



- 1 **Revisiting the eruptive nature of the 1108–1110 CE volcanic episode:**
2 **Insights from sulfur isotopes and cryptotephra in Greenland ice cores**
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14 ABSTRACT

15 Major volcanic eruptions inject large quantities of sulfur-rich gases into the atmosphere,
16 which can drive seasonal to multiyear cooling on hemispheric to global scales. Such
17 volcanically forced cooling can cause profound societal disruption. A prominent example is the
18 1108–1110 CE volcanic episode, marked by one of the largest sulfate deposition events
19 recorded in polar ice cores over the past 2500 years and linked to some of the most severe
20 Northern Hemisphere (NH) summer cooling and widespread societal disruption of the past
21 1500 years. Our understanding of this episode has relied largely on ice-core sulfate
22 concentration records, and key questions remain about whether it represents a single eruption
23 or multiple eruptions and about the plume injection heights, source regions, and climatic
24 impacts of the volcanism involved.

25 Here, we reassess this volcanic episode by integrating new high-resolution sulfur isotope
26 records and cryptotephra geochemistry from Greenland ice cores. Our isotopic data provide
27 direct evidence that the 1108–1110 CE episode, which is currently considered a single tropical
28 event in volcanic forcing databases, in fact comprises a cluster of eruptions. These include an
29 NH extratropical eruption beginning in mid-1108 CE, a subsequent tropical eruption, and a
30 minor NH extratropical eruption in early 1111 CE, with the first two injecting sulfur into the
31 stratosphere. Cryptotephra analysis identifies basaltic glass shards geochemically consistent
32 with Icelandic sources; however, this evidence does not provide definitive provenance
33 constraints for the two stratospheric sulfur-injecting eruptions.

34 Based on these new constraints, we revise the volcanic stratospheric sulfur injection for
35 this episode to 9.0 ± 1.4 Tg S, approximately 53 % lower than the previous estimate of $19.2 \pm$
36 4.7 Tg S, indicating that the stratospheric sulfur load associated with this episode has been
37 overestimated. The timing of sulfate deposition further indicates that the initial NH
38 extratropical eruption was the primary driver of the severe 1109 CE summer cooling, largely
39 through sulfate transported via the troposphere and lowermost stratosphere. We therefore
40 suggest that such aerosols, which are typically excluded from total forcing estimates, may play
41 an underappreciated role in the climate response through aerosol-cloud interactions. These
42 results provide a more nuanced representation of the volcanic forcing over this time period that
43 can be used in state-of-the-art climate model simulations.



44 1. INTRODUCTION

45 Major volcanic eruptions inject substantial amounts of sulfur (S)-rich gases, primarily
46 sulfur dioxide (SO₂), into the troposphere and stratosphere, where they oxidize to form sulfuric
47 acid (H₂SO₄) aerosols. These aerosols scatter incoming solar radiation and can also increase
48 cloud reflectivity and cover in the troposphere through aerosol-cloud interactions, inducing
49 seasonal to multiyear cooling on hemispheric to global scales (Robock, 2000; Malavelle et al.,
50 2017; Chen et al., 2022). Such volcanically forced climate changes can contribute to major
51 societal disruption, including widespread agricultural failure and economic instability, as well
52 as associated political transitions and demographic shifts (Büntgen et al., 2026). However,
53 these interactions among volcanism, climate, and society are often complex (Degroot et al.,
54 2021), and only through detailed reconstructions of past events can we disentangle the direct
55 volcanic forcing from natural climate variability, allowing more robust interpretations of the
56 associated societal impacts.

57 The reconstruction of past volcanic activity primarily relies on sulfate (SO₄²⁻)
58 concentration records from ice cores, which underpin volcanic forcing datasets prescribed in
59 state-of-the-art climate model simulations, such as CMIP historical and PMIP past1000
60 experiments (JungCLAUS et al., 2017; Aubry et al., 2026). Synchronous sulfate peaks in both
61 Greenland and Antarctic ice cores (bipolar signals) are typically interpreted as evidence of
62 tropical eruptions whose plumes reached the stratosphere, facilitating global aerosol dispersal
63 and driving sustained cooling through the annual-to-multiannual residence time of
64 stratospheric aerosols (Robock, 2000). In contrast, sulfate peaks confined to a single
65 hemisphere (unipolar signals) are generally attributed to extratropical eruptions. However,
66 coincident high-latitude eruptions in both hemispheres can produce bipolar deposition patterns
67 that are indistinguishable from those of a single tropical eruption, potentially leading to
68 inaccuracies in the assigned volcanic forcing (e.g., Gautier et al., 2019). Furthermore, standard
69 volcanic forcing estimates derived from ice-core sulfate concentrations typically assume that
70 all sulfate deposited on polar ice sheets originates from stratospheric injection (Toohey and
71 Sigl, 2017). Yet a large proportion of ice-core sulfate from extratropical eruptions is delivered
72 via the troposphere or lowermost stratosphere (T/LMS) (Burke et al., 2023), where aerosols
73 are removed from the atmosphere within weeks to months (Toohey et al., 2019). As the
74 radiative forcing associated with volcanic eruptions depends on a range of parameters including



75 both plume height and source latitude (Marshall et al., 2019, 2021), constraints on these
76 parameters are vital for improving volcanic forcing estimates in climate model simulations.

77 Recent advances in sulfur isotope and cryptotephra analyses of polar ice cores can provide
78 these constraints. The sulfur isotope approach relies on the principle that atmospheric UV
79 photochemistry occurring within and above the ozone layer in the stratospheric overworld
80 (potential temperature >380 K; Holton et al., 1995) imprints a mass-independent fractionation
81 (MIF) signature on the volcanic sulfate produced (Savarino et al., 2003). This signature is
82 preserved in ice cores as non-zero $\Delta^{33}\text{S}$ ($= \delta^{33}\text{S} - [(\delta^{34}\text{S} + 1)^{0.515} - 1]$) values and serves as a
83 robust proxy for stratospheric sulfur injection (Savarino et al., 2003; Baroni et al., 2007;
84 Gautier et al., 2018, 2019; Burke et al., 2019, 2023). However, a $\Delta^{33}\text{S}$ value near zero does not
85 necessarily indicate a tropospheric origin; sulfur injected into the high-latitude LMS may also
86 yield sulfate lacking an MIF signal because the sulfur remains below the ozone layer (Schmidt
87 et al., 2012). Atmospheric modeling indicates that such LMS-derived sulfate mixes rapidly into
88 the troposphere, resulting in shorter atmospheric residence times than those of sulfate
89 transported via the stratospheric overworld (Toohey et al., 2019). Consequently, ice-core $\Delta^{33}\text{S}$
90 effectively distinguishes long-lived stratospheric aerosols from shorter-lived T/LMS aerosols
91 (Burke et al., 2023). In addition, the stratospheric MIF process produces a systematic, coupled
92 evolution in $\delta^{34}\text{S}$ and $\Delta^{33}\text{S}$ that is preserved in the temporal sequence of sulfate deposition
93 (Baroni et al., 2007; Gautier et al., 2018), and the timing and pattern of this isotopic evolution
94 relative to the sulfate peak depend on the transport pathway from the eruption source to the
95 poles, thereby providing a means of differentiating tropical from extratropical eruptions (Burke
96 et al., 2019). Complementing this isotopic approach, geochemical analysis of microscopic
97 volcanic ash particles (i.e. cryptotephra) can constrain the geographic source of an eruption
98 (Plunkett et al., 2023; Hutchison et al., 2024, 2025). These particles are closely associated with
99 the ice core sulfate deposit (usually occurring at or just before the sulfate peak), and recent
100 improvements in sample preparation and electron microprobe analysis (EPMA, using 3 and 1
101 μm beams) are allowing identification of cryptotephra that were previously missed or deemed
102 too small for geochemical characterization (Innes et al., 2024; Ro et al., 2025; Smith et al.,
103 2026).

