). The NAO and AO may describe internal interannual-to-decadal climate variability but are also sensitive to volcanic forcing (e.g., Timmreck, 2012; Zanchettin et al., 2012).

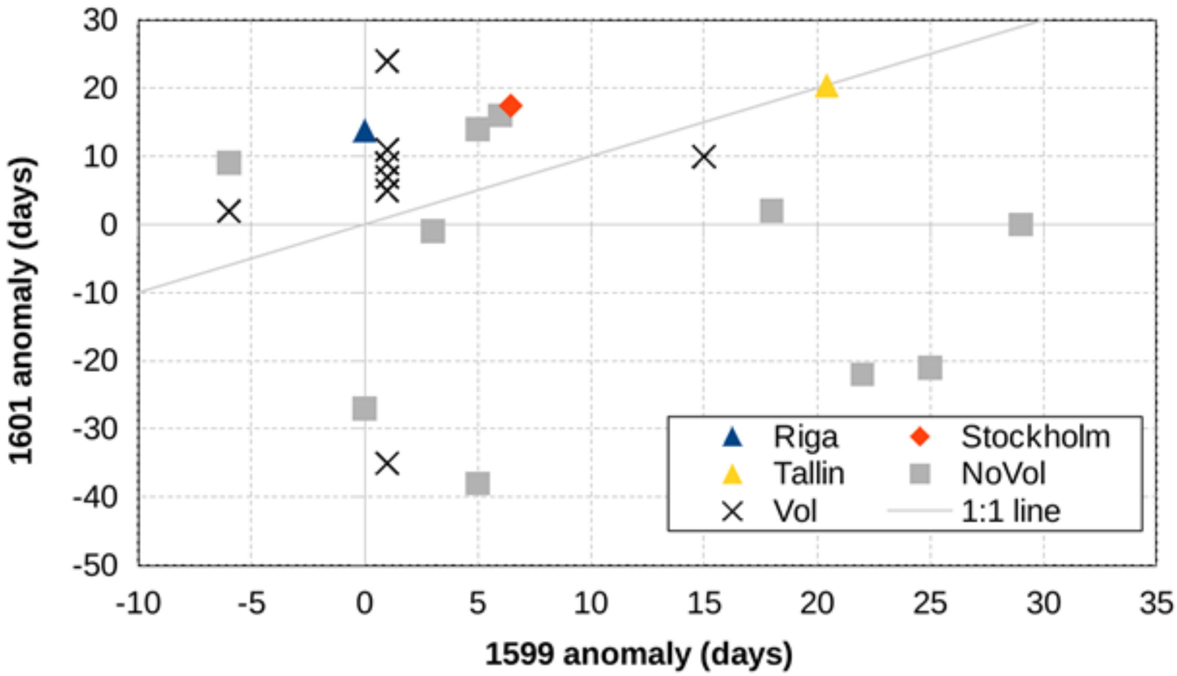


Figure 7. Anomaly in the sea ice breakup date in 1601 and 1599, wrt the 1550-1598 period. In the MPI-ESM-P simulations, breakup dates are defined as the first day of the year when sea-ice concentration in the Baltic Sea (averaged between 10˚E-30˚E and 50˚N-60˚N) is lower than 1%, after smoothing with a 7-day running mean. This threshold ensures similar dates in the model and historical records, around early April. Note that 6 simulations have the same 1599 anomaly, since the ensemble begins in 1600.

https://doi.org/10.5194/cp-2021-82
Preprint. Discussion started: 29 July 2021

Climate of the Past
Discussions
Open Access

As shown in **Figure 8**, changepoints in Tallinn harbor ice break-up dates were detected at 1597 and 1630. The change in mean of the Tallinn ice break-up data timeseries indicates colder winters and greater persistence of Baltic sea ice in the period 1597-1630 than during 1563-1596 or 1631-1664. Similarly, the winter ice severity index from the southwestern Baltic identifies a phase of increased severity between 1593 and 1630 (Koslowski and Glaser, 1999). The timing raises three at least possibilities: First, the winter cooling may have been unrelated to the SPG slowdown. Second, the cooling may have been caused by an SPG slowdown that began before the Huaynaputina eruption. Finally, there may have been other causes for cooling during the 1590s—such as the 1591 and 1595 eruptions described above—before the Huaynaputina eruption triggered an SPG slowdown and further cooling. Thus, the timing of winter cooling detected in this reconstruction could be consistent with an eruption-induced SPG slowdown but does not provide further evidence for it.

