

# Utilizing a Multi-Proxy to Model Comparison to Constrain the Season and Regionally Heterogeneous Impacts of the Mt. Samalas 1257 Eruption

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**Abstract.** The Mt. Samalas eruption, thought to have occurred in Summer 1257, ranks as one of the most explosive sulfur-rich eruptions of the Common Era. Despite recent convergence, several dates have been proposed for the eruption ranging between 1256-1258. As of yet, no single combination of evidence has been able to robustly distinguish between, and exclude the other dates proposed for the Mt Samalas eruption. Widespread surface cooling and hydroclimate perturbations following the eruption have been invoked as contributing to a host of 13th Century social and economic crises, although regional-scale variability in the post-eruption climate response remains uncertain. In this study we run ensemble simulations using the UK Earth System Model (UKESM1) with a range of eruption scenarios and initial conditions in order to compare our simulations with a globally-resolved multi-proxy database for the Mt. Samalas eruption, incorporating tree-rings, ice cores, and historical records. This allows precise constraints to be placed on the year and season of the Mt. Samalas eruption as well as an investigation into the regionally heterogeneous post-eruption climate response. Using a multi-proxy to model comparison, we are able to robustly distinguish between July 1257 and January 1258 eruption scenarios where the July 1257 ensemble simulation achieves considerably better agreement with spatially averaged and regionally resolved proxy surface temperature reconstructions. These reconstructions suggest the onset of significant cooling across Asia and Europe in early 1258, and thus support the plausibility of previously inferred historical connections. Model-simulated temperature anomalies also point to severe surface cooling across the Southern Hemisphere with as of yet unexplored historical implications for impacted civilizations. Model simulations of polar sulfate deposition also reveal distinct differences in the timing of ice sheet deposition between the two simulated eruption dates, although comparison of the magnitude or asymmetric deposition of sulfate aerosol remains limited by large inter-model differences and complex intra-model dependencies. Overall, the multi-proxy to model comparison employed in this study has strong potential in constraining similar uncertainties in eruption source parameters for other historical eruptions where sufficient coincident proxy records are available, although care is needed to avoid the pitfalls of model-multi proxy comparison.

## 34 **1 Introduction**

35 The Mt. Samalas eruption, which occurred on the Indonesian Island of Lombok between 1257 and 1258, is identified in ice  
36 cores as one of the largest volcanic sulfate deposition events of the last 2500 years (Palais et al., 1992, Zielinski et al., 1994,  
37 Sigl et al., 2015). Petrological analysis suggests a release of ~120Tg of sulfur dioxide (SO<sub>2</sub>) into the stratosphere, with a  
38 maximum estimated plume height of 43 km and volcanic explosivity index (VEI) of 7, meaning the eruption ranks as one of  
39 the most-explosive sulfur-rich eruptions of the Common Era (Lavigne et al., 2013, Vidal et al., 2015, Toohey and Sigl, 2017).  
40 Tree-ring reconstructions suggest a peak Northern Hemisphere (NH) average summer cooling of -0.8°C to -1.2°C between  
41 Summer 1258 and 1259 (Schneider et al., 2015, Guillet et al., 2017, Wilson et al., 2016, Büntgen et al., 2022). The surface air  
42 temperature (SAT) anomalies and potential hydroclimate perturbations induced by the eruption have been invoked by  
43 historians as contributing factors to a host of 13th Century social and economic crises (Campbell, 2017, Malawani et al., 2022,  
44 Guillet et al., 2017, Bierstedt, 2019, Stothers 2000, Green 2020, Di Cosmo et al., 2021).

45  
46 The full span of dates proposed for the Mt Samalas eruption ranges from 1256 to 1258, with suggestions including an eruption  
47 in spring 1256 (Bauch, 2019), summer 1257 (Lavigne et al., 2013, Oppenheimer 2003), and early 1258 (Stothers 2000). Whilst  
48 consensus has converged on a summer 1257 eruption date, as of yet, no single combination of evidence has been able to  
49 robustly distinguish between, and exclude other dates proposed for the Mt Samalas eruption. The suggestion of Spring 1256  
50 was based on historical evidence for a dust veil over Asia and the Middle East in late 1256 and early 1257 (Bauch, 2019). This  
51 is more likely attributed to a smaller eruption such as the 1256 Medina eruption which had only localized impacts (Saliba,  
52 2017). A mid-1257 eruption date was first proposed by Oppenheimer (2003) based on the spatial distribution of negative  
53 temperature anomalies across both hemispheres for 1257-59. Radiocarbon dating of the pyroclastic flow deposits associated  
54 with the eruption also yield a youngest eruption age boundary of 1257, with some samples being older, but no samples being  
55 younger than 1257 (Lavigne et al., 2013). Based on the westerly displacement of ash isopachs, Lavigne et al., (2013) proposed  
56 that easterly winds prevailed at the time of the eruption, indicative of the eruption occurring during the May-Oct dry season.  
57 Negative tree-ring width (TRW) growth anomalies in the late 1257-growth season (Büntgen et al., 2022) and frost rings (Salzer  
58 and Hughes, 2007) in the Western US in 1257 and 1259 also add support for a potential eruption date prior to August 1257.  
59 Modelling studies for the Mt. Samalas eruption have achieved best agreement with tree-ring reconstructions for a May-July  
60 eruption window (Stoffel et al., 2015). Nonetheless, Stothers (2000) suggests a later eruption date of early 1258 based on peak  
61 sulfate deposition for the Mt. Samalas eruption occurring in 1259 (from Hammer et al., 1980) and the first historical reports  
62 of a dust veil over Europe appearing in Summer 1258, which they suggest is most compatible with an early 1258 eruption.  
63 Therefore, there is still a need to constrain the year and season of the eruption with greater certainty, with implications for  
64 evaluating the robustness of inferred connections to synchronous historical events, as well as in the role of the eruption sulfate  
65 deposition spike as a key temporal calibration marker in ice core records.

66

67 The regionally heterogeneous climate response to the Mt. Samalas eruption also remains largely unconstrained at a global  
68 level. Guillet et al., (2017) utilized a wealth of historical records and tree-ring chronologies to assess the impact of the eruption  
69 across the NH, with particular focus on the climate response to the eruption revealed by historical sources in Western Europe.  
70 Medieval chronicles point to abnormal weather conditions in Summer 1258, with economic records highlighting delayed and  
71 poor harvests which likely aggravated ongoing grain shortages. Stothers (2000) suggested that frequently cold and rainy  
72 weather lead to widespread crop damage and famine, also noting the outbreak of plague across Europe and the Middle East in  
73 1258-59. Suggestions that the Samalas eruption can be linked to the initiation of the “Big Bang” diversification event which  
74 led to the Branch 1 strain of *Yersinia pestis* responsible for the Black Death in Europe (Fell et al., 2020) have recently been  
75 refuted, with consensus forming instead that this plague proliferation event can be traced to the Tian Shan region much earlier  
76 in the 13<sup>th</sup> century (Green, 2020, Green, 2022). Nonetheless, connections have still been drawn between the anomalous climatic  
77 conditions following the eruption and the fall of Bagdad to the Mongol empire in 1258, as well as the subsequent defeat of the  
78 Mongol Army at the battle of Ayn Jālūt in 1260 which marked the collapse of the Mongol westward advance (Green, 2020,  
79 Di Cosmo et al., 2021). Without a comprehensive understanding of the extent and chronology of climate response to the  
80 eruption on a regional-scale, the robustness of these inferred connections between post-eruption climate response and historical  
81 events remains difficult to constrain.

82

83 The magnitude and spatial distribution of the post-eruption climate response following large volcanic eruptions is known to  
84 show a strong seasonal dependency (Stevenson et al., 2017, Toohey et al., 2011). Asymmetric cooling between hemispheres  
85 occurs due to seasonal variation in Brewer-Dobson circulation which modulates hemispheric aerosol distribution (e.g Toohey  
86 et al., 2011). Asymmetric aerosol distribution combined with enhanced land-albedo feedbacks during NH winter can therefore  
87 increase the magnitude of temperature anomalies between hemispheres (Stevenson et al., 2017). Hemispheric temperature  
88 contrasts can subsequently drive latitudinal shifts in the Inter-Tropical Convergence Zone (ITCZ), an equatorial band of  
89 enhanced rainfall and lower pressure, away from the hemisphere of greatest cooling resulting in hydroclimate perturbations  
90 (Broccoli et al., 2006). Therefore, depending on if the Mt. Samalas eruption occurred in Summer 1257 or in early 1258,  
91 differences in the magnitude and spatial distribution of resulting SAT and precipitation anomalies are expected.

92

93 In this study we utilize a multi-proxy to model comparison to place more precise constraints on the year and season of the Mt.  
94 Samalas eruption, tested across the whole window of proposed eruption dates. We utilize both model and proxy constraints to  
95 assess regionally heterogeneous impacts of the Mt. Samalas eruption, with reference to proposed historical consequences. Van  
96 Dijk et al., 2023 demonstrated the effectiveness of this model multi-proxy approach in their investigation of the regional  
97 climatic and social consequences in Scandinavia following the 536/540 CE double eruption event, including incorporating  
98 additional archaeological evidence. UK Earth System Model (UKESM1) simulations were run across January and July  
99 eruption scenarios and a globally-resolved database of proxy records was collated consisting of tree-ring chronologies,  
100 historical sources, and ice core records (Supplementary Sheet 1). Our study demonstrates the ability of a multi-proxy to model

101 comparison to more precisely constrain the date of the Mt. Samalas eruption, where previous studies have tended to utilize  
102 only a single-proxy approach (Stothers, 2000, Bauch, 2019, Büntgen et al., 2022, Stoffel et al., 2015). The multi-proxy to  
103 model comparison employed in this study is shown to have significant potential in constraining similar uncertainties in eruption  
104 source parameters for other historical eruptions where sufficient coincident proxy records are available.  
105

## 106 **2 Methods**

### 107 **2.1 Model Simulations Using the UK Earth System Model (UKESM)**

108 The state-of-the-art interactive aerosol-climate model UKESM1 (Sellar et al., 2019) was used, consisting of the physical global  
109 climate model HadGEM3-GC3.1 with additional configurations for terrestrial and marine biogeochemistry, land and ocean  
110 physics, ocean-sea ice, and dynamic terrestrial vegetation. The model also includes the UK Chemistry and Aerosol (UKCA)  
111 interactive stratospheric-tropospheric chemistry and aerosol schemes (Archibald et al., 2020, Mulcahy et al., 2020). The full  
112 life cycle of stratospheric sulfur and sulfate aerosol particles is included, from injection of SO<sub>2</sub>, oxidation, particle formation  
113 and subsequent growth, to sedimentation and removal.  
114

115 The model has a horizontal atmospheric resolution of 1.875° by 1.25° and a 1° by 1° resolution in the ocean, giving a vertical  
116 resolution of 85 levels in the atmosphere and 75 levels in the ocean. This results in well-resolved ocean and atmosphere  
117 dynamics and an internally generated Quasi-Biennial Oscillation (QBO). Coupled ocean-atmosphere simulations were run  
118 with greenhouse gases set to a representative pre-industrial (AD 1850) background state. The difference between a  
119 preindustrial and bespoke 13th century background state is small compared to model internal variability and thus does not  
120 represent a significant limitation of the approach.  
121

122 Eighteen UKESM eruption-perturbed ensemble simulations were run, with nine simulating a January eruption (JAN1258) and  
123 nine a July eruption (JUL1257), where January and July are winter/summer representatives. Given the preindustrial  
124 background the January/July ensemble groupings are not constrained to a specific year and therefore the two ensemble  
125 groupings have been used to assess the full range of dates proposed for the Mt. Samalas eruption between the years 1256 and  
126 1258. The ensembles sample a range of initial conditions, with the starting phase of both QBO and the El Niño Southern  
127 Oscillation (ENSO) varying between ensembles. For full details of ensemble initial condition classification see Supplementary  
128 Document Table S1. Across the eighteen ensembles only the eruption season and initial conditions were varied, with all other  
129 eruption source parameters held constant and as listed in Table 1.  
130

131 119 Tg of SO<sub>2</sub> was taken from the updated database of VSSI estimates (eVolv2k; Toohey and Sigl, 2017), which is within  
132 error of the 126 +/- 9.6 Tg estimated by Vidal et al. (2016). In our simulations, the injection height is set at 18-20 km to be

133 consistent with the 1991 Mt. Pinatubo eruption and to allow for lofting of aerosol to higher altitudes in the stratosphere. This  
134 height is lower than the estimated 38-40 km column heights (Lavigne et al., 2013, Vidal et al., 2015) however, those column  
135 heights refer to the maximum altitude of tephra and ash rather than the height of sulfur injection in the stratosphere or the  
136 maximum altitude of the SO<sub>2</sub> plume. A 24-hr eruption duration agrees well with Lavigne et al., (2013) who estimated the  
137 eruption duration to be 23.8 +/- 10.3 hrs.

138  
139 An equivalent control ensemble, with identical starting conditions but no eruption perturbation, was run for each of the  
140 individual eruption-perturbed ensemble simulations. Anomalies were calculated with respect to a climatological background  
141 constructed from the control ensemble mean.

142

143 **Table 1: Eruption Source Parameter values used in the ensemble simulations.**

<b>Eruption Source Parameter</b>	<b>Ensemble Value</b>
Volcanic stratospheric sulfur injection (VSSI)	119 Tg of SO <sub>2</sub>
Injection Height	18-20 km
Duration	24hrs
Latitude	8°S (Single grid box)

## 144 **2.2 Multi-Proxy Database of Surface Temperature Impacts**

145 The proxy database (See Supplementary Sheets 1 and 2 for full details) was constructed by compiling records indicative of  
146 changes to climate (tree-ring chronologies, ice core records, and historical sources) that span the range of dates proposed for  
147 the eruption with a minimum of annual resolution. Whilst the aim was to create a global database of proxy records, this was  
148 significantly impeded by the limited global distribution of all types of proxy records, which show a strong bias to the NH,  
149 European, and North American localities. No suitable data was found for this study from Africa or South America.