104 An important period of clustered eruptive activity occurs in the 12th and 13th centuries
105 (Guillet et al., 2023; Büntgen et al., 2026). This period begins with a major volcanic sulfate



106 deposition event in 1108 CE, which is the 13th largest such event recorded in polar ice cores
107 over the past 2500 years (Sigl et al., 2015; Guillet et al., 2020). This volcanic episode likely
108 triggered one of the most severe Northern Hemisphere (NH) summer cooling events of the past
109 1500 years in 1109 CE (summer temperature anomalies of approximately -1 to -1.3 °C relative
110 to 1961–1990; Guillet et al., 2020) and induced anomalous precipitation across Western Europe;
111 these conditions contributed to widespread crop failures and food price inflation, establishing
112 the environmental preconditions for the devastating subsistence crises and famines between
113 1109 and 1111 CE (Guillet et al., 2020). Although this volcanic activity was initially interpreted
114 as a single tropical eruption (Sigl et al., 2015), Guillet et al. (2020) proposed that it instead
115 represented a multi-eruption cluster, based on high-resolution bipolar ice-core records that
116 revealed several sulfate peaks in Greenland between 1108 and 1113 CE and a distinct peak in
117 Antarctica in late 1109 CE. They also reported documentary evidence of a dark total lunar
118 eclipse across Europe in May 1110 CE, indicating the presence of a stratospheric aerosol veil.
119 Nevertheless, the eruptive nature of this volcanic episode—the number of eruptions involved
120 and their plume injection heights, source latitudes, and provenances—remains poorly
121 constrained.

122 Guillet et al. (2020) proposed three potential scenarios: (1) a tropical eruption in 1108 CE
123 together with two NH extratropical eruptions, one near-simultaneous and the other subsequent;
124 (2) multiple NH extratropical eruptions from 1108 CE onwards, combined with a subsequent
125 tropical eruption recorded in late 1109 CE; or (3) multiple NH extratropical eruptions from
126 1108 CE onwards without any tropical eruption, accompanied by an independent Southern
127 Hemisphere (SH) extratropical eruption in late 1108 to early 1109 CE. They also suggested that
128 the Volcanic Explosivity Index (VEI) 5 eruption of Asama (36.4° N, 138.5° E, Japan) on 29
129 August 1108 CE may have been the source of one of the NH extratropical eruptions.
130 Unfortunately, ice-core sulfate concentration records cannot differentiate between the tropical
131 and extratropical origins of closely spaced eruptions or determine whether their sulfur was
132 injected into the stratosphere. In addition, no diagnostic tephra has yet been reported, leaving
133 the proposed attribution to Asama untested. Other VEI 5 eruptions, such as Mashu (1125 ± 95
134 CE; 43.6° N, 144.5° E, Japan) and Ubinas (1082 ± 82 CE; 16.4° S, 70.9° W, Peru), also have
135 dates that fall within the 1108–1110 CE interval, further complicating source attribution. These
136 uncertainties prevent a robust assessment of the potential climatic forcing associated with this



137 volcanic episode.

138 In this study, we present new high-resolution sulfur isotope records and geochemical
139 compositions of cryptotephra glass shards from the North Greenland Ice Core Project 1
140 (NGRIP1) ice core, covering 1106–1112 CE, together with sulfur isotope records from the
141 TUNU2013 ice core, spanning 1107–1113 CE. These data allow us to test directly whether this
142 volcanic episode represents a single eruption or multiple eruptions, and to constrain the plume
143 injection height, source latitude, and provenance of each event, thereby providing a revised
144 estimate of the associated volcanic forcing and improved constraints on its timing.

145

146 2. MATERIALS AND METHODS

147 2.1. Ice core samples

148 This study used two ice cores drilled in Greenland: NGRIP1 (75.1° N, 42.3° W, 2917 m
149 a.s.l.) and TUNU2013 (78.0° N, 33.9° W, 2105 m a.s.l.; **Fig. 1a**). The NGRIP1 samples were
150 obtained from a depth interval of 188.65–189.58 m, corresponding to 1106–1112 CE based on
151 the GICC21 chronology (Sinnl et al., 2022). The TUNU2013 samples were obtained from a
152 depth interval of 120.46–121.13 m, corresponding to 1107–1113 CE based on the
153 DRI_NGRIP2 chronology (McConnell et al., 2018, 2019).

154 For sulfur isotope and cryptotephra analyses, the outer edges of the ice-core sticks were
155 scraped to remove potential surface contamination. Discrete samples were then cut across each
156 sulfate peak at a depth resolution of 3.1–4.5 cm for NGRIP1 and 3.0–4.3 cm for TUNU2013,
157 corresponding to approximately 5 and 3 samples per year, respectively. Background samples
158 were also collected from intervals preceding the volcanic sulfate peaks at a depth resolution of
159 20 cm for NGRIP1 and 4 cm for TUNU2013, corresponding to approximately 1 year and 3
160 months of accumulation, respectively. Following melting, the NGRIP1 samples were
161 centrifuged, and the supernatant was decanted for isotope analysis, while the remaining 1–2
162 mL of concentrate was reserved for cryptotephra sampling. The melted TUNU2013 samples
163 were used exclusively for isotope analysis.

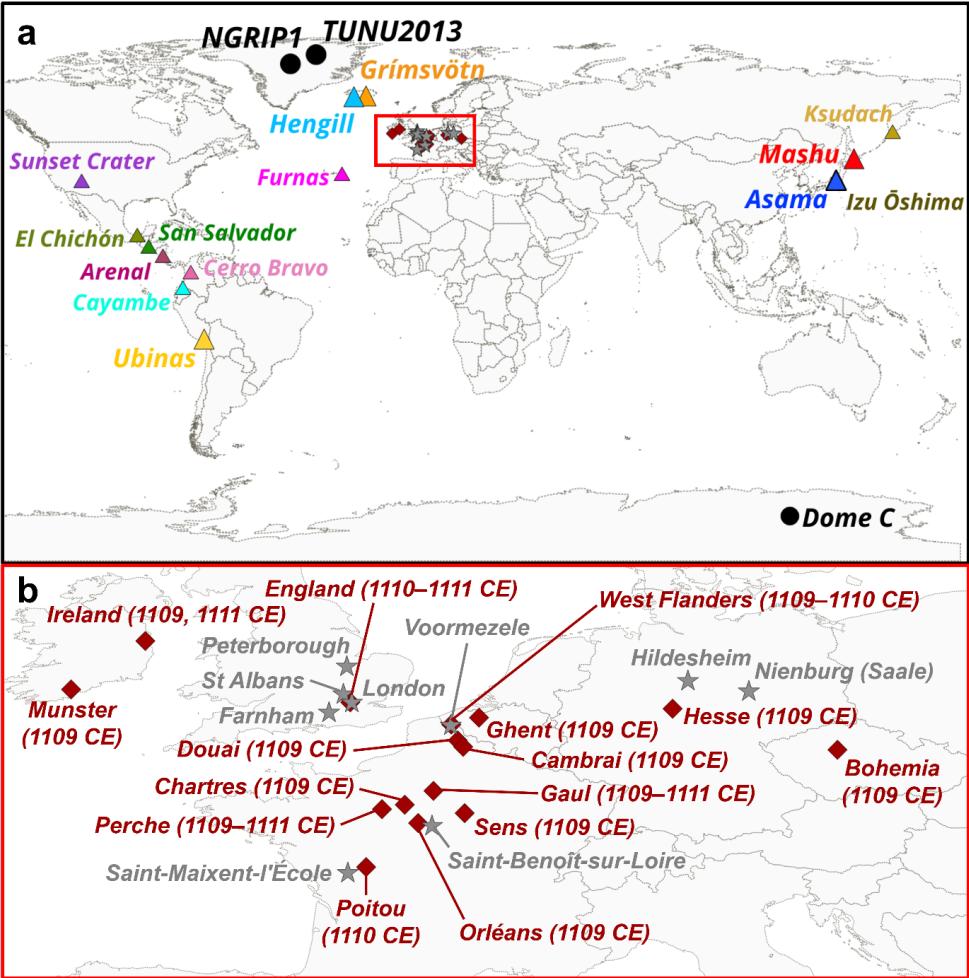


Figure 1. (a) A map showing the locations of NGRIP1 and TUNU2013 in Greenland and Dome C in Antarctica (black circles) and the volcanoes discussed in Sect. 3 and Tables 1 and 2 (triangles). The red rectangle indicates the area enlarged in (b). (b) The places where the total lunar eclipse of 5 May 1110 CE was recorded (grey stars) and the regions that experienced severe societal impacts between 1109 and 1111 CE (dark-red diamonds). Historical records are compiled from Guillet et al. (2020).