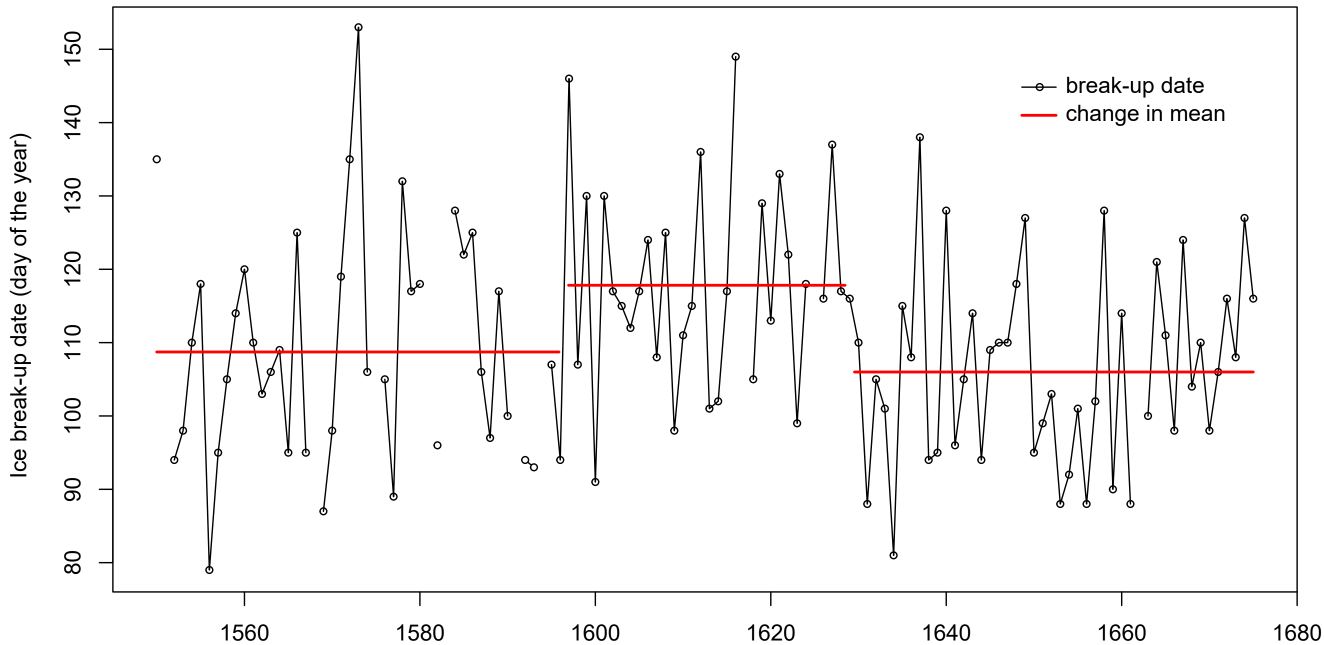


Figure 8. Tallinn ice break-up dates 1550–1670 and changepoints in mean.

4.3 North Sea Wind Direction

Observations in contemporary weather diaries indicate a marked reduction in the frequency of southwesterly winds and an increase in the frequency of easterly winds in the period 1590-1612 over the North Sea as compared to an early instrumental reference period, 1881-1925 (Lenke, 1968). The greatest change occurred during winter months. During the modern reference period the highest frequency of easterlies near East Friesland occurred in May-June, when they were predominant on only 10-15% of days. Averaged over the years during which David Fabricius kept daily observations, easterly winds were predominant on more than 20% of days during October, December, February, and March. As noted by Metzger and Tabeaud (2017),

https://doi.org/10.5194/cp-2021-82
Preprint. Discussion started: 29 July 2021

Fabricius's observations also indicate far more frequent snows and longer frosts during most of the winters in this period than those observed since the 20th century.