### 150 **2.2.1 Tree-Ring Chronologies**

151 We include 24 region-specific tree-ring studies, distributed predominantly across the NH, although two studies from Australia  
152 and New Zealand are included. The parameter used for temperature reconstruction varies between studies, as does the tree  
153 species analyzed and season of reconstruction depending on study location (see Supplementary Sheet 1 for details). SAT  
154 anomalies for 1258 and 1259 were taken either directly from the referenced study or calculated from reconstructed SAT

155 anomalies using a background climatology which was constructed from the 10-year average prior to the eruption. Where  
156 studies reported that frost rings were present, they have also been included in the database.

157

158 In addition to the 24 region-specific studies, four NH spatially averaged reconstructions have also been incorporated (Wilson  
159 et al., 2016, Schneider et al., 2015, Büntgen et al., 2021, Guillet et al., 2017), along with the N-TREND reconstruction which  
160 is spatially resolved for the NH (Anchukaitis et al., 2017, see figure S3 for the spatial distribution of records used in the N-  
161 TREND reconstruction). For consistency in this study all tree-ring-reconstructed SAT anomalies are calculated from the 10-  
162 year average prior to 1257.

### 163 **2.2.2 Ice Core Records**

164 We include six  $\delta^{18}\text{O}$  isotope series ice core records from Greenland and Antarctica, where records were chosen on the basis of  
165 both annual resolution and an age dating precision of +/- 1 year. Linear regression analysis was applied to calibrate the series  
166 to JJA gridded temperature anomalies (with respect to 1990-1960) from the BEST dataset (Rodhe et al., 2020). An additional  
167 SAT constraint is also included in Greenland for Summer 1258 from analysis by Guillet et al., (2017) who utilised three  
168 Greenland ice cores at GRIP, CRETE, and DYE3 (Vinther et al., 2010) to calculate a clustered SAT anomaly for the region.  
169 Additional ice core records were investigated to expand this analysis such as the Illimani Ice Core in Bolivia and the Belukha  
170 Ice Core in Altai, Siberia; however, these records lacked the annual resolution required to constrain abrupt temperature changes  
171 associated with volcanic eruptions and/or the age dating precision to clearly identify signatures from the Samalas eruption.

### 172 **2.2.3 Historical Sources**

173 Historical sources consist predominantly of medieval chronicles which refer to abnormal and/or extreme weather events in the  
174 years 1258-59. Analysis of medieval chronicles and economic records for the years 1258-59 by Guillet et al., (2017) form the  
175 basis of historical constraints in Europe. Additional chronicles include references from the Russian Annals in the Altai  
176 Mountains (Borisenkov and Pasetsky, 1988, Guillet et al., 2017), the Chronicle of Novgorod from Central Russia (Stothers,  
177 2000), the Þorgils Saga Skarða in Iceland (Bierstedt, 2019), and Azuma Kagami from Japan (Farris, 2006). References to  
178 abnormally dark lunar eclipses are also included from the Chronicle of the Abbey of St. Edmunds (Stothers, 2000) and the  
179 Annales Ianuenses (Guillet et al., 2017). Additional historical sources across Europe, as well as some from the Middle East,  
180 report plague, famine, and economic crises for 1258-60 (Stothers, 2000). These sources have not been incorporated into the  
181 database as they refer to social and economic disturbances rather than making direct references to abnormal climatological  
182 phenomena.

## 183 **2.3 Simulated SAT Anomalies**

### 184 **2.3.1 Northern Hemisphere Average**

185 Spatially averaged NH summer land SAT anomalies were calculated for the mean of the nine JUL1257 and JAN1258  
186 ensembles respectively. Constraints were applied to model-simulated surface temperature outputs to make them most  
187 comparable to tree-ring reconstructions (Latitudes of 40°N-75°N, land only, June-July-August (JJA) was taken as  
188 representative of the growth season.) Using four NH spatially averaged tree-ring chronologies (Wilson et al., 2016, Schneider  
189 et al., 2015, Büntgen et al., 2021, Guillet et al., 2017 – see Supplementary Document, Figure S1 for individual chronology  
190 comparison) a mean NH summer SAT anomaly time series was also calculated to which the model-simulated anomalies are  
191 compared.

### 192 **2.3.2 Spatially Resolved Comparison**

193 Spatially resolved model-simulated summer (JJA) SAT anomalies were calculated globally and for the NH between 1258-  
194 1259. This time window was chosen because the JUL1257 and JAN1258 ensemble simulations show the greatest divergence  
195 in SAT anomalies for Summers 1258-59. Model-simulated NH SAT anomalies were re-gridded and masked to facilitate more  
196 direct comparison with the N-TREND dataset (Anchukaitis et al., 2017). An analysis of variance (ANOVA) test was performed  
197 by eruption season to determine at which grid points the variance in means between the 1258 and 1259 perturbed ensembles  
198 exceeded 95% significance relative to the control ensemble simulations, where the null hypothesis was that there was no  
199 difference between the mean grid point anomalies for the perturbed 1258 and 1259 ensembles compared to the control  
200 ensemble.

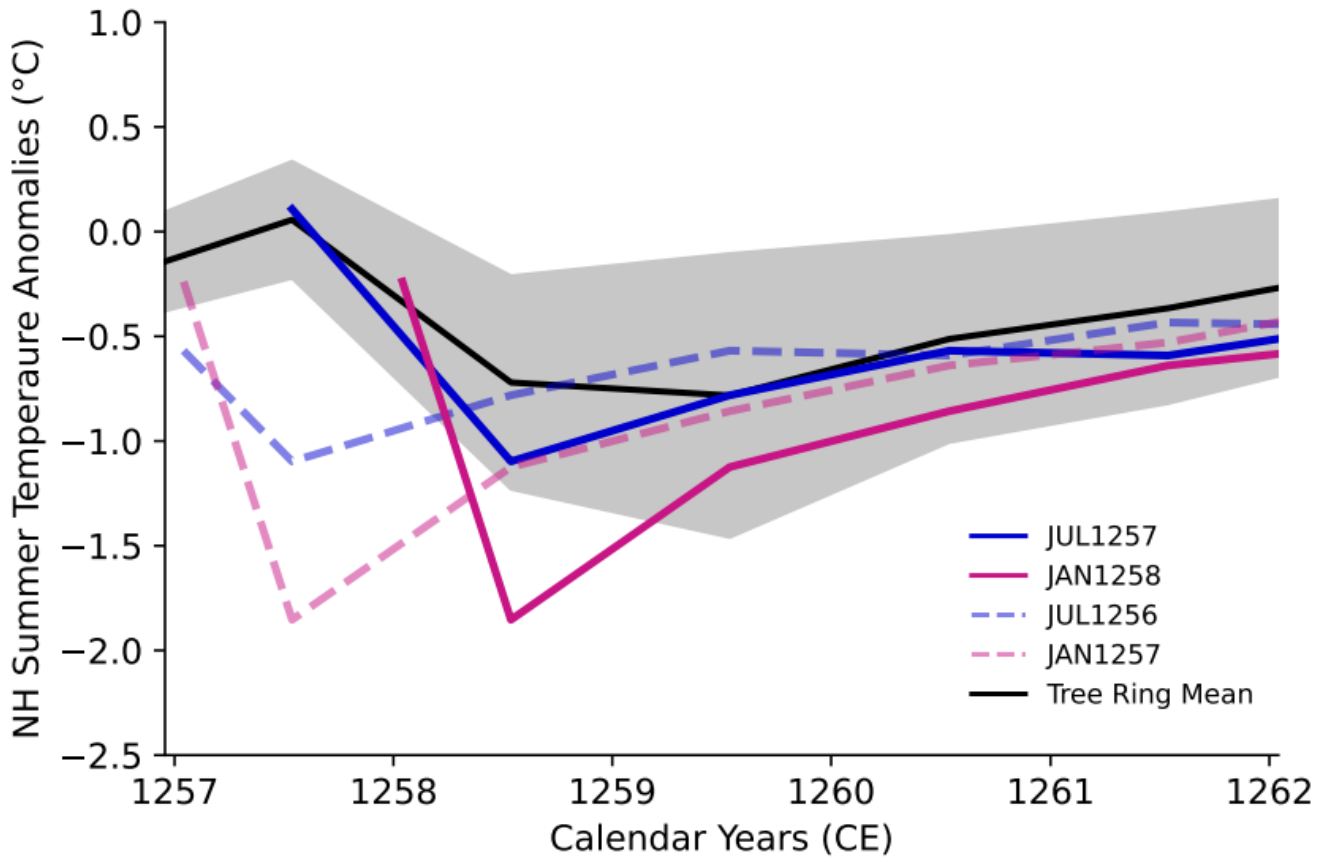
## 201 **3. Results**

### 202 **3.1 Multi-Proxy to Model Comparison**

#### 203 **3.1.1 Northern Hemisphere**

204 Figure 1 shows that the JUL1257 ensemble mean (solid blue line) is the only eruption scenario to lie consistently within  $2\sigma$  of  
205 the tree-ring mean (grey band around the black line), with good agreement with tree-ring-reconstructed anomalies for both the  
206 timing and magnitude of peak cooling across the whole period (1257-1262). The JAN1258 eruption ensemble mean (solid  
207 pink line) also results in peak cooling occurring in Summer 1258 although the magnitude of model-simulated cooling is much  
208 greater (by over 1°C) than the peak tree-ring reconstructed cooling. Across individual JAN1258 eruption ensembles (shown  
209 in Supplementary Document, Figure S2) only two lie within  $2\sigma$  of the tree-ring mean for the whole period. By contrast for the  
210 individual JUL1257 eruption ensembles seven lie within  $2\sigma$  of the tree-ring mean for the whole period (Supplementary  
211 Document, Figure S2). Both JUL1256 (dashed blue line) and JAN1257 (dashed pink line) eruption scenarios result in peak

212 cooling occurring a year early relative to the tree-ring-reconstructed mean and across both scenarios no individual ensemble  
213 members lie within  $2\sigma$  of the tree-ring mean for 1257-1260.

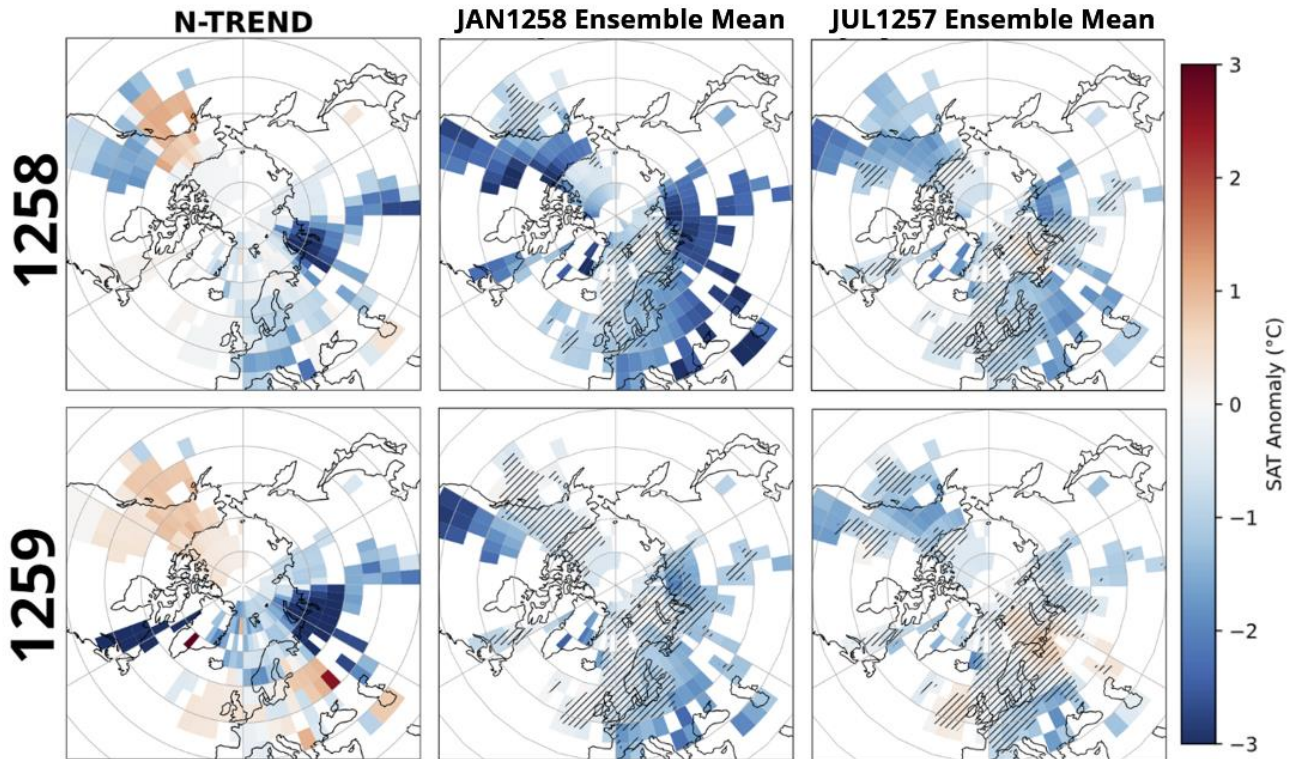


214  
215 **Figure 1. NH Summer June-July-August surface temperature anomalies. Blue: JUL1257 ensemble mean. Pink:**  
216 **JAN1258 ensemble mean. Solid and dashed lines indicate different eruption years. Black line shows the mean of the**  
217 **tree-ring-reconstructed summer surface air temperature anomalies and grey band shows  $2\sigma$  around the tree-ring mean.**  
218 **Tree-Ring data: Wilson et al., (2016), Schneider et al., (2015), Büntgen et al., (2021), Guillet et al., (2017). See**  
219 **Supplementary Sheet 2 for details of each tree-ring study.**