2.2. Sulfur isotopes

All sample preparation and analysis for sulfur isotopes were conducted at the St Andrews Isotope Geochemistry (STAiG) Laboratory, University of St Andrews (Burke et al., 2019, 2023). First, a 600 μL aliquot of each sample was analyzed for sulfate concentration using a 930 Compact IC Flex ion chromatograph (Metrohm, Switzerland). Based on the measured concentrations, aliquots containing 20 nmol of sulfate were evaporated to dryness in Teflon vials on a hotplate at 104 $^{\circ}\text{C}$ and re-dissolved in 100 μL of 0.01 % (v/v) distilled HCl. The resulting NGRIP1 and TUNU2013 solutions were transferred to autosampler vials for sulfate purification using an automated prepFAST MC system (Elemental Scientific Inc., USA) and manual column chemistry, respectively.

The automated prepFAST MC system was equipped with a single PFA column containing two PTFE frits and 50 μL of AG1-X8 resin. Prior to loading the NGRIP1 samples, the resin was sequentially pre-conditioned with 600 μL of 1.1 M HCl, 1 mL of Milli-Q water, 600 μL of 1.6 M HNO_3 , an additional 600 μL of 1.1 M HCl, and 600 μL of 0.6 M HCl. After sample loading, the column was rinsed with 750 μL of Milli-Q water to remove the matrix, and the sulfate fraction was then eluted and collected in 550 μL of 0.5 M HNO_3 .

For the manual column chemistry procedure (Burke et al., 2019), the TUNU2013 samples were loaded onto 35 μL of AG1-X8 anion-exchange resin within Teflon columns constructed from 4:1 PTFE shrink tubing. Prior to sample loading, the resin was pre-conditioned with 350 μL of 1.6 M HNO_3 , 350 μL of 1.2 M HCl, and 350 μL of 0.6 M HCl. After sample loading, the column was rinsed with 3×175 μL of Milli-Q water to remove the matrix, and the sulfate fraction was then eluted and collected in 3×70 μL of 0.5 M HNO_3 .

The collected eluates were subsequently evaporated to dryness in Teflon vials on a hotplate at 104 $^{\circ}\text{C}$ and re-dissolved in 0.5 M HNO_3 containing NaOH to yield a final concentration of 20 μM Na_2SO_4 , thereby matching the matrix of our in-house bracketing standards. Finally, triple sulfur isotopes (^{32}S , ^{33}S , and ^{34}S) were measured using a Neptune Plus (Thermo Fisher Scientific, USA) multicollector inductively coupled plasma mass spectrometer (MC-ICP-MS). Instrument accuracy and precision were monitored by preparing and analyzing in-house secondary standards (Switzer Falls river water: $\delta^{34}\text{S} = 4.11 \pm 0.07$ ‰; filtered seawater: $\delta^{34}\text{S} = 21.16 \pm 0.07$ ‰), an IAEA-certified reference material (IAEA-SO6: $\delta^{34}\text{S} = -34.19 \pm 0.07$ ‰), and a NIST-certified reference material (NBS-127: $\delta^{34}\text{S} = 21.17 \pm 0.07$ ‰) under the



202 same conditions as the samples. These values show good agreement with previously reported
203 values (e.g., Burke et al., 2019, 2023). The mean and standard deviation (σ) of the measured
204 procedural blanks (0.05 ± 0.02 nmol S, $\delta^{34}\text{S} = 2.00 \pm 0.57$ ‰) were used to correct the sample
205 isotope ratios.

206 All sulfur isotope values ($\delta^{34}\text{S}$ and $\delta^{33}\text{S}$; $\delta^x\text{S} = ({}^x\text{S}/{}^{32}\text{S})_{\text{sample}}/({}^x\text{S}/{}^{32}\text{S})_{\text{VCDT}} - 1$) are reported
207 relative to the Vienna Canyon Diablo Troilite (VCDT) standard. Background sulfate
208 concentrations and isotope ratios were used to isolate the volcanic sulfate component,
209 following Baroni et al. (2007):

$$210 \quad \delta^x\text{S}_{\text{volc}} = (\delta^x\text{S}_{\text{meas}} - \delta^x\text{S}_{\text{bkgd}}f_{\text{bkgd}})/f_{\text{volc}} \quad (1)$$

211 Here, $\delta^x\text{S}_{\text{volc}}$, $\delta^x\text{S}_{\text{meas}}$, and $\delta^x\text{S}_{\text{bkgd}}$ denote the $\delta^{34}\text{S}$ or $\delta^{33}\text{S}$ values of the volcanic, measured, and
212 background sulfate, respectively. $f_{\text{bkgd}} (= [\text{SO}_4]_{\text{bkgd}}/[\text{SO}_4]_{\text{sample}})$ and $f_{\text{volc}} (= 1 - f_{\text{bkgd}})$ represent
213 the mass fractions of background and volcanic sulfate, respectively. Uncertainties on this
214 correction were propagated using Monte Carlo methods (Burke et al., 2023). Samples
215 containing >50 % background sulfate yielded uncertainties too large for interpretation and were
216 therefore excluded from the analysis.

217 Furthermore, an isotope mass balance approach was employed to quantify the fraction of
218 sulfate originating from the stratospheric overworld (Burke et al., 2019, 2023). The measured
219 sulfate was modeled as a mixture of stratospheric overworld, T/LMS, and background
220 components:

$$221 \quad \delta^{34}\text{S}_{\text{meas}} = \delta^{34}\text{S}_{\text{strat}}f_{\text{strat}} + \delta^{34}\text{S}_{\text{trop}}f_{\text{trop}} + \delta^{34}\text{S}_{\text{bkgd}}f_{\text{bkgd}} \quad (2)$$

$$222 \quad \delta^{33}\text{S}_{\text{meas}} = \delta^{33}\text{S}_{\text{strat}}f_{\text{strat}} + \delta^{33}\text{S}_{\text{trop}}f_{\text{trop}} + \delta^{33}\text{S}_{\text{bkgd}}f_{\text{bkgd}} \quad (3)$$

223 where f_{strat} and f_{trop} are the proportions of sulfate from the stratospheric overworld and the
224 T/LMS, respectively, with $f_{\text{strat}} + f_{\text{trop}} + f_{\text{bkgd}} = 1$. The $\delta^{33}\text{S}$ value of each component is related
225 to its $\delta^{34}\text{S}$ value through the following fractionation relationships:

$$226 \quad \delta^{33}\text{S}_{\text{trop}} = ((\delta^{34}\text{S}_{\text{trop}} + 1)^{0.515} - 1) \quad (4)$$

$$227 \quad \delta^{33}\text{S}_{\text{bkgd}} = ((\delta^{34}\text{S}_{\text{bkgd}} + 1)^{0.515} - 1) \quad (5)$$

$$228 \quad \delta^{33}\text{S}_{\text{strat}} = ((\delta^{34}\text{S}_{\text{strat}} + 1)^{0.608} - 1) \quad (6)$$

229 Substituting Eqs. (4)–(6) into Eqs. (2) and (3) reduces the system to two equations with three
230 unknowns (f_{strat} , $\delta^{34}\text{S}_{\text{strat}}$, $\delta^{34}\text{S}_{\text{trop}}$). Given that $\delta^{34}\text{S}_{\text{trop}}$ can vary widely (Jongebloed et al., 2023),
231 we varied $\delta^{34}\text{S}_{\text{trop}}$ between -10 ‰ and 10 ‰ and numerically solved for f_{strat} and $\delta^{34}\text{S}_{\text{strat}}$ via
232 Monte Carlo simulations, following the implementation described in Burke et al. (2023).