The year 1601 has the highest frequency of days with winds predominantly out of the north, northeast, or east. As displayed in **Figure 9**, this short-term shift toward more frequent northerlies is found in simulations with the SPG shift but not in those without it. Otherwise, neither the reconstruction data nor simulations provide a clear signal of volcanic forcing or the SPG shift in the characteristics of winter winds over the North Sea.

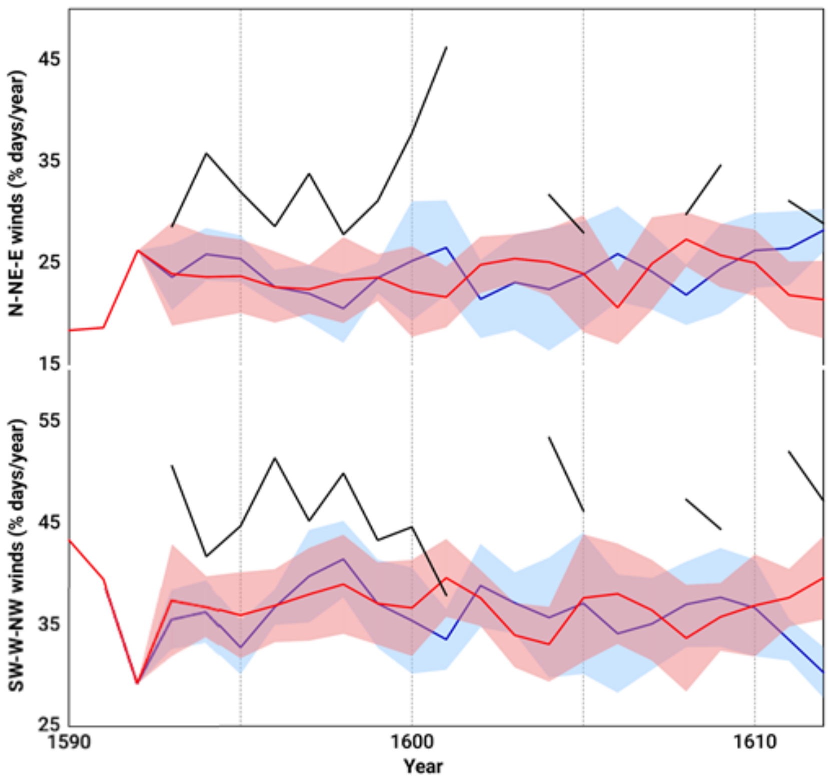


Figure 9. Percent of days in a year with winds predominantly from the North, Northeast, and East (top) and from the Southwest, West, and Northwest (bottom) in the historical data (black) and in the MPI-ESM-P ensemble of simulations with and without a SPG shift (blue and red respectively; with the ensemble mean as the solid line, and standard deviation in shading).

4.4 North Atlantic Sea Ice

Table 1 summarizes all observations that met the criteria listed in section 3.4. Where multiple sources were available for a single location and year, they agreed in all cases with the exception of mixed reports in 1612 at Bjørnøya and 1614 and 1617 around northern Spitsbergen, probably due to shifting ice conditions within the same season. The small number and heterogeneity of the observations do not permit a robust reconstruction of sea-ice conditions; however, the results appear to confirm the proxy evidence in natural archives indicating an increase in North Atlantic sea ice after ca.1600. During the 1570s-1590s several voyages were able to approach the southern tip of Greenland from the east or reach western Novaya Zemlya. Two of three could land at Labrador and four of five at Baffin Island. The Icelandic historical sources suggests that the 1560s

https://doi.org/10.5194/cp-2021-82
Preprint. Discussion started: 29 July 2021

were very cold with much sea ice but that the 1570s were mild; they provide indications that there was much sea ice off the coasts during the 1580s and years of the 1590s. From the period 1602-20, no voyages were able to land on the southern tip of Greenland, the northern Labrador coast, or Novaya Zemlya, and only two out of five reached Baffin Island. Icelandic observers did not report any years free of sea ice around Iceland and recorded several years with extensive sea ice. A few years, including 1602 and 1615, were notable for descriptions of severe sea ice in both ship observations and Icelandic records. Conditions around Spitsbergen and Bjørnøya remained variable.