220  
221 When compared to the N-TREND spatially resolved tree-ring reconstructed anomalies (first column in Figure 2) the JUL1257  
222 ensemble mean (third column in Figure 2) results in more consistent agreement with the magnitude and spatial distribution of  
223 SAT anomalies across the NH for Summers 1258 and 1259. For Summer 1258 the mean grid point difference between N-  
224 TREND reconstructions and the JUL1257 Ensemble mean is +0.19 ( $\sigma = 1.08$ ) whilst the difference with the JAN1258  
225 Ensemble mean is +0.78 ( $\sigma = 0.99$ ), with the JAN1258 ensemble mean (second column in Figure 2) tending to overpredict  
226 summer SAT anomalies relative to N-TREND reconstructions. Across the US West Coast and Central and Northern Europe



227 N-TREND reconstructions suggest cooling of  $-1^{\circ}\text{C}$  to  $-2^{\circ}\text{C}$  whereas model-simulated anomalies for the JAN1258 ensemble  
 228 mean are on the order of  $-2^{\circ}\text{C}$  to  $-3^{\circ}\text{C}$ . By contrast the JUL1257 ensemble mean shows widespread but more moderate negative  
 229 SAT anomalies of  $-1^{\circ}\text{C}$  to  $-2^{\circ}\text{C}$  across the NH for Summer 1258 and thus achieves better agreement with N-TREND  
 230 reconstructions. Nonetheless, model-simulated anomalies for the JUL1257 mean suggest cooling of up to  $-2^{\circ}\text{C}$  in Central and  
 231 Northwest Asia which is an underprediction relative to N-TREND tree-ring reconstructions which suggest cooling of up to -  
 232  $3^{\circ}\text{C}$  in Summer 1258.  
 233



234  
 235 **Figure 2. Spatially resolved N-TREND-Model comparison for June-July-August average 1258-59 for the Northern**  
 236 **Hemisphere (40-90°N). N-TREND data from Anchukaitis et al., (2017). First column shows N-TREND reconstructed**  
 237 **summer surface temperature anomalies. Second column shows model-simulated summer surface temperature**  
 238 **anomalies for the JAN1258 ensemble mean and third column shows the same for the JUL1257 ensemble mean. Hashed**  
 239 **areas show regions of less than 95% significance as determined using a grid box ANOVA analysis.**

240  
 241 Greater regional variability is seen in N-TREND SAT anomalies for summer 1259. N-TREND reconstructions suggest  
 242 negative SAT anomalies in Northern Eurasia and Quebec of up to  $-3^{\circ}\text{C}$  and between  $-1^{\circ}\text{C}$  to  $-2^{\circ}\text{C}$  in Northern Europe and  
 243 Central Asia. Positive SAT anomalies of up to  $+1^{\circ}\text{C}$  are seen in Alaska and Western Europe. The JAN1258 ensemble mean

244 shows continued widespread negative anomalies of  $-1^{\circ}\text{C}$  to  $-3^{\circ}\text{C}$  across the whole NH and thus does not achieve consistent  
245 agreement with reconstructed anomalies across North and Western Europe, US West Coast or Alaska. By contrast, the  
246 JUL1257 ensemble mean shows moderate positive SAT anomalies in Northern and Western Europe of up to  $+0.5^{\circ}\text{C}$  although  
247 still somewhat under predicts the magnitude of cooling in Central and Northern Asia and the US East coast relative to N-  
248 TREND reconstructions.

249  
250 Notably neither the JUL1257 or JAN1258 ensemble mean achieves agreement with positive SAT anomalies reconstructed in  
251 Alaska for Summers 1258 and 1259. Across individual ensembles (See Supplementary Document, Figures S4-S7) only 1  
252 JUL1257 and 4 JAN1258 ensembles show positive SAT anomalies in this region. Of these ensembles three are classified as  
253 having warm phase ENSO initial conditions. N-TREND reconstructions show moderate SAT anomalies in Alaska from 1255-  
254 56, with strong positive SAT anomalies first appearing in reconstructions for Summer 1257 (shown in Supplementary  
255 Document, Figure S8).

### 256 **3.1.2 Globally-Resolved Multi-Proxy Constraints**

257 Model-simulated SAT anomalies for a JUL1257 eruption (first row) and a JAN1258 eruption (second row) across Summers  
258 (JJA) 1258 (left) and 1259 (right) are shown in Figure 3 with symbols denoting the degree of agreement with multi-proxy-  
259 reconstructed SAT anomalies. The locations and SAT anomalies constrained by proxy records are shown in Figure S3. Overall,  
260 the JUL1257 ensemble mean shows more consistent agreement with proxy SAT constraints across Europe, Asia, and North  
261 America, whilst the JAN1258 ensemble mean tends to overpredict the magnitude of negative SAT anomalies relative to  
262 quantitative proxy constraints.

263

#### 264 Summer 1258

265 For Summer 1258 large negative SAT anomalies are well constrained across Central Asia, with tree-ring reconstructions  
266 suggesting cooling in the region of up to  $-0.4^{\circ}\text{C}$  in Tibet and  $-1.1^{\circ}\text{C}$  in Mongolia (Xu et al., 2019, Davi et al., 2015, Davi et  
267 al., 2021) as well as the presence of frost rings late in the 1258 growth season (D'Arrigo et al., 2001). In Japan the “Mirror of  
268 the East” historical chronicle refers to persistent cold and wet conditions. Negative SAT anomalies are constrained across  
269 northern Russia, with cooling up to  $-2.7^{\circ}\text{C}$  (Briffa et al., 2013), and frost rings also present in the Polar Urals, Siberia (Hantemirov  
270 et al., 2004). Both JAN1258 and JUL1257 model ensembles suggest strong cooling across Central and Northern Asia, however,  
271 only the JUL1257 ensemble mean lies consistently within  $\pm 1^{\circ}\text{C}$  of the proxy constraints whilst anomalies of up to  $-4^{\circ}\text{C}$  for  
272 the JAN1258 ensemble mean are an overprediction relative to proxy constraints (see filled symbols on Figures 3C and 3E).

273

274 Across Northern, Central, and Western Europe negative SAT anomalies are constrained by a combination of tree-ring  
275 reconstructions, with moderate cooling of up to  $-0.3^{\circ}\text{C}$  in Europe (Büntgen et al., 2011), and a multitude of medieval chronicles  
276 across France, Germany, and England which refer to cold and wet conditions (Guillet et al., 2017). The JUL1257 ensemble

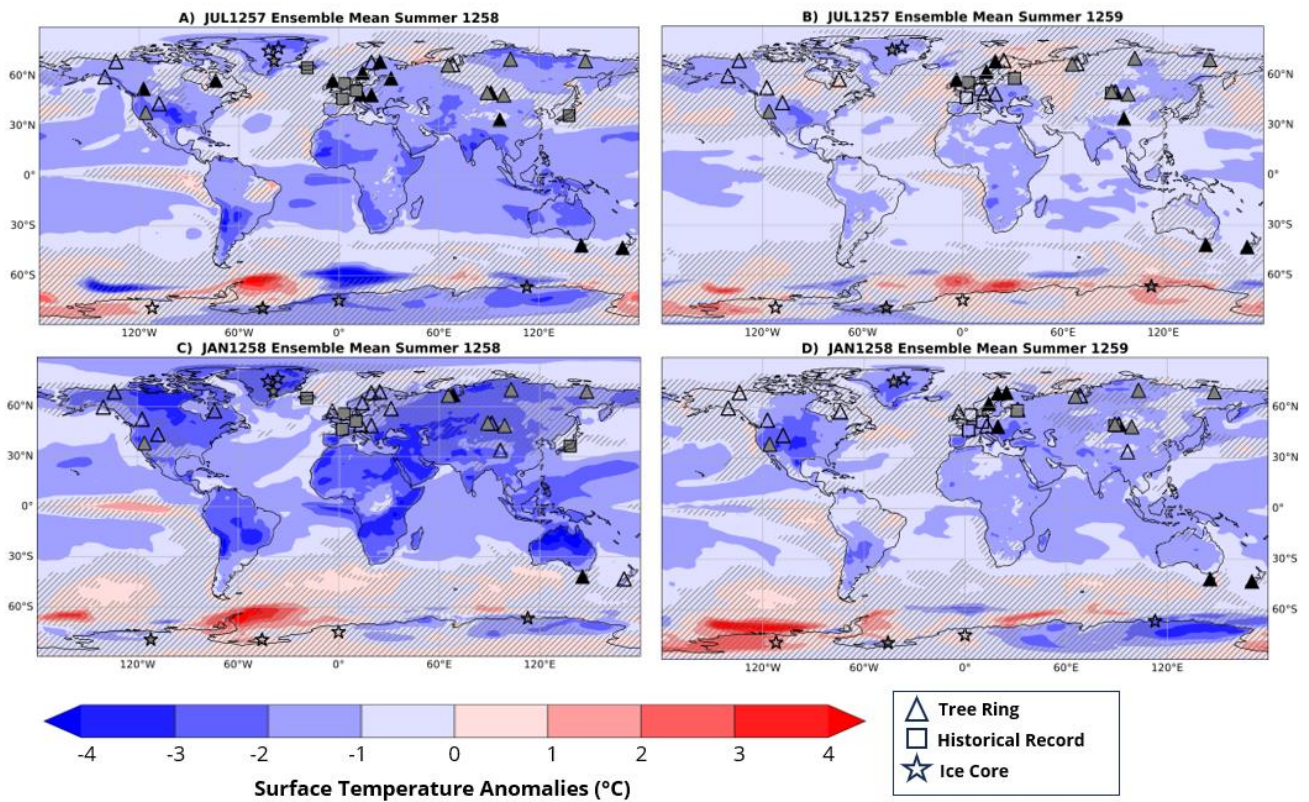
277 shows SAT anomalies of up to  $-1^{\circ}\text{C}$  in good agreement with proxy constraints on the magnitude of cooling (see filled triangles  
278 across Europe in Figure 3A), whereas the JAN1258 ensemble mean shows cooling between  $-2^{\circ}\text{C}$  and  $-3^{\circ}\text{C}$ . The Þorgils Saga  
279 Skarða in Iceland also refers to abnormally cold and wet weather during 1258 (Bierstedt, 2019) although ice core  
280 reconstructions in Greenland show only minor SAT anomalies of between  $+0.09^{\circ}\text{C}$  and  $-0.1^{\circ}\text{C}$  (Guillet et al., 2017, Vinther  
281 et al., 2010, Fischer et al., 1998, Vinther et al., 2006). JAN1258 and JUL1257 ensemble means suggest coolings of up to  $-3^{\circ}\text{C}$   
282 and  $-2^{\circ}\text{C}$  respectively and so both overpredict the magnitude of cooling across Greenland. Ice core records across Antarctica  
283 all show moderate negative SAT anomalies of up to  $-0.1^{\circ}\text{C}$ . The JUL1257 ensemble mean does show negative anomalies  
284 across three of the four ice core sites, although the magnitude of model simulated cooling ( $-2^{\circ}\text{C}$  to  $-3^{\circ}\text{C}$ ) is much greater. The  
285 JAN 1258 ensemble shows more variable SAT anomalies with both warming and cooling anomalies across Antarctica.

286

287 SAT anomalies are more variable across North America with reconstructions in Eastern Canada showing both positive and  
288 negative SAT anomalies of  $0.09^{\circ}\text{C}$  and  $-1.6^{\circ}\text{C}$  respectively (Gennaretti et al., 2014, Moore et al., 2001). The JUL1257  
289 ensemble shows agreement with both proxy constraints suggesting cooling of up to  $-2^{\circ}\text{C}$  on Baffin Island and showing no  
290 significant SAT anomaly in the Quebec region, whilst the JAN1258 ensemble mean shows cooling of up to  $-3^{\circ}\text{C}$  and thus  
291 overpredicts cooling in both regions. Tree-ring reconstructions in the Western US and Canadian Rockies suggest negative SAT  
292 anomalies of up to  $-0.5^{\circ}\text{C}$  and  $-1.6^{\circ}\text{C}$  (Martin et al., 2020, Luckman et al., 2005). The JAN1258 ensemble overpredicts the  
293 magnitude of cooling relative to proxy constraints with SAT anomalies of up to  $-3^{\circ}\text{C}$  whilst the JUL1257 ensemble mean  
294 shows more moderate anomalies of up to  $-2^{\circ}\text{C}$ . Positive SAT anomalies are well constrained by tree-ring reconstructions in  
295 the Gulf of Alaska with warming of up to  $0.1^{\circ}\text{C}$  and  $0.9^{\circ}\text{C}$  (Wiles et al., 2014), however, neither JAN1258 nor JUL1257  
296 ensemble means show positive SAT anomalies in this region.