233

234 **2.3. Cryptotephra glass shards**

235 All sample preparation and analysis of cryptotephra glass shards were conducted at the
236 University of St Andrews following the procedure of [Innes et al. \(2024\)](#). Subsamples were
237 pipetted onto the centers of plastic stubs and dried at 85 °C on a hotplate covered with Kapton
238 tape. Once fully dry, the stubs were embedded in EpoThin 2 epoxy resin (Buehler, USA). The
239 surfaces were lightly polished for 30 s to 2 min using 3 and 1 µm diamond pastes, followed by
240 a 0.3 µm aluminum oxide slurry, and subsequently carbon-coated.

241 Major and minor element compositions (Na, Mg, Al, Si, P, K, Ca, Ti, Mn, and Fe) were
242 determined using a JXA-iSP100 electron probe microanalyzer (EPMA; JEOL, Japan) equipped
243 with an energy-dispersive spectrometer (EDS) and five wavelength-dispersive spectrometers
244 (WDS). The operating conditions were set to an accelerating voltage of 15 kV, a beam current
245 of 1 nA, and a beam diameter of 3 µm. Peak counting times were as follows: 50 s for Mn; 30 s
246 for Si, Ti, Fe, Ca, and P; 20 s for Al, Mg, and K; and 7 s for Na. Secondary standards (ML3B-
247 G, StHs6/80-G, and Lipari obsidian; [Jochum et al., 2006](#); [Kuehn et al., 2011](#)) were analyzed
248 concurrently to monitor accuracy and precision (**Table S1**). All geochemical data were
249 normalized to 100 % on an anhydrous basis.

250

251 **2.4. Calculation of the revised volcanic stratospheric sulfur injection for the 1108–1110** 252 **CE volcanic episode**

253 We calculated the revised volcanic stratospheric sulfur injection (VSSI) for the 1108–1110
254 CE episode following the procedure described in [Sigl et al. \(2014\)](#) and [Gabriel et al. \(2025\)](#)
255 and applying the stratospheric sulfate fractions derived from our new sulfur isotope records
256 and isotope mass balance calculations (see **Sect. 2.2**). Briefly, volcanic sulfate signals were
257 first isolated from natural background signals in previously reported bipolar ice-core records
258 (Greenland: NGRIP1, NEEM-2011-S1, TUNU2013, TUNU2022A, TUNU2022B, and GISP2,
259 n = 6; Antarctica: WDC06A, SP1, SP4, SPC14, DML05, DML07, B40, NUS08-5, NUS07-5,
260 NUS07-7, DF01, DFS10, EDC96, and Taylor Dome, n = 14; **Figs. S1 and S2**). The non-
261 volcanic background was defined using a 91-year running median of annually averaged non-
262 sea-salt (nss) or total sulfate records across the 950–1850 CE period. Volcanic sulfate flux was
263 derived by subtracting this background from the total sulfate data and multiplying the residual



264 by site-specific accumulation rates. Cumulative deposition (kg km^{-2}) was then summed over
265 the event duration, defined by upper ($K = 3$) and lower ($K = 1$ for Greenland; $K = 1.5$ for
266 Antarctica) thresholds, where K denotes the multiple of the median absolute deviation (MAD),
267 as validated against historical eruptions (Gabriel et al., 2025). We then estimated the mean
268 cumulative volcanic sulfate deposition for each ice sheet; a simple arithmetic mean was
269 calculated for Greenland, whereas a representative mean for Antarctica was derived by
270 applying a weighting of 20 % to high-accumulation sites and 80 % to low-accumulation sites
271 (Toohey and Sigl, 2017).

272 We then used the sub-annually resolved NGRIP1 nss-sulfate record reported by Sigl et al.
273 (2015) to separate the overlapping sulfate signals within the 1108–1110 CE episode into
274 individual events. This high-resolution record allowed us to approximate the start and end dates
275 of individual events and calculate the fractional contribution of each event to the cumulative
276 volcanic sulfate deposition, ensuring more robust quantification of sulfur loading for individual
277 eruptions within a multi-year cluster (see Sects. 3.1 and 3.3). The mean cumulative volcanic
278 sulfate deposition on the Greenland Ice Sheet was then apportioned among the individual
279 events. By contrast, all volcanic sulfate deposition associated with this episode on the Antarctic
280 Ice Sheet was attributed to the event interpreted as tropical in origin.

281 Finally, VSSI was estimated assuming that ice-sheet sulfate deposition is proportional to
282 the injected sulfur mass (M_s ; Tg S). Using the mean cumulative deposition for Greenland and
283 Antarctica (D_G and D_A , respectively; $\text{kg SO}_4^{2-} \text{ km}^{-2}$), we calculated M_s as follows:

$$284 \quad M_s = \frac{1}{3} [L_G D_G + L_A D_A] \quad (7)$$

285 where L_G and L_A are transfer functions representing NH and SH spatial distributions,
286 respectively. We applied transfer function values of $1 \times 10^9 \text{ km}^2$ for tropical and $0.57 \times 10^9 \text{ km}^2$
287 for NH extratropical eruptions (Toohey and Sigl, 2017; Sigl et al., 2022). The final revised
288 VSSI for this episode was obtained by multiplying M_s for each event by its stratospheric sulfate
289 fraction and summing the resulting values.

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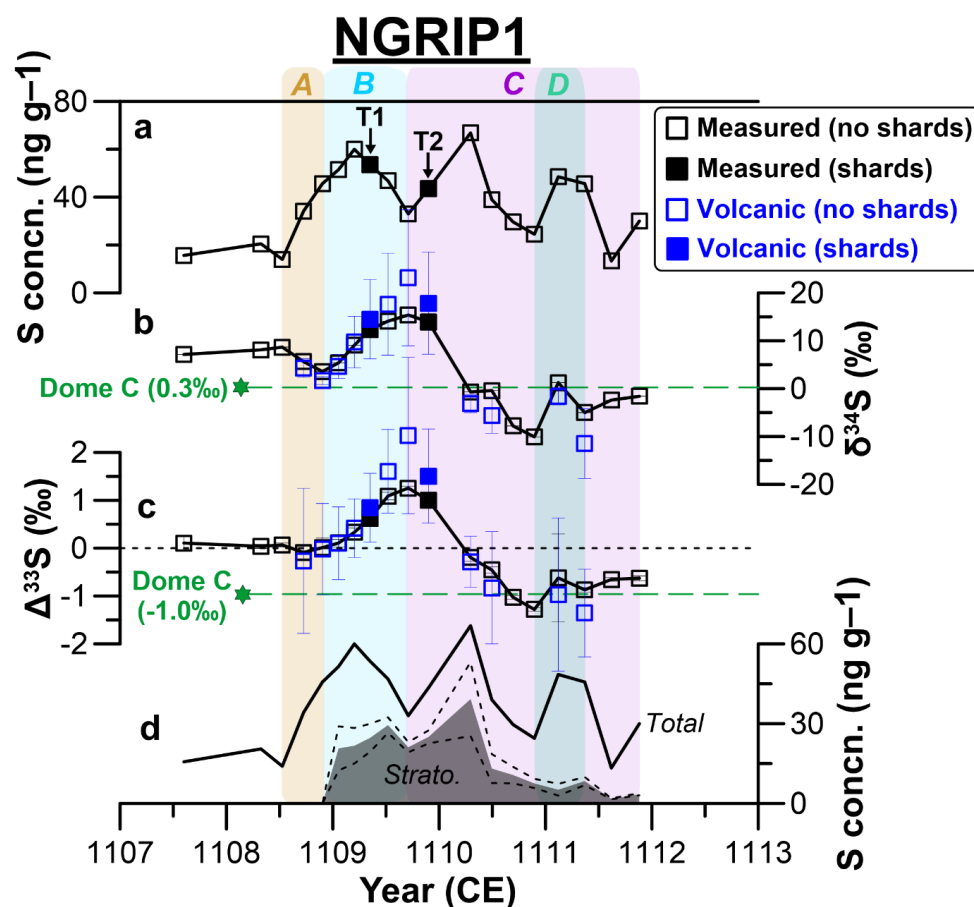
291 3. RESULTS AND DISCUSSION