Thus, results from historical observations appear to confirm a change in sea-ice conditions during the early 17th century consistent with simulations of a post-eruption SPG slowdown. However, the precise timing of the sea-ice shift and its association with a change in SPG strength remain ambiguous. On the one hand, it could be the case that post-eruption cooling generated a short-term expansion in sea ice after 1600, including the severe conditions observed in 1602, and only then did an eruption-triggered SPG slowdown initiate a long-term change in sea-ice conditions. On the other hand, the apparent increase in sea ice in 1602-20 with respect to 1570-1590s would also be consistent with a pre-eruption onset of the SPG slowdown.

Voyages of Exploration	Pre-eruption (1570-1600)		Post-eruption (1601-1620)	
	Dates passable	**Dates impassable**	**Dates passable**	**Dates impassable**
SE Greenland	6/1578, 7/1587	7/1576, su1577, 7/1585, 6-7/1586	--	6-7/1602, 6-7/1605, 6-7/1606, 6/1607, 6/1610, 5/1615
W Greenland	8/1585, 7/1586, 7/1587	9/1576	7/1605, 8/1606, 7/1616	--
Baffin	8/1576, su1577, 8/1585, 8-9/1586	7/1578	7/1610, 7/1611	7-8/1602, 7/1606, 8/1616
Labrador	su1577, 8/1586	7/1576	--	8/1602, 7/1606
W Novaya Zemlya	8/1596, 8/1597	7-8/1594	--	6-7/1608
Bjørnøya	--	6/1596	--	5/1612
Spitsbergen	--	6-7/1596	su1612	7/1607, su1614
Sea Ice Descriptions	Pre-eruption (1570-1600)		Post-eruption (1601-1620)	
	Light/no sea ice	**Heavy/late sea ice**	**Light/no sea ice**	**Heavy/late sea ice**
Iceland	1570, 1572, 1592, *early 1590s*	*1580s*	--	1602, 1604, 1605, 1608, 1612, 1615, 1618
Bjørnøya	(no data)		1603, 1608	1604, 1605, 1606, 1609
Spitsbergen	(no data)		1612, 1617, 1619	1611, 1614, 1615, 1616, 1618

Table 1. Summary of sea-ice observations from written sources. Numbers before years indicate months of year; sp[ring] = AMJ; su[mmer] = JAS. All dates converted to Gregorian calendar. Entries in italics indicate indirect observations or implied conditions based on information in chronicles. See Supplement for underlying sources and information

5. Discussion

5.1 The Huaynaputina Eruption as Possible Trigger of Persistent Winter Cooling

https://doi.org/10.5194/cp-2021-82
Preprint. Discussion started: 29 July 2021

The high-resolution reconstructions analyzed in this study exhibit several anomalies consistent with those in simulations that include both the 1600 Huaynaputina eruption and subsequent SPG slowdown. Specifically, tree-ring densities indicate strong short-term summer cooling; Baltic sea-ice records demonstrate multidecadal winter cooling and more persistent sea ice; a rising frequency of winter northeasterly winds was observed from the North Sea coast following the eruption; and Arctic voyages and Icelandic sources indicate more extensive sea ice during the first two decades of the 17th century than the final decades of the 16th century. Nevertheless, the last three sources all indicate that these anomalies began before or at ca.1600 CE, rather than developing in the years following the Huaynaputina eruption, as expected in an eruption-triggered SPG slowdown.