297

298



299

300

301 **Figure 3: Globally-resolved multi-proxy-model comparison visualized for summers (JJA) 1258 and 1259. Symbols**  
 302 **denote proxy data type and red/blue shading shows model-simulated surface air temperature anomalies for JUL1257**  
 303 **eruptions (a-b) and JAN1258 eruptions (c-d) ensemble means. Surface air temperature anomalies were calculated**  
 304 **relative to a 10-year background climatology constructed from the control ensemble mean. Hatched lines denote**  
 305 **anomalies at <95% significance as determined by a grid point ANOVA analysis. Black filled symbols denote agreement**  
 306 **within +/- 1°C between model-simulated anomalies and quantitative proxy records. Grey filled symbols denote**  
 307 **qualitative agreement with proxy records. Locations and proxy-constrained SAT anomalies are shown in Figure S3.**

308

### 309 Summer 1259

310 Persistent strong negative SAT anomalies are well constrained for Central Asia with tree-ring reconstructions suggesting  
 311 continued cooling of up to -2.3°C in Mongolia and -0.3°C in Tibet (Davi et al., 2015, Xu et al., 2019) as well as frost rings  
 312 present early in the 1259 growth season (Churakova et al., 2019, D'Arrigo et al., 2001). The Russian Chronicle of Novgorod  
 313 refers to abnormal summer snowfall in the Altai mountains and unusual summer frost days (Stothers, 2000, Borisenkov and  
 314 Pasetsky, 1988) with tree-ring reconstructions showing continued negative SAT anomalies across Northern Russia of up to -  
 315 4°C (Briffa et al., 2013). Both JAN1258 and JUL1257 ensemble means show moderate negative anomalies of between -1°C

316 and  $-2^{\circ}\text{C}$  in Central Asia, however, only the JAN1258 ensemble mean shows stronger negative SAT anomalies of up to  $-3^{\circ}\text{C}$   
317 in Northern Russia whilst anomalies for the JUL1257 ensemble mean do not exceed 95% significance. Tree-ring  
318 reconstructions show some positive SAT anomalies across Central Europe of up to  $0.2^{\circ}\text{C}$  (Büntgen et al., 2011) and in Western  
319 Europe medieval chronicles refer to a hot and dry summer (Guillet et al., 2017). Very moderate SAT anomalies across Europe  
320 are only shown by the JUL1257 ensemble mean, whilst the JAN1258 ensemble continues to show cooling in the region of up  
321 to  $-2^{\circ}\text{C}$ . Ice core records in Greenland show very moderate SAT anomalies of  $0.02^{\circ}\text{C}$  (Fischer et al., 1998) and  $-0.009^{\circ}\text{C}$   
322 (Vinther et al., 2006) however, both JAN1258 and JUL1257 ensemble means show cooling across Greenland of  $-1^{\circ}\text{C}$  to  $-2^{\circ}\text{C}$ .

323  
324 Negative SAT anomalies persist across Eastern North America with cooling of up to  $-2.6^{\circ}\text{C}$  (Gennaretti et al., 2014). Cooling  
325 of up to  $-2^{\circ}\text{C}$  is shown by both ensemble means across Baffin Island (NE Canada), however, neither ensemble achieves  
326 agreement with the stronger cooling signal in Quebec. Along the Western coast of North America tree-ring reconstructions  
327 yield positive SAT anomalies of up to  $+0.8^{\circ}\text{C}$  in the Missouri River Basin and  $+0.3^{\circ}\text{C}$  in the Canadian Rockies (Martin et al.,  
328 2020, Luckman et al., 2005). Both ensembles continue to show cooling along the US West Coast although the magnitude is  
329 more moderate for the JUL1257 ensemble. In the Gulf of Alaska tree-ring reconstructions suggest continued positive SAT  
330 anomalies of up to  $+0.7^{\circ}\text{C}$  (Wiles et al., 2014). The JAN1258 ensemble mean shows no anomalies exceeding 95% significance  
331 in Alaska whilst the JUL1257 ensemble mean shows continued cooling of up to  $-2^{\circ}\text{C}$  and so neither eruption scenario achieves  
332 good agreement with proxy constraints. Ice core records across Antarctica continue to show moderate cooling anomalies of up  
333 to  $-0.15^{\circ}\text{C}$  apart from the Law Dome (Plummer et al., 2012 -  $112^{\circ}\text{E}$ ) core which shows a moderate warming anomaly of  $+0.16$ .  
334 The JUL1257 ensemble mean does show a warming anomaly in the east of Antarctica, however cooling anomalies are more  
335 isolated in west Antarctica compared to those reconstructed from ice core records. The JAN1258 ensemble mean shows poor  
336 agreement with ice core records with warm SAT anomalies in the east of Antarctica and cool SAT anomalies in the west.

### 337 **3.2 Re-evaluating Evidence for a January 1258 Eruption Date**

338 Two lines of evidence have previously been invoked to support an early 1258 eruption date: references in medieval chronicles  
339 to a “dark lunar eclipse” in mid-May 1258 (Stothers, 2000) and peak sulfate fall out in Greenland ice cores in early 1259  
340 (Hammer et al., 1980). Figure 4a shows model-simulated Stratospheric Aerosol Optical Depth (SAOD) averaged across  
341 western Europe for the years following the eruption with the black line showing the eVolv2k SAOD for the PMIP4 forcing  
342 which is based on a July 1257 eruption date (Toohey & Sigl, 2017). Marked by the vertical grey lines are the dark lunar eclipses  
343 of 18-19th May 1258 and 12-13th Nov 1258 identified by Stothers (2000) in the Bury Saint Edmunds Abbey chronicle and  
344 Guillet et al., (2017) in the Annales Ianuenses respectively, where for the moon to appear dark an SAOD  $> 0.1$  is needed  
345 (Stothers, 2000). Whilst a JAN1258 eruption scenario does result in a later SAOD peak, both a JUL1257 and a JAN1258  
346 eruption result in SAOD  $\gg 0.1$  during both May and November 1258 and therefore either eruption scenario could account for  
347 observations of darkened lunar eclipses. This is supported by findings from Guillet et al., (2023) that a dark total lunar eclipse  
348 is most likely to be observed 3 to 20 months following an explosive eruption with a VSSI exceeding 10 Tg. SAOD in the

349 UKESM for both JUL1257 and JAN1258 ensembles decays more rapidly than the eVol2k SAOD used for the PMIP4 forcing,  
 350 for which SAOD remains > 0.1 until mid-1260 AD..

351

352 **Table 2: Comparison between ice core average (from Table S2) and ensemble means for JUL1257 and**  
 353 **JAN1258 model simulations for the timing of SO<sub>4</sub> rise, peak, and total deposition in Greenland and**  
 354 **Antarctica.**

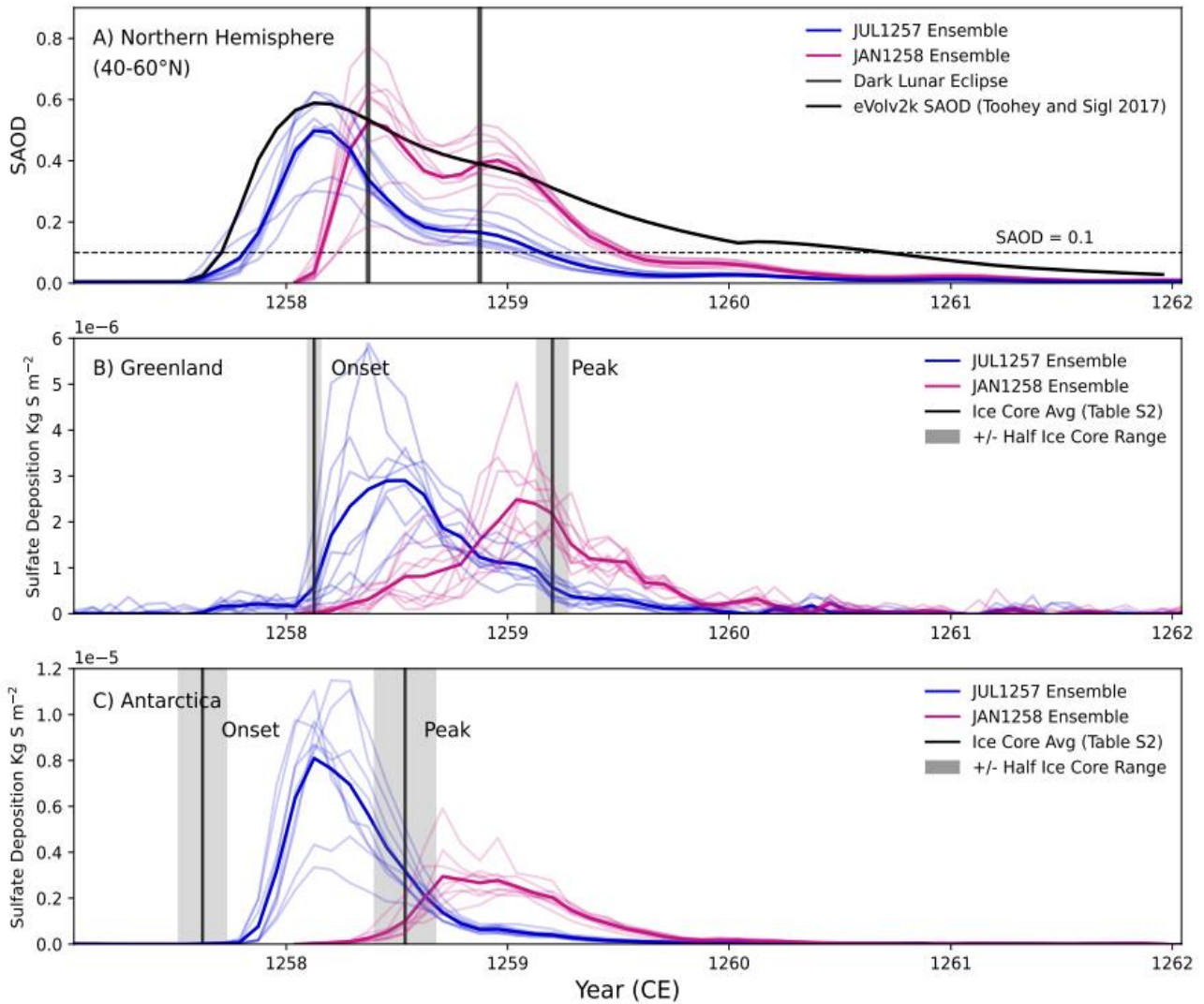
	<b>JUL1257 Ensemble Mean</b>	<b>JAN1258 Ensemble Mean</b>	<b>Ice Core Mean (Average from Table S2)</b>
SO <sub>4</sub> Rise Greenland	Jan 1258	May 1258	February 1258
SO <sub>4</sub> Rise Antarctica	Oct 1257	March 1258	August 1257
<b>Offset in SO<sub>4</sub> Rise (months, Antarctica - Greenland)</b>	<b>3</b>	<b>2</b>	<b>6</b>
SO <sub>4</sub> Peak Greenland	July 1258	Jan 1259	March 1259
SO <sub>4</sub> Peak Antarctica	February 1258	Sept 1258	July 1258
<b>Offset in SO<sub>4</sub> Peak (months, Antarctica - Greenland)</b>	<b>5</b>	<b>4</b>	<b>8</b>
Greenland Sulfate Deposition (Kg km <sup>-2</sup> ) (Toohey & Sigl 2017)	<b>26.8</b>	<b>23.2</b>	105
Antarctica Sulfate Deposition (Kg km <sup>-2</sup> ) (Toohey & Sigl 2017)	<b>54.3</b>	<b>28.0</b>	73
<b>Greenland/Antarctica</b>	<b>0.49</b>	<b>0.83</b>	<b>1.4</b>

355

356

357 Figures 4B and 4C show ice sheet-averaged model-simulated sulfate deposition across Greenland and Antarctic ice sheets  
 358 respectively, for JUL1257 and JAN1258 eruption scenarios. The mean timing of the onset of sulfate rise and the timing of  
 359 peak sulfate deposition across four high resolution ice cores (see Table S2) are also shown as vertical black lines. Table 2  
 360 shows a comparison between ice core average (from Table S2) and ensemble means for JUL1257 and JAN1258 model  
 361 simulations for the timing of SO<sub>4</sub> rise, peak, and total deposition in Greenland and Antarctica. The JUL1257 ensemble shows

362 that the onset of sulfate deposition occurs in January 1285 in Greenland and October 1257 in Antarctica, with peak deposition  
363 occurring in July 1258 and February 1258 respectively. By contrast, the JAN1258 ensemble shows sulfate rise beginning in  
364 May and March 1258 in Greenland and Antarctica respectively, with peak deposition in Jan 1259 and September 1258. Across  
365 four high resolution ice cores (Table S2) the mean timing of the onset of sulfate deposition occurs in February 1258 in  
366 Greenland and August 1257 in Antarctica, thus showing closest agreement with the timing of the simulated onset in sulfate  
367 deposition for a July 1257 eruption date, where only an eruption in summer 1257 can account for the beginning of sulfate  
368 deposition in Antarctica in autumn 1257. Across the four ice core records mean peak deposition in Greenland occurs in March  
369 1259 whilst mean peak deposition in Antarctica occurs earlier in July 1258. The timing of peak deposition therefore shows  
370 better agreement with simulated peak deposition for a January 1258 eruption date with peak model-simulated deposition  
371 occurring too early in the JUL1257 ensemble relative to ice core records. Ice core records also suggest an asymmetry in sulfate  
372 dispersal and deposition, with the onset of sulfate deposition and peak deposition in Antarctica being 6 and 8 months ahead of  
373 Greenland respectively. The JUL1257 ensemble shows a greater degree of asymmetry compared to the JAN1258 ensemble,  
374 although offsets of 3 and 5 months between Antarctica and Greenland are still lower than the offset suggested by the ice core  
375 record. Whilst the offset between the onset in sulfate deposition and peak sulfate deposition is approximately a year in both  
376 Greenland (13 months) and Antarctica (11 months), the offset for model-simulated deposition in the UKESM is considerably  
377 shorter across both JUL1257 and JAN1258 ensembles (ranging from 4-8 months). This may represent a limitation specific to  
378 the UKESM, for example with too weak poleward transport of volcanic aerosol (or midlatitude deposition too strong). This  
379 weaker poleward transport may also contribute to the lower magnitude of total polar deposition simulated by the UKESM  
380 relative to ice core records, also seen in an earlier version of the UK climate model for the 1815 eruption of Mt. Tambora  
381 (Marshall et al., 2018). Both model-simulated ensembles show greater total sulfate deposition in Antarctica compared to  
382 Greenland, whilst ice core records suggest the opposite asymmetry favouring greater deposition in Greenland.