292 3.1. Sulfur isotopes: Constraints on plume injection height and source latitude



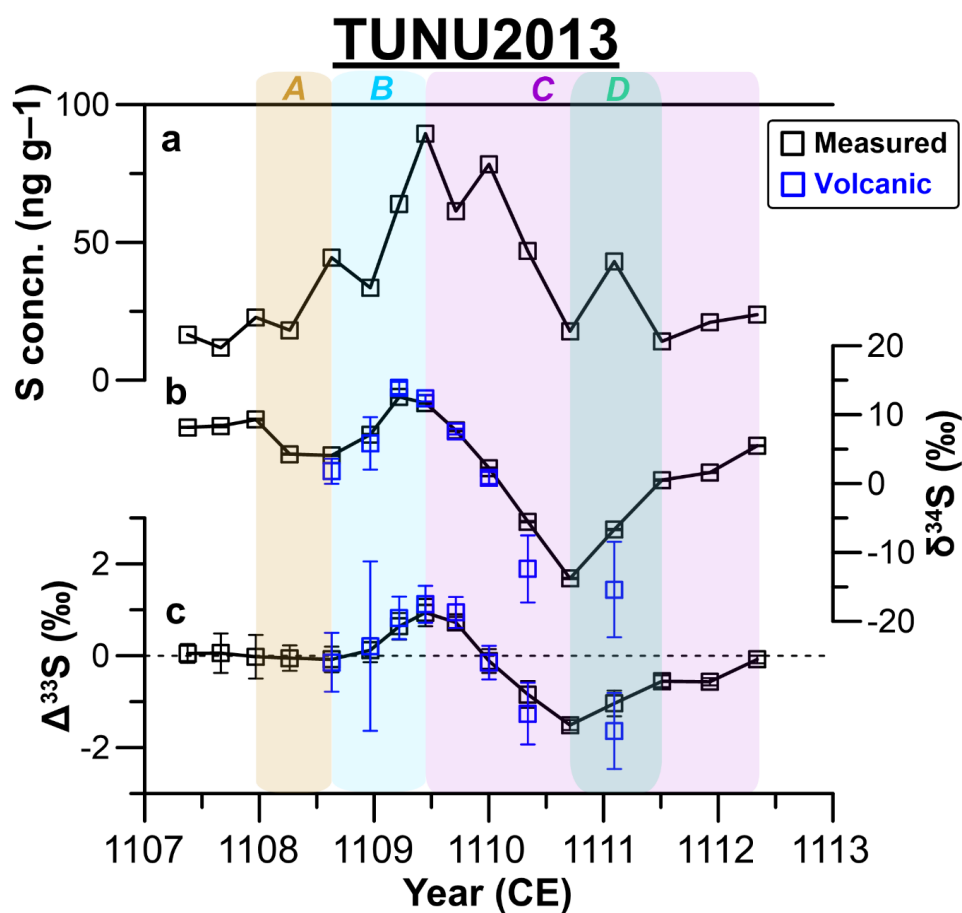
293 Sulfate concentrations and isotopic compositions of the NGRIP1 and TUNU2013 ice-core
294 samples are shown in **Figs. 2 and 3** and listed in **Table S2**. The NGRIP1 record exhibits three
295 main sulfate peaks at 1109.2, 1110.3, and 1111.1 CE, with concentrations of 60, 67, and 49 ng
296 S g⁻¹, respectively. The corresponding peaks in TUNU2013 occur at 1109.4, 1110.0, and
297 1111.1 CE, reaching 90, 78, and 43 ng S g⁻¹, respectively. Measured $\delta^{34}\text{S}$ ranges from -10.1 to
298 15.4 ‰ in NGRIP1 and from -13.8 to 12.6 ‰ in TUNU2013, and $\Delta^{33}\text{S}$ from -1.3 to 1.3 ‰ in
299 NGRIP1 and from -1.5 to 0.9 ‰ in TUNU2013.

300 The non-volcanic background samples yield $\delta^{34}\text{S}$ values of 7.1 ‰ for NGRIP1 and 8.2–
301 8.4 ‰ for TUNU2013, with $\Delta^{33}\text{S}$ indistinguishable from zero within uncertainty. These values
302 fall within the range of previously reported background $\delta^{34}\text{S}$ values for Greenland ice cores (7–
303 14 ‰; [Burke et al., 2019, 2023](#); [Jongebloed et al., 2023](#); [Hutchison et al., 2024](#)). The initial
304 isotopic evolution around 1108 CE is characterized by a decline in $\delta^{34}\text{S}$ from 8.7 to 3.6 ‰ in
305 NGRIP1 and from 9.3 to 4.1 ‰ in TUNU2013, while $\Delta^{33}\text{S}$ remains near zero. These changes
306 are accompanied by an overall increase in sulfate concentrations from 14 to 46 ng S g⁻¹ in
307 NGRIP1 and from 23 to 44 ng S g⁻¹ in TUNU2013 over approximately half a year (section A
308 in **Figs. 2 and 3**). Subsequently, both $\delta^{34}\text{S}$ and $\Delta^{33}\text{S}$ increase, reaching maxima in late 1109 CE
309 in NGRIP1 ($\delta^{34}\text{S} = 15.4$ ‰, $\Delta^{33}\text{S} = 1.3$ ‰) and in early to mid-1109 CE in TUNU2013 ($\delta^{34}\text{S}$
310 $= 12.6$ ‰, $\Delta^{33}\text{S} = 0.9$ ‰). These isotopic maxima occur approximately half a year after the first
311 peak in NGRIP1 but are nearly synchronous with it in TUNU2013 (section B). Thereafter, $\delta^{34}\text{S}$
312 and $\Delta^{33}\text{S}$ decline overall, with $\Delta^{33}\text{S}$ changing from positive to negative values around the
313 second sulfate peak, and reach negative minima in late 1110 CE (NGRIP1: $\delta^{34}\text{S} = -10.1$ ‰,
314 $\Delta^{33}\text{S} = -1.3$ ‰; TUNU2013: $\delta^{34}\text{S} = -13.8$ ‰, $\Delta^{33}\text{S} = -1.5$ ‰), followed by a general increase
315 toward background values (section C). Within this interval, the third sulfate peak, which
316 emerges shortly after the isotopic minimum, is accompanied by transient isotopic excursions
317 in NGRIP1, with $\delta^{34}\text{S}$ increasing to 1.2 ‰ (compared with -10.1 ‰ immediately before and –
318 5.0 ‰ after) and $\Delta^{33}\text{S}$ to -0.6 ‰ (compared with -1.3 ‰ before and -0.9 ‰ after; section D
319 in **Fig. 2**). No comparable excursion is evident in TUNU2013 (section D in **Fig. 3**).



320

321 **Figure 2.** Sulfur concentration and isotopic records from the NGRIP1 ice core samples for the
 322 1108–1110 CE eruption episode. (a) Sulfur concentration, (b) $\delta^{34}\text{S}$, and (c) $\Delta^{33}\text{S}$ records (black
 323 squares). For samples with more than 50 % volcanic sulfate, estimates of the volcanic sulfur
 324 isotopic composition—calculated from isotope mass balance (see Sect. 2.2)—are plotted as
 325 blue squares in (b) and (c). Uncertainties for $\delta^{34}\text{S}$ and $\Delta^{33}\text{S}$ are represented as $\pm 1\sigma$. $\delta^{34}\text{S}$ and
 326 $\Delta^{33}\text{S}$ values measured from the Antarctic Dome C ice core during the 1108–1110 CE volcanic
 327 episode (Gautier et al., 2019) are also shown in (b) and (c) as green stars and dashed lines.
 328 Open symbols indicate samples where no glass shards were detected, while solid symbols
 329 indicate samples where glass shards were found (T1: 189.14–189.17 m, 1109.4 CE; T2:
 330 189.02–189.06 m, 1109.9 CE). (d) Sulfur concentration as in (a); filled areas represent the
 331 estimated concentration of volcanic sulfate originating from the stratospheric overworld based
 332 on the isotope mass balance calculations. Dashed lines represent $\pm 1\sigma$ of the stratospheric sulfate
 333 fraction estimated from Monte Carlo simulations. Gold, light-blue, purple, and teal shaded
 334 bands represent sections A, B, C, and D, respectively (see Sect. 3.1).