Therefore, the reconstructions are consistent with at least two climatic scenarios, each found in different sets of simulations. In the first scenario, the 1600 Huaynaputina eruption has triggered the SPG slowdown, but the shift to colder and icier conditions had already begun by 1600 due to intrinsic climate variability or a different externally forced mechanism. This latter possibility stems from the fact that the 1600 Huaynaputina eruption ends a volcanic cluster that starts in 1585. In the second scenario, the SPG slowdown has commenced by 1600 without any role for the Huaynaputina eruption.

Our results, although inconclusive, suggest two ways forward. First, a larger ensemble of climate simulations could improve the chances for determining whether or not the 1600 Huaynaputina eruption triggered the SPG slowdown. With a larger ensemble, it may become possible to identify an initial seed for the SPG slowdown mechanism, as the signal-to-noise ratio of emergent features increases with the ensemble size. Identification of such an initial seed in proxy reconstructions and historical observations may enable more certain identification of a SPG slowdown and its causes than the attempts to find characteristic effects of such a slowdown in this study. This type of study may require large ensembles of simulations (20 members or more) in order to detect a clear signal for the onset and evolution of a climate mechanism above the noise of interannual climate variability. Second, additional high-resolution climate proxies and historical records covering more locations in the North Atlantic could help determine whether anomalies during the late 1590s were indicative of a SPG slowdown preceding the 1600 Huaynaputina eruption.

The results also reveal strengths and weaknesses in the use of historical records to assess model-derived mechanisms of climate variability. Compared to proxies in natural archives, information drawn from archives of societies often has greater specificity and resolution but less continuity and homogeneity (Brönnimann et al., 2018). These sources may thus be more suitable for testing the presence of specific initial conditions than identifying spatial patterns in anomalies or long-term climatic trends, with the possible exception of very consistent and precise historical sources such as the Baltic harbor dates. Historical climatology may also contribute more to the analysis of climatic events and changes during the 18th and 19th centuries, for which there are more consistent and widespread historical observations than those found in earlier centuries.

5.2 Implications for the History of Climate and Society

As described in the introduction, previous studies have identified widespread mortality crises and conflict in Europe during the 1590s and early 17th century associated with cool, wet summers and resulting harvest failures. Our results confirm the

https://doi.org/10.5194/cp-2021-82
Preprint. Discussion started: 29 July 2021

occurrence of exceptionally cool summers in Europe and around the North Atlantic following the 1595 Nevado del Ruiz eruption and especially the 1600 Huaynaputina eruption. In addition, our results indicate the onset of colder winters and more sea ice in the North Atlantic by 1600 and lasting into the 1620s, thus preceding the eruption and persisting through and beyond the post-eruption summer cooling.

These long, cold and snowy winters can affect human societies in many ways, especially in more marginal areas of agriculture. In areas where cattle were kept on winter pastures, such as parts of Ireland, severe winters could kill stock and reduce births (Ludlow and Crampsie, 2018). In addition, in high-latitude and altitude agricultural areas where the animals had to be kept indoors over winter, such as Northern Europe and the Alps, farmers may have run out of fodder during prolonged winter seasons, resulting in reduced dairy production or emergency sale or slaughter of animals (Soininen, 1974; Pfister, 2005). Long snowy winters also posed a risk to winter grains, since heavy long-lasting snow cover created optimal conditions for snow mold fungi (*Microdocium nivale*) to damage seedlings wintering under the snow cover (Pfister, 2005; Solantie, 2012). Furthermore, in northernmost Europe, increased snow depth and prolonged winter delayed the start of the growing season and thus postponed the harvest to less favorable times in late August or early September, when the first autumn frosts commonly occurred (Huhtamaa et al., 2015). In Finland, contemporaries witnessed this delay in 1601. That year, in southern parts of the country, the amount of harvested grain barely exceeded the amount sown, and further north the harvest was lost altogether due to the autumn frost (Voipio, 1914; Huhtamaa et al., 2020).