383  
 384 **Figure 4: A) Model-simulated Stratospheric Aerosol Optical Depth (SAOD) timeseries averaged across Western**  
 385 **Europe (Lat: 40-60°N, Longitude: 10W-10°E) for JUL1257 (blue line) and JAN1258 (pink line) eruption scenarios**  
 386 **where bold lines are the ensemble means. Vertical grey bars denote historical records of dark lunar eclipses in May**  
 387 **1258 (England) and November 1258 (Genoa). The dashed horizontal line at SAOD = 0.1 denotes the minimum SAOD**  
 388 **required for a dark lunar eclipse. B-C) Model-simulated sulfate deposition (kg S m<sup>-2</sup>) over Greenland and Antarctica**  
 389 **ice sheets for JUL1257 (Blue) and JAN1258 (Pink) eruption scenarios. The region of deposition is limited to the single**  
 390 **model grid box containing NEEM (Greenland) and WDC (Antarctica) ice core drill sites respectively. Black lines are**  
 391 **timeseries from NEEM and WDC ice core records respectively (Sigl et al., 2015) with grey bars showing +/- 1-year**  
 392 **uncertainty for the timing of peak deposition.**



## 393 4. Discussion

### 394 4.1 Constraining the Eruption Year and Season of the Mt. Samalas eruption

395 As shown in Figure 1, both JUL1256 and JAN1257 eruption scenarios result in peak surface cooling occurring a year too early  
396 relative to tree-ring reconstructed SAT anomalies. Thus, an early eruption date such as Spring 1256, as proposed by Bauch  
397 (2019), is, based on our analysis, unfeasible for the Mt. Samalas eruption. Across the two remaining eruption scenarios only  
398 the JUL1257 eruption ensemble lies consistently within  $2\sigma$  of the tree ring-reconstructed mean between 1258-1262. By  
399 contrast, the JAN1258 eruption ensemble results in peak cooling being over  $1^{\circ}\text{C}$  greater than tree ring reconstructions for  
400 Summer 1258.

401  
402 When compared with spatially resolved N-TREND reconstructions model-simulated SAT anomalies for the JAN1258  
403 ensemble mean continue to consistently overpredict the magnitude of cooling for both 1258 and 1259 across the NH (Figure  
404 2). Whilst the spatial agreement of SAT anomalies for the JUL1257 ensemble mean does not precisely replicate those  
405 reconstructed from the N-TREND tree ring record it does show widespread but less extreme negative SAT anomalies across  
406 the NH and thus achieves better agreement with reconstructed anomalies on a regional basis. This is supported by comparison  
407 between globally-resolved model-simulated ensemble mean SAT anomalies and multi-proxy-reconstructed anomalies where  
408 the JAN1258 ensemble mean consistently over predicts the magnitude of cooling in key regions across Europe, Central Asia,  
409 and the US, whilst the JUL1257 ensemble mean results in generally better agreement across these regions with more moderate  
410 negative SAT anomalies.

411  
412 A re-evaluation of evidence for a January 1258 eruption finds that a Summer 1257 eruption can also satisfy constraints from  
413 both ice core records and historically documented dark lunar eclipses. Model-simulated ice sheet sulfate deposition shows that  
414 for eruption scenarios 6 months apart there is a distinguishable off-set in the timing and magnitude of deposition, with the bi-  
415 polar Greenland/Antarctica ratio of simulated sulfate depositions between northern and southern hemisphere ice sheets varying  
416 depending on eruption season. The onset of model-simulated sulfate deposition in Greenland and Antarctica for the JUL1257  
417 ensemble mean achieves better agreement with ice core record tie points, where the onset of sulfate deposition in Antarctica  
418 in August 1257 can only be aligned with an eruption during or prior to Summer 1257. Nonetheless, the timing of peak model-  
419 simulated sulfate deposition achieves better agreement with the JAN1258 ensemble mean, potentially reflecting limitations in  
420 aerosol transport within the UKESM leading simulated aerosol decay being too-rapid (Marshall et al., 2018).

421  
422 Overall, the multi-proxy to model comparison utilized in this study provides a clear distinction between JULY1257 and  
423 JAN1258 eruption scenarios, with better agreement between proxy reconstructions and model-simulated anomalies being  
424 shown for a July 1257 eruption date. This is consistent with the May-August 1257 date constraint suggested by Guillet et al.,  
425 (2023) based on their analysis of contemporary reports of total lunar eclipses, combined with tree ring-based climate proxies

426 and aerosol model simulations. Although consensus has converged on a Summer 1257 date for the Samalas eruption, it remains  
427 to be seen if a more precise constraint (i.e. to a specific month) could be achieved given current model and proxy uncertainties  
428 (as discussed in Section 4.3 below). Nonetheless, this four-month window remains an improvement upon previous dating  
429 uncertainty and is still sufficient at present for interrogating the both climatic and human consequences following the eruption.

430

#### 431 **4.2 Regionally Heterogenous Climate Response**

432

433 Multi-Proxy SAT reconstructions highlight the regionally heterogenous climate response following the Mt. Samalas eruption.  
434 The largest negative SAT anomalies occur across Central Asia and Northern Russia, with cooling of  $-2^{\circ}\text{C}$  to  $-4^{\circ}\text{C}$  between  
435 1258 and 1259, making these as some of the most severely impacted regions in the NH. The role of sudden and severe cooling  
436 associated with the Mt. Samalas eruption in the collapse of the Mongol westward advance is therefore plausible (Di Cosmo et  
437 al., 2021). Alongside references to extreme and abnormal weather conditions, the Azuma Kagami in Japan also highlights the  
438 severity of the Shôga famine between 1257-60 (Farris, 2006). Model-simulated cooling of up to  $-1^{\circ}\text{C}$  across Japan suggest the  
439 severity of this famine could have been amplified by the climate response to the Mt. Samalas eruption in 1258-59. Similar  
440 evidence exists in the Middle East with famine and pestilence reported across Syria, Iraq, and Southern Turkey in 1258  
441 (Stothers, 2000). A model-simulated JUL1257 eruption scenario suggests cooling of up to  $-2^{\circ}\text{C}$  in the region and thus supports  
442 a possible association with the Mt. Samalas eruption. Reconstructed and model-simulated SAT anomalies suggest less severe  
443 cooling across Europe for 1258 ( $<-1^{\circ}\text{C}$ ) and relative warming in the region for 1259. Nonetheless, this is still associated with  
444 significant economic and social disturbances, with historical records reporting famine and social unrest (Guillet et al., 2017,  
445 Stothers, 2000). Van Dijk et al., (2023) found similar regional variability in both the climatic effects and social consequences  
446 across Scandinavia following the 536/540 double eruption event, finding the severity of impacts depended heavily both on  
447 local topography and the subsistence methods employed by different communities.

448

449 Büntgen et al., (2022) found negative TRW growth anomalies along the US West Coast for Summer 1257 and suggest this is  
450 evidence for the onset of climate perturbations in the NH before the end of the 1257 growth season and therefore support a  
451 Summer 1257 eruption date. Model-simulated anomalies for a JUL1257 eruption, however, show no significant anomalies  
452 occurring across the US or NH during Summer 1257 (see Supplementary Document, Figure S9). The presence of frost rings  
453 in 1257 and 1259 (Salzer and Hughes, 2007), but not in 1258, also contradicts model-simulated cooling, which consistently  
454 shows the strongest cooling in 1258. Model-simulated SAT anomalies for a JUL1257 eruption only show significant negative  
455 SAT anomalies in late summer 1257 across South America, Africa, and Oceania, with cooling in some regions of up to  $-2^{\circ}\text{C}$ ,  
456 although without well-dated, geographically distributed climate proxies in the SH it will remain difficult to resolve whether  
457 this is potential model/proxy discrepancy in the SH (Neukom et al., 2014). However, if extreme and sudden cooling did occur  
458 in the SH would be expected to have significant, but as yet unknown, consequences for communities and civilizations in the  
459 Southern Hemisphere. Additional large volcanic eruptions in 1269, 1276 (UE5), and 1286 (UE6), which combined with the

460 eruption in 1257, make the sulfate loading in the 13th century two to ten times larger than any other century in the last 1500  
461 years (Gao et al., 2008, Guillet et al., 2023), and may have led to further climate anomalies with effects on impacted  
462 civilisations being prolonged throughout the latter half of the 13<sup>th</sup> century. For example, the first settlement of New Zealand  
463 most likely occurred between 1250–1275, with suggestions this may have reflected a climate-induced migration associated at  
464 least in part with the impacts of the Mt. Samalas, and subsequent, eruptions (Anderson, 2016, Bunbury et al., 2022). Apart  
465 from recent analysis of the localized impacts and recovery following the Mt. Samalas eruption (Malawani et al., 2022) the  
466 general sparsity of currently available proxy data across the SH precludes definitive conclusions as to climate and social  
467 response in the SH following the Mt. Samalas eruption.

468

469 Notably, neither the JAN1258 or JUL1257 ensemble mean replicates reconstructed positive SAT anomalies in Alaska for  
470 Summer 1258 or the wider US west coast for Summer 1259. Across individual realisations only five show positive SAT  
471 anomalies in this region for 1258-59, and of these ensembles 3 are classified as having warm phase ENSO initial conditions.  
472 N-TREND spatially resolved reconstructions show positive SAT anomalies over Alaska from Summer 1257-59, with warm  
473 phase El Niño-like conditions during this period being supported by positive tree ring-reconstructed temperature anomalies in  
474 Alaska (Guillet et al., 2017) and the muted absolute temperature signal seen in  $\delta^{18}\text{O}$  coral reconstructions from the central  
475 tropical Pacific (most likely explained by the superposition of a volcanic cooling signal and a warm El Niño-like signal (Dee  
476 et al. 2020, Robock 2020)). The role of volcanic eruptions in perturbing ENSO remains debated (Mann et al. 2005, Stevenson  
477 et al. 2018, Dee et al. 2020, Robock 2020) although climate simulations and proxy records suggest an increased probability  
478 for the occurrence of an El Niño event in the first or second year after a large volcanic eruption (see McGregor et al 2020 for  
479 a full review). El Niño-like warm conditions may therefore have prevailed at the time of the Mt. Samalas eruption with  
480 subsequent volcano-ENSO interactions potentially acting to enhance these pre-existing El Niño-like conditions.

### 481 **4.3 Model and Proxy Limitations**

482 The majority of temperature anomaly constraints applied in this study are provided by SAT tree ring reconstructions, where  
483 tree-ring data has been shown to effectively record extreme cooling events synchronous with evidence for explosive volcanic  
484 eruptions over the last two millennia without chronological errors (Stoffel et al., 2015, Büntgen et al., 2020; Büntgen et al.,  
485 2016; Sigl et al., 2015) and in good agreement with instrumental measurements following large eruptions (Esper et al., 2013).  
486 Lücke et al., (2019) highlight the importance of accounting for biologically based memory effects, which can lead to  
487 dampening of volcanic cooling signals in ring width-based chronologies. However, many of the tree ring-based reconstructions  
488 utilised in this work do incorporate maximum latewood density (MXD) data as well as TRW (see Figure S3) which has been  
489 shown to reduce attenuation of volcanic cooling signals (Esper et al., 2014, Stoffel et al., 2015). Additional temperature  
490 anomaly constraints are also provided by historical sources, although these records are inherently more limited being non-  
491 quantitative and subjective accounts. The application of historical records to the model-multi proxy framework is limited by  
492 the sparsity of historical records, both spatially, being predominantly concentrated in the NH, and temporally, being biased in

493 frequency, dating accuracy, and traceability towards more recent volcanic eruptions, although have been applied effectively to  
494 understand the aftermath and impacts of the Mt Samalas 1257 eruption (Malawani et al., 2022). In this study ensemble  
495 simulations do show hydroclimate perturbations following the Samalas eruption for both JUL1257 and JAN1258 scenarios  
496 (Figure S10), although due to the limited number of realisations these are not included in our model-proxy framework.  
497 Nonetheless, the magnitude of precipitation anomalies for Summer 1258 being greatest in the equatorial regions, potentially  
498 indicative of a shift in the ITCZ due to asymmetric cooling between hemispheres. Therefore, future studies may also consider  
499 including hydroclimate proxies alongside SAT reconstructions in order to add an additional constraint on model-proxy  
500 agreement.

501

502 Model-simulated anomalies are strongly dependent on model set up, including model resolution, modelled stratospheric winds,  
503 aerosol microphysics and sedimentation and deposition schemes (Marshall et al., 2018, Quaglia et al., 2023). In recent model  
504 intercomparison studies (Marshall et al., 2018, Quaglia et al., 2023) UM-UKCA, a previous version of the UKESM, showed  
505 a bias towards stronger transport to the NH extratropics, resulting in a hemispherically asymmetric aerosol load. The spatial  
506 distribution of volcanic forcing can influence subsequent growth of sulfate aerosols and their global distribution, in turn  
507 affecting the persistence of aerosols in the stratosphere (Quaglia et al., 2023). Compared to other global aerosol models UM-  
508 UKCA also has relatively weaker poleward transport, with stronger meridional deposition which may lead to a more  
509 equatorially focussed aerosol distribution and deposition (Marshall et al., 2018). Disentangling large inter-model differences  
510 from the range of model components that contribute to this uncertainty remains challenging, although future multi-model  
511 multi-proxy studies may be of use.