335

336 **Figure 3.** Sulfur concentration and isotopic records from the TUNU2013 ice core samples for
 337 the 1108–1110 CE eruption episode. (a) Sulfur concentration, (b) $\delta^{34}\text{S}$, and (c) $\Delta^{33}\text{S}$ records
 338 (black squares). For samples with more than 50 % volcanic sulfate, estimates of the volcanic
 339 sulfur isotopic composition—calculated from isotope mass balance (see Sect. 2.2)—are plotted
 340 as blue squares in (b) and (c). Uncertainties for $\delta^{34}\text{S}$ and $\Delta^{33}\text{S}$ are represented as $\pm 1\sigma$. Gold,
 341 light-blue, purple, and teal shaded bands represent sections A, B, C, and D, respectively (see
 342 Sect. 3.1).



343 Overall, the two ice-core records show good agreement in their temporal variations,
344 providing a consistent reconstruction of volcanic sulfate isotopic compositions during the early
345 12th century. The minor discrepancies in the detailed timing and magnitude of the variations
346 between NGRIP1 and TUNU2013 may reflect differences in sampling resolution
347 (approximately 5 and 3 samples per year, respectively; see **Sect. 2.1**), sub-annual age
348 interpolation in the two chronologies (GICC21 and DRI_NGRIP2), and post-depositional
349 stratigraphic noise arising from wind-driven snow redistribution and surface roughness at the
350 drilling sites (Wolff et al., 2005; Gautier et al., 2016). Given the lower accumulation rate at
351 TUNU2013 (0.10 m w.e. yr⁻¹; Sigl et al., 2015) compared with NGRIP1 (0.17 m w.e. yr⁻¹;
352 North Greenland Ice Core Project members, 2004), the TUNU2013 record is more susceptible
353 to such post-depositional alteration. We therefore primarily use the NGRIP1 record to
354 investigate the temporal variability of the volcanic sulfate signals.

355 The first sulfate peak is preceded by a decrease in $\delta^{34}\text{S}$ with no MIF signature (section A
356 in **Fig. 2a–c**), indicative of an influx of volcanic sulfate transported through the T/LMS from
357 an extratropical eruption (Burke et al., 2019). This interpretation is consistent with the
358 negligible stratospheric sulfate contribution estimated for this interval from the isotope mass
359 balance calculations (**Fig. 2d**). A positive MIF signature subsequently develops, with the
360 isotopic maxima lagging the first sulfate peak by approximately half a year in NGRIP1 (section
361 B). This sequence, an initial non-MIF sulfate influx followed by stratospheric sulfate, differs
362 from the pattern typical of tropical eruptions. Because volcanic sulfate from the tropics can
363 reach the poles exclusively via the stratospheric overworld, where the MIF occurs, tropical
364 eruptions are characterized by large positive $\delta^{34}\text{S}$ and $\Delta^{33}\text{S}$ anomalies at the onset of sulfate
365 deposition, which precede the sulfate concentration peak by 6 months to a year and then
366 decrease monotonically across the rest of the deposition (Burke et al., 2019). Instead, the
367 observed sequence resembles the pattern documented for well-known extratropical eruptions,
368 such as the 1912 CE Katmai/Novarupta eruption in Alaska (Burke et al., 2019). We therefore
369 attribute the first sulfate peak predominantly to an NH extratropical eruption that injected sulfur
370 into both the T/LMS and the stratospheric overworld, with sulfate deposition beginning in mid-
371 1108 CE. The negative MIF signature expected in the later stage of this signal is not observed,
372 likely because it was masked by the arrival of positive-MIF sulfate from a subsequent eruption
373 (see below). The isotope mass balance calculations yield a stratospheric sulfate fraction of 37



374 $\pm 8 \%$ (1σ) for the volcanic sulfate deposited over the interval attributed to this eruption
375 (sections A and B in **Fig. 2d**).

376 The second sulfate peak is also associated with a clear MIF signature (section C in **Fig.**
377 **2**), characteristic of volcanic sulfate transported through the stratospheric overworld ([Gautier](#)
378 [et al., 2018](#); [Burke et al., 2019](#)). However, the close temporal spacing of the first and second
379 peaks makes it difficult to resolve the isotopic signatures associated with the end of the first
380 deposition event and the onset of the second. This ambiguity complicates the latitudinal
381 attribution of the source of the second peak, because the isotope pattern at the onset of sulfate
382 deposition is key to distinguishing between extratropical and tropical eruptions ([Burke et al.,](#)
383 [2019, 2023](#)). Nevertheless, a non-zero $\Delta^{33}\text{S}$ value of -1.0 ‰ reported for the corresponding
384 volcanic signal in the Antarctic Dome C ice core (**Fig. 2c**; [Gautier et al., 2019](#)) indicates that
385 substantial stratospheric sulfate also reached the SH. This evidence points to a tropical eruption
386 as the most likely source of the second peak, although high-resolution sulfur isotope records
387 from Antarctica are required to definitively rule out a coincident SH stratospheric eruption.
388 This highlights the importance of paired sulfur isotope records from both hemispheres for
389 characterizing the eruptive nature of closely spaced eruptions.

390 Finally, the third sulfate peak and the accompanying transient isotopic excursions in early
391 1111 CE (section D in **Fig. 2**) likely reflect an input of volcanic sulfate transported via the
392 T/LMS. Such an input would have $\delta^{34}\text{S}$ values of $0\text{--}5 \text{ ‰}$ ([Nielsen et al., 1991](#)) and $\Delta^{33}\text{S}$ of ~ 0
393 ‰ and would explain the shift of $\delta^{34}\text{S}$ toward more positive values and of $\Delta^{33}\text{S}$ toward 0 ‰
394 during an interval otherwise dominated by the negative isotopic signatures of the tropical
395 eruption. A comparable shift could in principle also be produced by an input of biogenic sulfur,
396 which is characterized by substantially higher $\delta^{34}\text{S}$ values ($15.6\text{--}22.0 \text{ ‰}$; [Calhoun et al., 1991](#);
397 [McArdle and Liss, 1995](#); [Sanusi et al., 2006](#); [Amrani et al., 2013](#)). However, a biogenic origin
398 is difficult to reconcile with the observations, because the sharp increase in sulfate
399 concentration is atypical of biogenic inputs and the high $\delta^{34}\text{S}$ values of biogenic sulfur would
400 produce a substantially larger positive shift in $\delta^{34}\text{S}$ than observed. Consequently, the third
401 sulfate peak is interpreted as the signature of a minor NH extratropical eruption. Although the
402 isotope mass balance calculations indicate some stratospheric sulfate contribution during this
403 interval, we attribute this to the residual MIF signal of stratospheric sulfate from the tropical
404 eruption rather than to the minor eruption itself.