The agricultural hardships of the persistent winter cooling culminated in widespread famine in 1601 in the Swedish Realm (roughly the areas of present-day Sweden, Finland and Estonia) (Lilja, 2006; Seppel, 2014; Huhtamaa, 2018). However, the cold alone does not solely explain the human suffering, since the whole northern Baltic region experienced political instability and distress at this time. King Sigismund and Duke Charles fought over the throne in the late 16th century, western Finland underwent a peasant uprising (1596-1597), and Estonia was the battleground of warfare between Sweden and Poland from 1600-1611 (Seppel, 2014; Huhtamaa, 2018). These existing conditions increased social vulnerability across the Swedish Realm and arguably exacerbated the human consequences.

Historical case studies demonstrate that these multidecadal changes, characteristic of the SPG slowdown mechanism described in this study, also featured in societal impacts and adaptations in the Arctic. Expanding sea ice, for example, shaped the possibilities for European competition in the Arctic. Extensive but fluctuating sea ice redirected expeditions led by captains such as Willem Barents and Henry Hudson towards lucrative sites for colonial exploitation, such as the bowhead whale feeding grounds off Svalbard and the Hudson River (Degroot, 2015a; Degroot, 2015b; White 2017). When European whalers competed for access to preferred whaling locations along the coast of Svalbard and the relatively nearby island of Jan Mayen, extensive sea ice discouraged conflict by either separating whalers from one another or by concentrating bowhead pods in just a few of Svalbard's many bays (Degroot, 2020). In European waters, winter sea ice suffocated seaborne trade but provided new possibilities for transportation and encouraged new transportation technologies in the densely-populated coastal regions of the Dutch Republic (Degroot, 2018).

https://doi.org/10.5194/cp-2021-82
Preprint. Discussion started: 29 July 2021

Climate of the Past Discussions

Recurring cold winters and sea-ice expansion had significant repercussions for early European colonization in eastern North America as well. Expeditions often arrived poorly supplied and vulnerable to extreme weather. Early and late frosts limited planting and harvesting of crops; long freezes and snow cover limited foraging; and early winters and late springs made for long periods without fresh food, contributing to deadly outbreaks of scurvy. Attempted settlements at Tadoussac, Quebec (1601); St. Croix, Maine (1605); and Sagadahoc, Maine (1608) were each abandoned in less than a year after experiencing extremely cold winters. The first settlers at Jamestown, Virginia (1607) and Quebec (1608) barely survived hunger and scurvy, respectively. The poor reputation of northern colonies as well as expansion of sea ice and difficult sailing conditions may have diverted interest and investments into more southern colonies during the 17th century (White, 2017; Zilberstein, 2016). During the same period some indigenous nations of today's northeastern US and eastern Canada migrated south while others adapted their hunting and horticulture to longer and colder winters (Hall, 2015; Wickman, 2018).

6. Conclusions

This study has examined high-resolution proxies and historical observations to investigate whether the 1600 Huaynaputina eruption triggered persistent cooling in the North Atlantic region by initiating a regime-shift of the North Atlantic subpolar gyre toward a persistent weak phase in the early 17th century, as shown by paleoclimate model simulations. Although the high-resolution reconstructions and historical observations of summer and winter temperature, wind direction, and sea-ice extent are consistent with such an eruption-induced mechanism, the results are inconclusive, particularly since the onset of winter cooling and increased sea ice may have preceded the Huaynaputina eruption.

By identifying potential strengths and weaknesses in data from historical climatology for the testing of model-derived climate mechanisms, our assessment may guide future research both for the specific case of the early 17th-century climate shift and for other episodes of paleoclimate variability. Our study underlines the potential of historical climatology to reconstruct highly resolved local climatic and environmental conditions relevant to studies of model-derived mechanisms. Moreover, the reconstructions and observations presented in this study have helped clarify human dimensions of climate variability during this period, including roles of short-term post-eruption anomalies as well as longer-term cooling and sea-ice expansion in societal impacts and adaptations.

Author Contributions

SW, DZ, and EM designed the study. DZ and EM provided paleoclimate simulations and analysis. MS, CC, HH, SW, and DG provided paleoclimate data and historical observations. SW, DG, and HH provided societal impacts and adaptation analysis. All authors provided figures, discussed methods and results, and commented on the manuscript.

Competing Interests

The authors declare that they have no conflict of interest.

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