512

513 Even within a single model, uncertainties persist in initial eruption conditions (e.g phase of the QBO,) and volcanic forcing  
514 parameters (e.g timing, magnitude, injection height and latitude of eruption). Timmerick et al., (2021) demonstrated that both  
515 the magnitude of forcing as well as its spatial structure can similarly affect proxy–simulation comparisons, particularly in the  
516 NH extratropics. This is supported by Lücke et al., (2023) who demonstrate a significant spread in the temperature response  
517 due to volcanic forcing uncertainties which can strongly affect the agreement with proxy reconstructions. The VSSI estimate  
518 used in our study was taken from the eVolV2k reconstruction (Toohey and Sigl, 2017) and is within error of other SO<sub>2</sub> emission  
519 estimates for the Mt Samalas eruption (Vidal et al., 2016, Lavigne et al., 2013). Whilst maximum plume heights have been  
520 estimated for the Samalas eruption (~ 43 km Vidal et al., 2015) the SO<sub>2</sub> injection height remains unknown. In our simulations  
521 the injection height is set to 18-20 km, which may be too low, but does allow for lofting of sulfate aerosol. Moreover, Stoffel  
522 et al., (2015) found that increasing plume height from 22-26 km to 33-36 km for simulations of the Mt Samalas 1257 eruption  
523 increased the magnitude of the peak post-eruption NH JJA temperature anomaly to -4°C for a January eruption and -1°C and  
524 -2°C for May and July eruption scenarios respectively. A plume height greater than the 20km used in our study would therefore  
525 likely further enforce our central conclusion that better agreement between proxy and model-simulated temperature anomalies  
526 is achieved for a summer 1257 eruption date, whilst a greater plume height for a January 1258 eruption would only further

527 overpredict the magnitude of post-eruption cooling. Timmreck et al., (2009) also highlight the strong dependence of model-  
528 simulated post-eruption climate responses following large volcanic eruptions on the aerosol particle size distribution due to  
529 self-limiting effects of larger particles (Pinto et al., 1989). These particle characteristics are difficult to constrain retrospectively  
530 for historical eruptions such as Mt. Samalas and thus represent a significant uncertainty when simulating historical eruptions.

531  
532 Model-simulated ice sheet sulfate deposition (Fig. 4) shows a clear distinction between JAN1258 and JUL1257 eruption  
533 scenarios in both the timing and hemispheric distribution of deposition. For the same stratospheric SO<sub>2</sub> injection, an eruption  
534 during NH summer results in more pronounced asymmetric polar deposition, with the magnitude of peak deposition in  
535 Antarctica being nearly a factor of 3x greater than for an eruption during the SH summer, likely due to the stronger branch of  
536 the Brewer-Dobson circulation and seasonal effects on aerosol transport and depositional processes. Nonetheless, model  
537 simulations of the Tambora 1815 eruption by Marshall et al., 2018 found a strong model dependency for the timing, magnitude,  
538 and spatial distribution of sulfate deposition. Polar sulfate deposition in the UKESM was half that reconstructed using ice cores  
539 for the Tambora 1815 eruption and considerably lower than deposition in the other MAECHAM5-HAM and SOCOL-AER  
540 models analysed (See Table S3). In our JUL1257 and JAN1258 simulations for the Samalas 1257 eruption mean total  
541 deposition in Antarctica (55-28 Kg km<sup>-2</sup>) and Greenland (26-23 Kg km<sup>-2</sup>) was also lower than mean total deposition from ice  
542 cores records (73 and 105 Kg km<sup>-2</sup> respectively, see Table 2). The overall lower polar total sulfate deposition in the UKESM  
543 may be caused by too weak polarward transport (or stronger meridional deposition) (Marshall et al., 2018). For our JUL1257  
544 and JAN1258 model-simulated sulfate deposition for both eruption scenarios result in Greenland to Antarctica ratios < 1, with  
545 a Summer 1257 eruption showing a much more asymmetric distribution (0.49) than a January 1258 eruption (0.83). These  
546 ratios however, disagree with the Greenland to Antarctica ratio derived from ice cores (1.4, Toohey and Sigl, 2017) suggesting  
547 that, for our Samalas scenarios, sulfate aerosol transport to the NH is too low, thus favouring transport to the SH. Of the four  
548 models analysed in Marshall et al., (2018) for the Tambora 1815 eruption, UM-UKCA was one of two models that had greater  
549 deposition in Greenland compared to Antarctica (Table S3, Greenland/Antarctica = 1.7). Both the Tambora 1815 and Mt  
550 Samalas 1257 were large magnitude eruptions at a similar latitude, therefore this intra-model difference in the asymmetric  
551 distribution of sulfate aerosol most likely results from differences in initial conditions used for our simulations (such as the  
552 phase of the QBO). Overall, this further highlights the complications of disentangling inter-model differences and intra-model  
553 variation due to initial conditions. Given both these inter and intra-model differences in relative hemispheric aerosol  
554 distribution and deposition, the current robustness of using hemispheric sulfate deposition ratios to distinguish between  
555 eruption scenarios when compared to ice core records is limited, with further work being needed to understand model and  
556 starting condition specific effects.

557 Whilst comparisons of simulated and reconstructed climate responses following large volcanic eruptions have been used  
558 routinely and effectively by a multitude of studies (van Dijk et al., 2023, Büntgen et al., 2022, Stoffel et al., 2015 ), there are  
559 several limitations to this approach. Given the uncertainties associated with both proxy reconstructions and model simulations  
560 neither can be taken as the inherently “correct” baseline with which to fit the other, and thus particular care should be taken

561 when using model-proxy comparison to validate the correctness of underlying records or model input parameters. For example,  
562 the missing tree ring hypothesis (Mann et al., 2012), which has since been widely rejected, proposed that the mismatch between  
563 climate simulations and proxy reconstructions resulted from chronological errors due to missing growth rings. This has,  
564 however, since been resolved with improved estimates of volcanic forcing (Toohey & Sigl 2017), the inclusion of climate-  
565 independent geochronological data, and the greater inclusion of MXD records in tree ring reconstructions (Stoffel et al., 2015;  
566 Schneider et al. 2015; Wilson et al., 2016; Anchukaitis et al. 2017). When utilising an array of different proxy data there is the  
567 risk of confirmation bias meaning records which show significant agreement with model simulations being given greater  
568 weighting than those which show less agreement. A more quantifiable approach to model-proxy comparison, where a greater  
569 number of model realisations would allow for more robust statistical evaluation would therefore be a considerable  
570 improvement for the application of the model-multi proxy framework. Further to this, better uncertainty quantification for  
571 proxy data would enable more robust comparison with model outputs, particularly in multi-model comparison studies.

## 572 **5 Conclusions**

573 We have utilized eighteen aerosol-climate UKESM1 ensemble simulations for the Mt. Samalas eruption in combination with  
574 an extensive globally-resolved multi-proxy database for the Mt. Samalas eruption in order to constrain the year and season as  
575 well as the regionally heterogeneous climate response following the eruption. Comparison with NH averaged and spatially  
576 resolved tree ring reconstructions showed that a Summer 1257 eruption scenario agrees best with reconstructed SAT  
577 anomalies, while a January 1258 eruption consistently overpredicts the magnitude of cooling relative to reconstructions. The  
578 regionally variable SAT response following the eruption is revealed by multi-proxy reconstructions which lend support to  
579 inferred social, economic and historical consequences across Europe and Asia following the eruption. Model-simulated SAT  
580 anomalies also suggest the onset of sudden a severe cooling across the SH, with the potential for significant social and  
581 economic consequences in impacted communities across South America, Africa, and Oceania. The spatial distribution of SAT  
582 anomalies shows sensitivity to initial atmospheric-ocean conditions, with positive SAT anomalies in Alaska being potentially  
583 indicative of warm El Niño-like conditions at the time of the eruption, with potential ENSO-Volcano interactions enhancing  
584 these conditions further.

585  
586 Overall, the proxy to model comparison employed in this study has been shown as an effective approach to constrain uncertain  
587 eruption source parameters. Model-multi proxy frameworks have similarly been employed by other recent studies (see Guillet  
588 et al., 2023 and van Dijk et al., 2023) which have also demonstrated its potential in constraining unknown eruption source  
589 parameters as well as regional climatic impacts for historic eruptions where there is sufficient concurrent proxy evidence. A  
590 greater global distribution of proxy evidence, especially in the Southern Hemisphere where all types of proxy evidence are  
591 sparse, will strengthen this proxy-model framework approach for future analysis. The incorporation of hydroclimate anomalies  
592 in particular has the potential to add further independent constraints, although relies on the development of higher resolution  
593 records, especially at low latitude sites. Model simulations of polar sulfate deposition also reveal distinct differences in the

594 timing of ice sheet deposition between the two simulated eruption seasons, although comparison of the magnitude or  
595 asymmetric deposition of sulfate aerosol remains limited by large inter-model differences and complex intra-model  
596 dependencies.

597

598 Finally, both proxy-reconstructed and model-simulated surface temperature anomalies highlight the severity of the global  
599 climate response following a large tropical explosive eruption like Mt. Samalas, with historical records confirming widespread  
600 and severe economic and social consequences. This adds further weight to recent calls (Cassidy and Mani, 2022) for increased  
601 global preparedness for the next large magnitude explosive volcanic eruption, given profound global consequences that would  
602 be expected, as clearly demonstrated by the Mt. Samalas eruption.

603

#### 604 **Data Availability**

605 Data has been uploaded to the CEDA archive and is pending review. Catalogue record can be found at  
606 <https://catalogue.ceda.ac.uk/uuid/e0221b37aa174dd290c5e105263b59d1>.

607

#### 608 **Supplement**

609 The supplement related to this article is available online at:

610

#### 611 **Author Contributions**

612

613 LW, LM, and AS jointly conceived the project methodology. LM ran the UKESM model simulations. LW performed the  
614 analysis, visualisation, and writing of the manuscript with supervision from LM and AS. All authors jointly reviewed and  
615 edited the paper.

616

#### 617 **Competing Interests**

618 The authors declare that they have no conflict of interest.

619

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### 632 **References**

633 Anchukaitis, K.J., Wilson, R., Briffa, K.R., Büntgen, U., Cook, E.R., D'Arrigo, R., Davi, N., Esper, J., Frank, D., Gunnarson,  
634 B.E., Hegerl, G., Helama, S., Klesse, S., Krusic, P.J., Linderholm, H.W., Myglan, V., Osborn, T.J., Zhang, P., Rydval, M. and  
635 Schneider, L. (2017). Last millennium Northern Hemisphere summer temperatures from tree rings: Part II, spatially resolved  
636 reconstructions. *Quaternary Science Reviews*, [online] 163, pp.1–22. doi:10.1016/j.quascirev.2017.02.020.

637

638 Anderson, A. *The first migration : Māori origins 3000BC - AD1450*. Wellington, New Zealand: Bridget Williams Books. Doi:  
639 10.7810/9780947492793, 2016.

640

641 Archibald, A.T., O'Connor, F.M., Abraham, N.L., Archer-Nicholls, S., Chipperfield, M.P., Dalvi, M., Folberth, G.A.,  
642 Dennison, F., Dhomse, S.S., Griffiths, P.T., Hardacre, C., Hewitt, A.J., Hill, R.S., Johnson, C.E., Keeble, J., Köhler, M.O.,  
643 Morgenstern, O., Mulcahy, J.P., Ordóñez, C. and Pope, R.J. (2020). Description and evaluation of the UKCA stratosphere–  
644 troposphere chemistry scheme (StratTrop vn 1.0) implemented in UKESM1. *Geoscientific Model Development*, 13(3),  
645 pp.1223–1266. doi:10.5194/gmd-13-1223-2020.

646

647 Bauch, M. (2019). *Chronology and impact of a global moment in the 13th century*. In: *The Dance of Death in Late Medieval  
648 and Renaissance Europe*. Routledge. ISBN 9781032083391

649

650 Bierstedt, A. (2019). *Weather and Ideology in Íslendinga saga A Case Study of the Volcanic Climate Forcing of the 1257  
651 Samalas eruption*. M.A. thesis, University of Iceland

652

653 Borisenkov, Y.P. and Pasetky, V.M., 1988. *Millennium-Long Chronicle of Unusual Natural Events*. Misl', Moscow.

654

655 Broccoli, A.J., Dahl, K.A. and Stouffer, R.J. (2006). Response of the ITCZ to Northern Hemisphere cooling. *Geophysical  
656 Research Letters*, 33(1), p.n/a-n/a. doi:10.1029/2005gl024546.

657

658 Bunbury, M.M.E., Petchey, F. and Bickler, S.H. (2022). A new chronology for the Māori settlement of Aotearoa (NZ) and the  
659 potential role of climate change in demographic developments. *Proceedings of the National Academy of Sciences*, [online]  
660 119(46). doi:10.1073/pnas.2207609119.

661

662 Büntgen, U., Allen, K., Anchukaitis, K.J., Arseneault, D., Boucher, É., Bräuning, A., Chatterjee, S., Cherubini, P., Churakova  
663 (Sidorova), O.V., Corona, C., Gennaretti, F., Gießinger, J., Guillet, S., Guiot, J., Gunnarson, B., Helama, S., Hochreuther, P.,  
664 Hughes, M.K., Huybers, P. and Kirilyanov, A.V. (2021). The influence of decision-making in tree ring-based climate  
665 reconstructions. *Nature Communications*, [online] 12(1), p.3411. doi:10.1038/s41467-021-23627-6.

666

667 Büntgen, U., Kyncl, T., Ginzler, C., Jacks, D.S., Esper, J., Tegel, W., Heussner, K.-U. and Kyncl, J. (2013). Filling the Eastern  
668 European gap in millennium-long temperature reconstructions. *Proceedings of the National Academy of Sciences*, 110(5),  
669 pp.1773–1778. doi:10.1073/pnas.1211485110.

670



671 Büntgen, U., Smith, S.H., Wagner, S., Krusic, P., Esper, J., Piermattei, A., Crivellaro, A., Reinig, F., Tegel, W., Kirilyanov,  
672 A., Trnka, M. and Oppenheimer, C. (2022). Global tree-ring response and inferred climate variation following the mid-  
673 thirteenth century Samalas eruption. *Climate Dynamics*, 59(1-2), pp.531–546. doi:10.1007/s00382-022-06141-3.