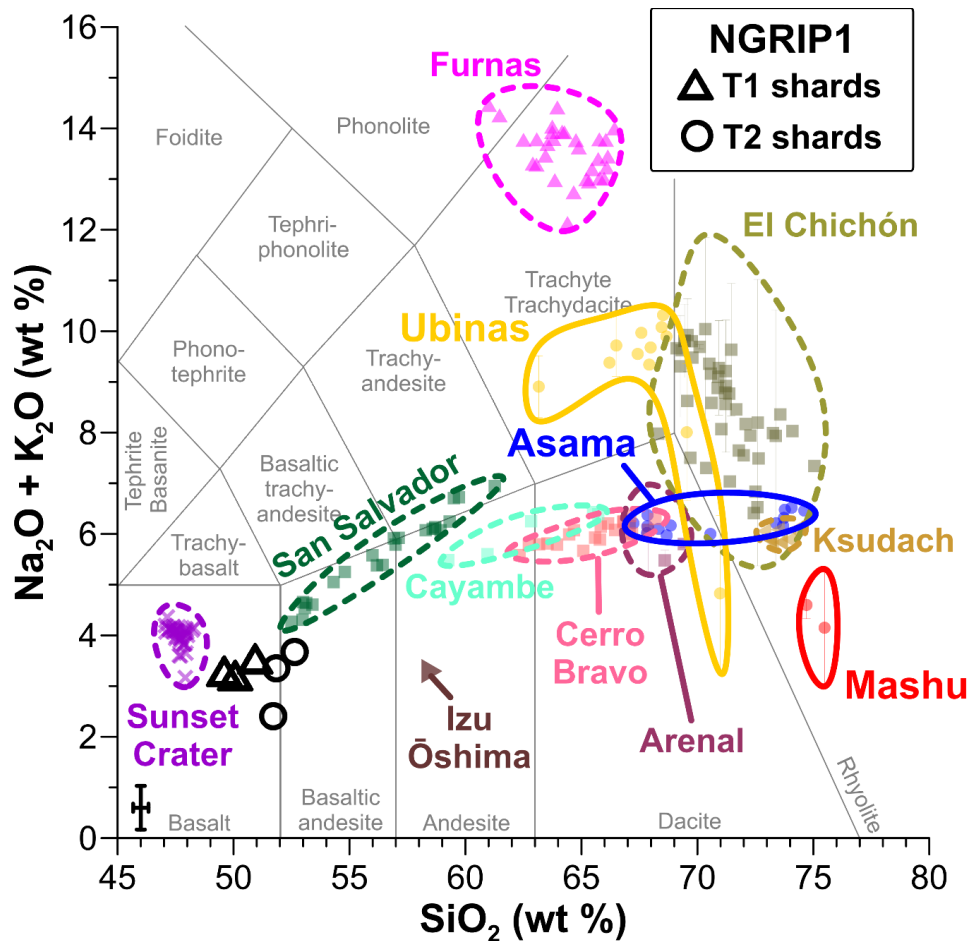


405 Overall, our sulfur isotope records indicate that the 1108–1110 CE volcanic episode
406 comprises an NH extratropical eruption beginning in mid-1108 CE, which primarily accounts
407 for the first sulfate peak in early 1109 CE (sections A and B in **Fig. 2**); a subsequent tropical
408 eruption, which accounts for the second peak in early 1110 CE (section C); and a minor NH
409 extratropical eruption, which accounts for the third peak in early 1111 CE (section D).

410

411 **3.2. Cryptotephra geochemistry: Constraints on volcanic provenance**

412 Among the 19 NGRIP1 samples examined, seven volcanic glass shards were identified
413 (**Fig. S3**) and chemically analyzed from two discrete depths: 189.15 m ($n = 4$; T1) and 189.04
414 m ($n = 3$; T2), corresponding to 1109.4 and 1109.9 CE, respectively (shown as solid squares in
415 **Fig. 2**). Full geochemical data for all measured shards are provided in **Table S1**. The identified
416 shards were typically micrometer-sized, making them challenging to polish and occasionally
417 too small for a 3 μm electron beam (see **Sect. 2.3**), which resulted in analytical totals of 68–
418 96 %. Nevertheless, these shards yielded geochemically coherent populations after
419 normalization. Given the limited shard count, we considered whether they were related to
420 reworking and redeposition of ash at high-elevation Greenland ice-core sites. Although
421 [Plunkett et al. \(2020\)](#) suggested that such remobilized shards are rare in Greenland, we applied
422 a conservative screening criterion by excluding tephra with normalized Na_2O values <1 wt %
423 ([Hutchison et al., 2024](#)), because sodium is among the major elements most readily leached by
424 meteoric fluids during environmental exposure ([Cerling et al., 1985](#)). On this basis, one dacitic
425 shard from T1 was excluded ($\text{Na}_2\text{O} = 0.8$ wt %; **Table S1**). Of the remaining shards, the three
426 basaltic T1 shards are characterized by lower SiO_2 (mean $\pm \sigma$: 50.2 ± 0.7 wt %) and higher
427 TiO_2 (3.1 ± 0.03 wt %) and FeO (13.4 ± 0.2 wt %) values than the T2 shards (52.1 ± 0.5 , $2.5 \pm$
428 0.04 , and 12.5 ± 0.3 wt %, respectively), with the latter straddling basalt and basaltic andesite
429 compositions (**Figs. 4 and 5**).



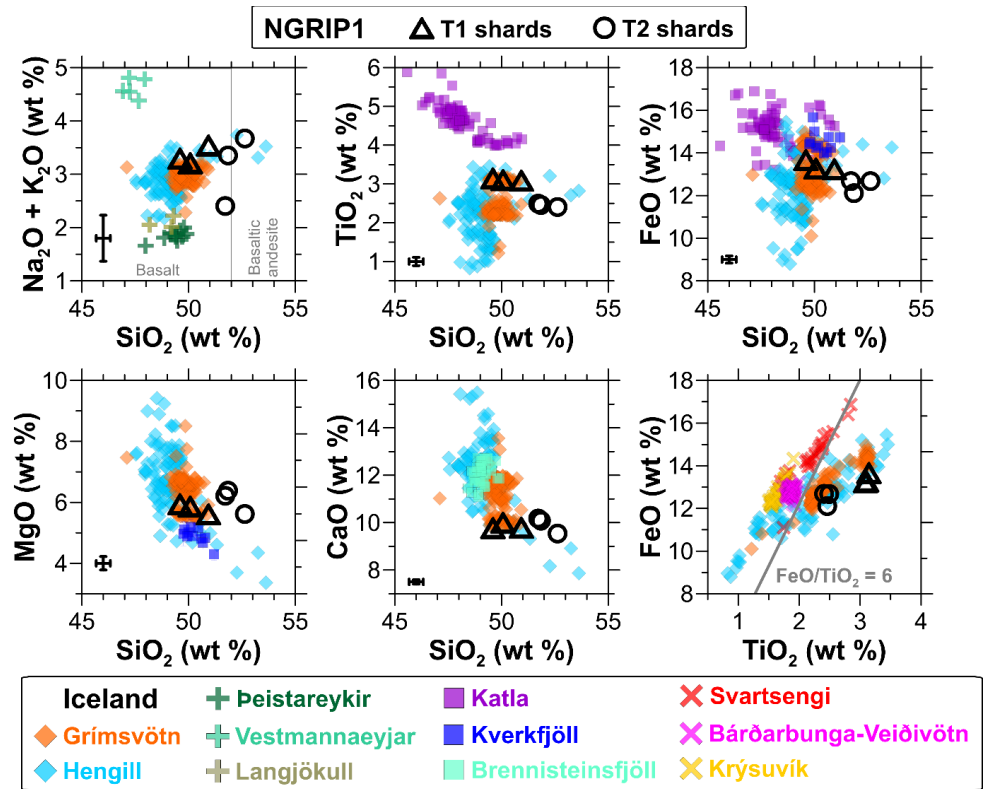
430

431 **Figure 4.** Total alkali versus silica (TAS) diagram of glass compositions from the NGRIP1 T1
432 (open triangles) and T2 (open circles) cryptotephra samples, compared with volcanic materials
433 from high-magnitude eruptions that occurred during the period 1000–1200 CE (see **Table 1** for
434 full data sources). VEI 5: Mashu (red), Asama (blue), Ubinas (yellow); VEI 4: Furnas
435 (magenta), Izu Ōshima (dark-brown), Ksudach (gold), Sunset Crater (purple), El Chichón
436 (khaki), Arenal (crimson), San Salvador (green), Cerro Bravo (pink), Cayambe (turquoise) (see
437 **Fig. 1a**). The error bar (bottom left) represents the maximum analytical uncertainty ($\pm 1\sigma$) based
438 on repeated measurements of secondary standards.