674  
675 Buntgen, U., Tegel, W., Nicolussi, K., McCormick, M., Frank, D., Trouet, V., Kaplan, J.O., Herzig, F., Heussner, K.-U. ,  
676 Wanner, H., Luterbacher, J. and Esper, J. (2011). 2500 Years of European Climate Variability and Human Susceptibility.  
677 *Science*, 331(6017), pp.578–582. doi:10.1126/science.1197175.

678  
679 Büntgen, U., Arseneault, D., Boucher, É., Churakova, O. V., Gennaretti, F., Crivellaro, A., Hughes, M. K., Kirilyanov, A. V.,  
680 Klippel, L., Krusic, P. J., Linderholm, H. W., Ljungqvist, F. C., Ludescher, J., McCormick, M., Myglan, V. S., Nicolussi, K.,  
681 Piermattei, A., Oppenheimer, C., Reinig, F., Sigl, M., Vaganov, E. A., and Esper, J.: Prominent role of volcanism in Common  
682 Era climate variability and human history, *Dendrochronologia*, 64, 125757, 2020.

683  
684 Büntgen, U., Myglan, V. S., Ljungqvist, F. C., McCormick, M., Di Cosmo, N., Sigl, M., Jungclauss, J., Wagner, S., Krusic, P.  
685 J., Esper, J., Kaplan, J. O., de Vaan, M. A. C., Luterbacher, J., Wacker, L., Tegel, W., and Kirilyanov, A. V.: Cooling and  
686 societal change during the Late Antique Little Ice Age from 536 to around 660 AD, *Nat Geosci*, 9, 231-236, 2016.

687  
688 Campbell, B.M.S. (2017). Global climates, the 1257 mega-eruption of Samalas volcano, Indonesia, and the English food crisis  
689 of 1258. *Transactions of the Royal Historical Society*, 27, pp.87–121. doi:10.1017/s0080440117000056.

690  
691 Cassidy, M. and Mani, L. (2022). Huge volcanic eruptions: time to prepare. *Nature*, [online] 608(7923), pp.469–471.  
692 doi:10.1038/d41586-022-02177-x.

693  
694 Clyne, M., Lamarque, J. F., Mills, M. J., Khodri, M., Ball, W., Bekki, S., Dhomse, S. S., Lebas, N., Mann, G., Marshall, L.,  
695 Niemeier, U., Poulain, V., Robock, A., Rozanov, E., Schmidt, A., Stenke, A., Sukhodolov, T., Timmreck, C., Toohey, M.,  
696 Tummon, F., Zanchettin, D., Zhu, Y., and Toon, O. B.: Model physics and chemistry causing intermodel disagreement within  
697 the VolMIP-Tambora Interactive Stratospheric Aerosol ensemble, *Atmos. Chem. Phys.*, 21, 3317-3343, 2021.

698  
699 Davi, N.K., D’Arrigo, R., Jacoby, G.C., Cook, E.R., Anchukaitis, K.J., Nachin, B., Rao, M.P. and Leland, C. (2015). A long-  
700 term context (931–2005 C.E.) for rapid warming over Central Asia. *Quaternary Science Reviews*, 121, pp.89–97.  
701 doi:10.1016/j.quascirev.2015.05.020.

702  
703 Davi, N.K., Rao, M.P., Wilson, R., Andreu-Hayles, L., Oelkers, R., D’Arrigo, R., Nachin, B., Buckley, B., Pederson, N.,  
704 Leland, C. and Suran, B. (2021). Accelerated Recent Warming and Temperature Variability Over the Past Eight Centuries in  
705 the Central Asian Altai From Blue Intensity in Tree Rings. *Geophysical Research Letters*, [online] 48(16).  
706 doi:10.1029/2021gl092933.

707  
708 Dee, S.G., Cobb, K.M., Emile-Geay, J., Ault, T.R., Edwards, R.L., Cheng, H. and Charles, C.D. (2020). No consistent ENSO  
709 response to volcanic forcing over the last millennium. *Science*, [online] 367(6485), pp.1477–1481.  
710 doi:10.1126/science.aax2000.

711  
712 Dhomse, S. S., Emmerson, K. M., Mann, G. W., Bellouin, N., Carslaw, K. S., Chipperfield, M. P., Hommel, R., Abraham, N.  
713 L., Telford, P., Braesicke, P., Dalvi, M., Johnson, C. E., O’Connor, F., Morgenstern, O., Pyle, J. A., Deshler, T., Zawodny, J.  
714 M., and Thomason, L. W.: Aerosol microphysics simulations of the Mt. Pinatubo eruption with the UM-UKCA composition-  
715 climate model, *Atmos. Chem. Phys.*, 14, 11221–11246, <https://doi.org/10.5194/acp-14-11221-2014>, 2014.

716  
717 Divine, D., Isaksson, E., Martma, T., Meijer, H.A.J., Moore, J., Pohjola, V., van de Wal, R.S.W. and Godtlielsen, F. (2011).  
718 Thousand years of winter surface air temperature variations in Svalbard and northern Norway reconstructed from ice-core  
719 data. *Polar Research*, 30(1), p.7379. doi:10.3402/polar.v30i0.7379.

720

721 Esper, J., DÜthorn, E., Krusic, P.J., Timonen, M. and Büntgen, U. (2014). Northern European summer temperature variations  
722 over the Common Era from integrated tree-ring density records. *Journal of Quaternary Science*, 29(5), pp.487–494.  
723 doi:10.1002/jqs.2726.

724

725 Esper, J., George, S.St., Anchukaitis, K., D’Arrigo, R., Ljungqvist, F.C., Luterbacher, J., Schneider, L., Stoffel, M., Wilson,  
726 R. and Büntgen, U. (2018). Large-scale, millennial-length temperature reconstructions from tree-rings. *Dendrochronologia*,  
727 [online] 50, pp.81–90. doi:10.1016/j.dendro.2018.06.001.

728

729 Fell, H.G., Baldini, J.U.L., Dodds, B. and Sharples, G.J. (2020). Volcanism and global plague pandemics: Towards an  
730 interdisciplinary synthesis. *Journal of Historical Geography*, [online] 70, pp.36–46. doi:10.1016/j.jhg.2020.10.001.

731

732 Gennaretti, F., Arseneault, D., Nicault, A., Perreault, L. and Begin, Y. (2014). Volcano-induced regime shifts in millennial  
733 tree-ring chronologies from northeastern North America. *Proceedings of the National Academy of Sciences*, 111(28),  
734 pp.10077–10082. doi:10.1073/pnas.1324220111.

735

736 Grissino-Mayer, Henri D. 1996. A 2129 year annual reconstruction of precipitation for northwestern New Mexico, USA. In  
737 Dean, J.S., Meko, D.M., and Swetnam, T.W., eds., *Tree Rings, Environment, and Humanity*. Radiocarbon 1996, Department  
738 of Geosciences, The University of Arizona, Tucson: 191-204.

739

740 Guillet, S., Corona, C., Stoffel, M., Khodri, M., Lavigne, F., Ortega, P., Eckert, N., Sielenou, P.D., Daux, V., Churakova  
741 (Sidorova), Olga V., Davi, N., Edouard, J.-L., Zhang, Y., Luckman, Brian H., Myglan, V.S., Guiot, J., Beniston, M., Masson-  
742 Delmotte, V. and Oppenheimer, C. (2017). Climate response to the Samalas volcanic eruption in 1257 revealed by proxy  
743 records. *Nature Geoscience*, 10(2), pp.123–128. doi:10.1038/ngeo2875.

744

745 Guillet, S., Corona, C., Oppenheimer, C., Lavigne, F., Khodri, M., Ludlow, F., Sigl, M., Toohey, M., Atkins, P. S., Yang, Z.,  
746 Muranaka, T., Horikawa, N., and Stoffel, M.: Lunar eclipses illuminate timing and climate impact of medieval volcanism,  
747 *Nature*, 616, 90-95, 2023.

748

749 Helama, S., Huhtamaa, H., Verkasalo, E. and Läänelaid, A. (2017). Something old, something new, something borrowed: New  
750 insights to human-environment interaction in medieval Novgorod inferred from tree rings. *Journal of Archaeological Science:*  
751 *Reports*, 13, pp.341–350. doi:10.1016/j.jasrep.2017.04.008.

752

753 Kern, Z., Pow, S., Pinke, Z. and Ferenczi, L. (2021). Samalas and the Fall of the Mongol Empire: A volcanic eruption’s  
754 influence on the dissolution of history’s largest contiguous empire. *EGU General Assembly*. doi:10.5194/egusphere-egu21-  
755 3460.

756

757 Lavigne, F., Degeai, J.-P. ., Komorowski, J.-C. ., Guillet, S., Robert, V., Lahitte, P., Oppenheimer, C., Stoffel, M., Vidal, C.M.,  
758 Surono, Pratomo, I., Wassmer, P., Hajdas, I., Hadmoko, D.S. and de Belizal, E. (2013). Source of the great A.D. 1257 mystery  
759 eruption unveiled, Samalas volcano, Rinjani Volcanic Complex, Indonesia. *Proceedings of the National Academy of Sciences*,  
760 110(42), pp.16742–16747. doi:10.1073/pnas.1307520110.

761

762 Lücke, L. J., Schurer, A. P., Toohey, M., Marshall, L. R., and Hegerl, G. C.: The effect of uncertainties in natural forcing  
763 records on simulated temperature during the last millennium, *Clim. Past*, 19, 959–978, [https://doi.org/10.5194/cp-19-959-](https://doi.org/10.5194/cp-19-959-2023)  
764 [2023](https://doi.org/10.5194/cp-19-959-2023), 2023.

765

766 Lücke, L.J., Hegerl, G.C., Schurer, A.P. and Wilson, R. (2019). Effects of Memory Biases on Variability of Temperature  
767 Reconstructions. *Journal of Climate*, [online] 32(24), pp.8713–8731. doi:10.1175/jcli-d-19-0184.1.

768

769 Luckman, B.H. and Wilson, R.J.S. (2005). Summer temperatures in the Canadian Rockies during the last millennium: a revised  
770 record. *Climate Dynamics*, 24(2-3), pp.131–144. doi:10.1007/s00382-004-0511-0.

771  
772 Malawani, M.N., Lavigne, F., Sastrawan, W.J., Jamaluddin, Sirulhaq, A. and Hadmoko, D.S. (2022). The 1257 CE cataclysmic  
773 eruption of Samalas volcano (Indonesia) revealed by indigenous written sources: Forgotten kingdoms, emergency response,  
774 and societal recovery. *Journal of Volcanology and Geothermal Research*, [online] 432, p.107688.  
775 doi:10.1016/j.jvolgeores.2022.107688.  
776  
777 Mann, M.E., Cane, M.A., Zebiak, S.E. and Clement, A. (2005). Volcanic and Solar Forcing of the Tropical Pacific over the  
778 Past 1000 Years. *Journal of Climate*, [online] 18(3), pp.447–456. doi:10.1175/jcli-3276.1.  
779  
780 Mann, M.E., Fuentes, J.D. and Rutherford, S. (2012). Underestimation of volcanic cooling in tree-ring-based reconstructions  
781 of hemispheric temperatures. *Nature Geoscience*, 5(3), pp.202–205. doi:10.1038/ngeo1394.  
782  
783 Marshall, L.R., Schmidt, A., Johnson, J.S., Mann, G.W., Lee, L.A., Rigby, R. and Carslaw, K.S. (2021). Unknown Eruption  
784 Source Parameters Cause Large Uncertainty in Historical Volcanic Radiative Forcing Reconstructions. *Journal of Geophysical*  
785 *Research: Atmospheres*, [online] 126(13). doi:10.1029/2020jd033578.  
786  
787 Martin, J.T., Pederson, G.T., Woodhouse, C.A., Cook, E.R., McCabe, G.J., Anchukaitis, K.J., Wise, E.K., Erger, P.J., Dolan,  
788 L., McGuire, M., Gangopadhyay, S., Chase, K.J., Littell, J.S., Gray, S.T., George, S.S., Friedman, J.M., Sauchyn, D.J., St-  
789 Jacques, J.-M. and King, J. (2020). Increased drought severity tracks warming in the United States’ largest river basin.  
790 *Proceedings of the National Academy of Sciences*, [online] 117(21), pp.11328–11336. doi:10.1073/pnas.1916208117.  
791  
792 McCarroll, D., Loader, N.J., Jalkanen, R., Gagen, M.H., Grudd, H., Gunnarson, B.E., Kirchhefer, A.J., Friedrich, M.,  
793 Linderholm, H.W., Lindholm, M., Boettger, T., Los, S.O., Remmele, S., Kononov, Y.M., Yamazaki, Y.H., Young, G.H. and  
794 Zorita, E. (2013). A 1200-year multiproxy record of tree growth and summer temperature at the northern pine forest limit of  
795 Europe. *The Holocene*, 23(4), pp.471–484. doi:10.1177/0959683612467483.  
796  
797 McGregor, S., Khodri, M., Maher, N., Ohba, M., Pausata, F.S.R. and Stevenson, S. (2020). The Effect of Strong Volcanic  
798 Eruptions on ENSO. *Geophysical Monograph Series*, [online] pp.267–287. doi:10.1002/9781119548164.ch12.  
799  
800 Melvin, T.M., Grudd, H. and Briffa, K.R. (2012). Potential bias in ‘updating’ tree-ring chronologies using regional curve  
801 standardisation: Re-processing 1500 years of Torneträsk density and ring-width data. *The Holocene*, 23(3), pp.364–373.  
802 doi:10.1177/0959683612460791.  
803  
804 Moore, J.J., Hughen, K.A., Miller, G.H. and Overpeck, J.T. (2001). *Journal of Paleolimnology*, [online] 25(4), pp.503–517.  
805 doi:10.1023/a:1011181301514.  
806  
807 Mulcahy, J.P., Johnson, C., Jones, C.G., Povey, A.C., Scott, C.E., Sellar, A., Turnock, S.T., Woodhouse, M.T., Abraham, N.L.,  
808 Andrews, M.B., Bellouin, N., Browse, J., Carslaw, K.S., Dalvi, M., Folberth, G.A., Glover, M., Grosvenor, D.P., Hardacre,  
809 C., Hill, R. and Johnson, B. (2020). Description and evaluation of aerosol in UKESM1 and HadGEM3-GC3.1 CMIP6 historical  
810 simulations. *Geoscientific Model Development*, 13(12), pp.6383–6423. doi:10.5194/gmd-13-6383-2020.  
811  
812 Oppenheimer, C. (2003). Ice core and palaeoclimatic evidence for the timing and nature of the great mid-13th century volcanic  
813 eruption. *International Journal of Climatology*, 23(4), pp.417–426. doi:10.1002/joc.891.  
814  
815 PAGES2k Consortium: Data Descriptor: A global multiproxy database for temperature reconstructions of the Common Era,  
816 *Sci Data*, 4, 2017.  
817  
818 Palais, J.M., Germani, M.S. and Zielinski, G.A. (1992). Inter-hemispheric Transport of Volcanic Ash from a 1259 A.D.  
819 Volcanic Eruption to the Greenland and Antarctic Ice Sheets. *Geophysical Research Letters*, 19(8), pp.801–804.  
820 doi:10.1029/92gl00240.

821  
822 Pinto, J.P., Turco, R.P. and Toon, O.B. (1989). Self-limiting physical and chemical effects in volcanic eruption clouds. *Journal*  
823 *of Geophysical Research*, [online] 94(D8), p.11165. doi:10.1029/jd094id08p11165.  
824  
825 Porter, T.J., Pisaric, M.F.J., Kokelj, S.V. and deMontigny, P. (2013). A ring-width-based reconstruction of June–July minimum  
826 temperatures since AD 1245 from white spruce stands in the Mackenzie Delta region, northwestern Canada. *Quaternary*  
827 *Research*, 80(2), pp.167–179. doi:10.1016/j.yqres.2013.05.004.  
828  
829 Quaglia, I., Timmreck, C., Niemeier, U., Visionsi, D., Pitari, G., Brodowsky, C., Brühl, C., Dhomse, S. S., Franke, H., Laakso,  
830 A., Mann, G. W., Rozanov, E., and Sukhodolov, T.: Interactive stratospheric aerosol models' response to different amounts  
831 and altitudes of SO<sub>2</sub> injection during the 1991 Pinatubo eruption, *Atmos. Chem. Phys.*, 23, 921–948, 2023.  
832 Robock, A. (2000). Volcanic eruptions and climate. *Reviews of Geophysics*, [online] 38(2), pp.191–219.  
833 doi:10.1029/1998rg000054.  
834  
835 Robock, A. (2005). Cooling following large volcanic eruptions corrected for the effect of diffuse radiation on tree rings.  
836 *Geophysical Research Letters*, 32(6). doi:10.1029/2004gl022116.  
837  
838 Robock, A. (2020). Comment on ‘No consistent ENSO response to volcanic forcing over the last millennium’. *Science*,  
839 [online] 369(6509). doi:10.1126/science.abc0502.  
840  
841 Rohde, R.A. and Hausfather, Z. (2020). The Berkeley Earth Land/Ocean Temperature Record. *Earth System Science Data*,  
842 12(4), pp.3469–3479. doi:https://doi.org/10.5194/essd-12-3469-2020.  
843  
844 Rydval, M., Loader, N.J., Gunnarson, B.E., Druckenbrod, D.L., Linderholm, H.W., Moreton, S.G., Wood, C.V. and Wilson,  
845 R. (2017). Reconstructing 800 years of summer temperatures in Scotland from tree rings. *Climate Dynamics*, 49(9–10),  
846 pp.2951–2974. doi:10.1007/s00382-016-3478-8.  
847  
848 Saliba, G. (2017). Cultural Implications of Natural Disasters: Historical Reports of the Volcano Eruption of July, 1256 CE.  
849 *Transcultural Research – Heidelberg Studies on Asia and Europe in a Global Context*, [online] pp.139–154. doi:10.1007/978-  
850 3-319-49163-9\_7.  
851  
852 Salzer, M. W. and Hughes, M. K.: Bristlecone pine tree rings and volcanic eruptions over the last 5000 yr, *Quaternary Res*,  
853 67, 57–68, 2007.  
854  
855 Schneider, L., Smerdon, J.E., Büntgen, U., Wilson, R.J.S., Myglan, V.S., Kirilyanov, A.V. and Esper, J. (2015). Revising  
856 midlatitude summer temperatures back to A.D. 600 based on a wood density network. *Geophysical Research Letters*, 42(11),  
857 pp.4556–4562. doi:10.1002/2015gl063956.  
858  
859 Sellar, A.A., Jones, C.G., Mulcahy, J.P., Tang, Y., Yool, A., Wiltshire, A., O’Connor, F.M., Stringer, M., Hill, R., Palmieri,  
860 J., Woodward, S., Mora, L., Kuhlbrodt, T., Rumbold, S.T., Kelley, D.I., Ellis, R., Johnson, C.E., Walton, J., Abraham, N.L.  
861 and Andrews, M.B. (2019). UKESM1: Description and Evaluation of the U.K. Earth System Model. *Journal of Advances in*  
862 *Modeling Earth Systems*, 11(12), pp.4513–4558. doi:10.1029/2019ms001739.  
863  
864 Sigl, M., Winstrup, M., McConnell, J.R., Welten, K.C., Plunkett, G., Ludlow, F., Büntgen, U., Caffee, M., Chellman, N., Dahl-  
865 Jensen, D., Fischer, H., Kipfstuhl, S., Kostick, C., Maselli, O.J., Mekhaldi, F., Mulvaney, R., Muscheler, R., Pasteris, D.R.,  
866 Pilcher, J.R. and Salzer, M. (2015). Timing and climate forcing of volcanic eruptions for the past 2,500 years. *Nature*, [online]  
867 523(7562), pp.543–549. doi:10.1038/nature14565.  
868

869 Stevenson, S., Fasullo, J.T., Otto-Bliesner, B.L., Tomas, R.A. and Gao, C. (2017). Role of eruption season in reconciling  
870 model and proxy responses to tropical volcanism. *Proceedings of the National Academy of Sciences*, 114(8), pp.1822–1826.  
871 doi:10.1073/pnas.1612505114.

872

873 Stevenson, S., Overpeck, J.T., Fasullo, J., Coats, S., Parsons, L., Otto-Bliesner, B., Ault, T., Loope, G. and Cole, J. (2018).  
874 Climate Variability, Volcanic Forcing, and Last Millennium Hydroclimate Extremes. *Journal of Climate*, [online] 31(11),  
875 pp.4309–4327. doi:10.1175/jcli-d-17-0407.1.

876

877 Stoffel, M., Khodri, M., Corona, C., Guillet, S., Poulain, V., Bekki, S., Guiot, J., Luckman, B.H., Oppenheimer, C., Lebas, N.,  
878 Beniston, M. and Masson-Delmotte, V. (2015). Estimates of volcanic-induced cooling in the Northern Hemisphere over the  
879 past 1,500 years. *Nature Geoscience*, 8(10), pp.784–788. doi:10.1038/ngeo2526.

880

881 Stothers, R.B. (2000). Climatic and Demographic Consequences of the Massive Volcanic Eruption of 1258. *Climatic Change*,  
882 45(2), pp.361–374. doi:10.1023/a:1005523330643.

883

884 Stothers, Richard B. (2005). Stratospheric Transparency Derived from Total Lunar Eclipse Colors, 1801–1881. *Publications*  
885 *of the Astronomical Society of the Pacific*, [online] 117(838), pp.1445–1450. doi:10.1086/497016.

886

887 Tejedor, E., Steiger, N.J., Smerdon, J.E., Serrano-Notivoli, R. and Vuille, M. (2021). Global hydroclimatic response to tropical  
888 volcanic eruptions over the last millennium. *Proceedings of the National Academy of Sciences*, 118(12), p.e2019145118.  
889 doi:10.1073/pnas.2019145118.

890

891 Timmreck, C., Lorenz, S.J., Crowley, T.J., Kinne, S., Raddatz, T.J., Thomas, M.A. and Jungclaus, J.H. (2009). Limited  
892 temperature response to the very large AD 1258 volcanic eruption. *Geophysical Research Letters*, 36(21).  
893 doi:10.1029/2009gl040083.

894

895 Timmreck, C., Toohey, M., Zanchettin, D., Brönnimann, S., Lundstad, E., & Wilson, R. (2021). The unidentified eruption of  
896 1809: a climatic cold case. *Climate of the Past*, 17(4), 1455-963 1482. <https://doi.org/10.5194/cp-17-1455-2021>.

897

898 Toohey, M., Krüger, K., Niemeier, U. and Timmreck, C. (2011). The influence of eruption season on the global aerosol  
899 evolution and radiative impact of tropical volcanic eruptions. *Atmospheric Chemistry and Physics*, 11(23), pp.12351–12367.  
900 doi:10.5194/acp-11-12351-2011.

901

902 Toohey, M. and Sigl, M. (2017). Volcanic stratospheric sulfur injections and aerosol optical depth from 500 BCE to 1900 CE.  
903 *Earth System Science Data*, 9(2), pp.809–831. doi:10.5194/essd-9-809-2017.

904

905 van Dijk, E., Mørkestøl Gundersen, I., de Bode, A., Høeg, H., Loftsgarden, K., Iversen, F., Timmreck, C., Jungclaus, J., and  
906 Krüger, K.: Climatic and societal impacts in Scandinavia following the 536 and 540 CE volcanic double event, *Clim. Past*, 19,  
907 357–398, <https://doi.org/10.5194/cp-19-357-2023>, 2023.

908

909 Vidal, C.M., Komorowski, J.-C., Métrich, N., Pratomo, I., Kartadinata, N., Prambada, O., Michel, A., Carazzo, G., Lavigne,  
910 F., Rodysill, J., Fontijn, K. and Surono (2015). Dynamics of the major plinian eruption of Samalas in 1257 A.D. (Lombok,  
911 Indonesia). *Bulletin of Volcanology*, 77(9). doi:10.1007/s00445-015-0960-9.

912

913 Vidal, C.M., Métrich, N., Komorowski, J.-C., Pratomo, I., Michel, A., Kartadinata, N., Robert, V. and Lavigne, F. (2016). The  
914 1257 Samalas eruption (Lombok, Indonesia): the single greatest stratospheric gas release of the Common Era. *Scientific*  
915 *Reports*, [online] 6(1). doi:10.1038/srep34868.

916

917 Vinther, B.M., Jones, P.D., Briffa, K.R., Clausen, H.B., Andersen, K.K., Dahl-Jensen, D. and Johnsen, S.J. (2010). Climatic  
918 signals in multiple highly resolved stable isotope records from Greenland. *Quaternary Science Reviews*, [online] 29(3-4),  
919 pp.522–538. doi:10.1016/j.quascirev.2009.11.002.  
920  
921 Wade, D.C., Vidal, C.M., Abraham, N.L., Dhomse, S., Griffiths, P.T., Keeble, J., Mann, G., Marshall, L., Schmidt, A. and  
922 Archibald, A.T. (2020). Reconciling the climate and ozone response to the 1257 CE Mount Samalas eruption. *Proceedings of*  
923 *the National Academy of Sciences*, 117(43), pp.26651–26659. doi:10.1073/pnas.1919807117.  
924  
925 Wiles, G.C., D’Arrigo, R.D., Barclay, D., Wilson, R.S., Jarvis, S.K., Vargo, L. and Frank, D. (2014). Surface air temperature  
926 variability reconstructed with tree rings for the Gulf of Alaska over the past 1200 years. *The Holocene*, 24(2), pp.198–208.  
927 doi:10.1177/0959683613516815.  
928  
929 William Wayne Farris (2006). *Japan’s medieval population : famine, fertility, and warfare in a transformative age*. Honolulu:  
930 University Of Hawai’i Press. ISBN-13: 9780824834241  
931  
932 Wilson, R., Anchukaitis, K., Briffa, K.R., Büntgen, U., Cook, E., D’Arrigo, R., Davi, N., Esper, J., Frank, D., Gunnarson, B.,  
933 Hegerl, G., Helama, S., Klesse, S., Krusic, P.J., Linderholm, H.W., Myglan, V., Osborn, T.J., Rydval, M., Schneider, L. and  
934 Schurer, A. (2016). Last millennium northern hemisphere summer temperatures from tree rings: Part I: The long term context.  
935 *Quaternary Science Reviews*, [online] 134, pp.1–18. doi:10.1016/j.quascirev.2015.12.005.  
936  
937 Zhang, P., Linderholm, H.W., Gunnarson, B.E., Björklund, J. and Chen, D. (2016). 1200 years of warm-season temperature  
938 variability in central Scandinavia inferred from tree-ring density. *Climate of the Past*, 12(6), pp.1297–1312. doi:10.5194/cp-  
939 12-1297-2016.  
940  
941 Zielinski, G.A., Mayewski, P.A., Meeker, L.D., Whitlow, S., Twickler, M.S., Morrison, M., Meese, D.A., Gow, A.J. and Alley,  
942 R.B. (1994). Record of Volcanism Since 7000 B.C. from the GISP2 Greenland Ice Core and Implications for the Volcano-  
943 Climate System. *Science*, 264(5161), pp.948–952. doi:10.1126/science.264.5161.948.  
944  
945