Table 1. Summary of volcanic eruptions with Volcanic Explosivity Index (VEI) ≥ 4 that occurred during the period 1000–1200 CE.

Volcano	Eruption name	Region		Eruption date (CE)	VEI	Tephra data		
						Geochemistry	Age	Reference
Furnas	E caldera floor	Azores (37.8° N, 25.3° W)	Northern Hemisphere extratropics	1170 ± 100 (Wastegård et al., 2020)	4	Trachyte–Phonolite	1170 ± 100 CE	Wastegård et al. (2020)
Asama		Japan (36.4° N, 138.5° E)		1108–08–29 (Takahashi et al., 2022)	5	Dacite–Rhyolite	Late Quaternary	Aoki (2020)
Izu Ōshima	Y6.0	Japan (34.7° N, 139.4° E)		~1183 (Koyama and Hayakawa, 1996)	4	Andesite	Not available	Ikehata et al. (2010)
Mashu	Ma-b	Japan (43.6° N, 144.5° E)		1125 ± 95 (Razzhigaeva et al., 2016; Nakamura, 2016)	5	Rhyolite	1125 ± 95 CE	Razzhigaeva et al. (2016), Nakamura (2016)
Ksudach	KSht1	Kamchatka (51.8° N, 157.5° E)		1000 ± 50 (Zelenin et al., 2020)	4	Rhyolite	~1.7 kyr	Zelenin et al. (2020)
Sunset Crater		Arizona, USA (35.4° N, 111.5° W)		1075 ± 25 (Elson et al., 2011)	4	Basalt	1075 ± 25 CE	Alfano et al. (2019)
El Chichón	Unit C	Mexico (17.4° N, 93.2° W)	Tropics	1190 ± 150 (Nooren et al., 2017)	4	Rhyolite	~540 CE / ~1050 CE	Macias et al. (2003), Nooren et al. (2017)
Arenal		Costa Rica (10.5° N, 84.7° W)		~1030 (Soto and Alvarado, 2006)	4	Dacite	1990–2001 CE	Szramek et al. (2006)
San Salvador	Boquerón	El Salvador (13.7° N, 89.3° W)		~1200 (Ferrés et al., 2013)	4	Basaltic andesite–Andesite	Not available	Fairbrothers et al. (1978)
Cerro Bravo	CB3	Colombia (5.1° N, 75.3° W)		1129 ± 154 (Lescinsky, 1990)	4	Dacite	1129 ± 154 CE	Lescinsky (1990)
Cayambe	Lava dome near eastern summit	Ecuador (0.0° N, 78.0° W)		~1040 (Samaniego et al., 2005)	4	Andesite–Dacite	<~4 kyr	Samaniego et al. (2005)
Ubinas		Peru (16.4° S, 70.9° W)		1082 ± 82 (Rivera et al., 2014)	5	(Trachy)dacite–Rhyolite	1667 CE / 2006 CE	Rivera et al. (2014), Samaniego et al. (2020)



442

443 **Figure 5.** Glass compositions from the NGRIP1 T1 (open triangles) and T2 (open circles)
444 cryptotephra samples, compared with basaltic materials from Grímsvötn (orange diamonds),
445 Hengill (light-blue diamonds), Þeistareykir (dark-green crosses), Vestmannaeyjar (light-green
446 crosses), Langjökull (khaki crosses), Katla (purple squares), Kverkfjöll (blue squares),
447 Brennisteinsfjöll (turquoise squares), Svartsengi (red X-marks), Bárðarbunga-Veiðivötn
448 (magenta X-marks), and Krýsuvík (yellow X-marks) in Iceland (see **Table 2** for full data
449 sources and **Fig. 1a**). The grey line in the FeO versus TiO₂ panel denotes FeO/TiO₂ = 6. Error
450 bars (bottom left) represent the maximum analytical uncertainty ($\pm 1\sigma$) based on repeated
451 measurements of secondary standards.



452 **Table 2.** Summary of Icelandic volcanoes active during the late Holocene.

Volcano	Region		Tephra data		
			Geochemistry	Age	Reference
Askja	65.0° N, 16.8° W	Northern Volcanic Zone	Rhyolite	1875 CE	Gudmundsdottir et al. (2016) , Wulf et al. (2016)
Krafla	65.7° N, 16.8° W		Rhyolite	Not available	Masotta et al. (2018)
Þeistareykir (Theistareykir)	65.9° N, 17.0° W		Basalt	Not available	Sigurdsson et al. (2000) , Thomson and MacLennan (2013)
Kverkfjöll	64.7° N, 16.7° W		Basalt	Holocene	Óladóttir et al. (2011)
Torfajökull	63.9° N, 19.1° W	Eastern Volcanic Zone	Rhyolite	~2.2 kyr / ~798 kyr	Lacasse and Garbe-Schonberg (2001) , Gudmundsdottir et al. (2016)
Eyjafjallajökull	63.6° N, 19.6° W		(Trachy)andesite– Trachy(dac)ite	2010 CE	Sigmarsson et al. (2011)
Bárðarbunga- Veidivötn	64.6° N, 17.5° W		Basalt	1477 CE	Caracciolo et al. (2020)
Grímsvötn	64.4° N, 17.3° W		Basalt	~8.6–8.7 kyr	Harning et al. (2018)
Vestmannaeyjar	63.4° N, 20.3° W		(Trachy)basalt–Basaltic (trachy)andesite	1964 CE / 1973 CE	Thy (1983)
Hekla	64.0° N, 19.7° W		Andesite	1693 CE	Harning et al. (2018)
Katla	63.6° N, 19.1° W		Basalt and rhyolite (bimodal)	~8.4–34.3 kyr	Thornalley et al. (2011)
Prestahnúkur	64.6° N, 20.7° W	Western Volcanic Zone	Rhyolite	~89 kyr	McGarvie et al. (2007)
Hengill	64.1° N, 21.4° W		Basalt–Basaltic andesite	Not available	Tronnes (1990)
Langjökull	64.7° N, 20.2° W		Basalt	Not available	Meyer et al. (1985)
Öræfajökull	64.0° N, 16.7° W	Öræfajökull Volcanic Belt	Rhyolite	~1362 CE	Selbekk and Tronnes (2007) , Sharma et al. (2008)
Snæfellsjökull	64.8° N, 23.8° W	Snæfellsnes Volcanic Belt	Trachydacite–Rhyolite	~1.8 kyr / ~3.6 kyr	Sigvaldason et al. (1992) , Larsen et al. (2002) , Wastegård et al. (2009)
Svartsengi	63.9° N, 22.4° W	Reykjanes Volcanic Belt	Basalt	~1210–1235 CE	Meyer et al. (1985) , Geirsdottir et al. (2022) , Caracciolo et al. (2023)
Brennisteinsfjöll	63.9° N, 21.8° W		Basalt	~800 CE / ~1000 CE	Caracciolo et al. (2023)
Krýsuvík	63.9° N, 22.1° W		Basalt	~800 CE / ~1151 CE	Caracciolo et al. (2023)

453



454 Volcanic sulfate peaks from proximal eruptions typically occur nearly simultaneously
455 with the deposition of associated tephra, as much of the sulfur reaches the ice sheet via the
456 same T/LMS pathway, whereas sulfate injected into the stratospheric overworld by distal
457 eruptions is deposited with a lag relative to the arrival of tephra (Koffman et al., 2013; Plunkett
458 et al., 2023; Abbott et al., 2024). While the absence of a high-resolution insoluble particle
459 record for NGRIP1 precludes a definitive assessment, this depositional sequence suggests that
460 the T1 shards may be temporally associated with the initial NH extratropical eruption, as they
461 were deposited close to the first sulfate peak (section B in Fig. 2). Nevertheless, the T1 shards
462 postdate the non-MIF sulfate influx phase, during which tephra from the same proximal
463 eruption would be expected to arrive via the T/LMS pathway (section A). The T2 shards were
464 deposited shortly before the second sulfate peak, within the interval attributed to the tropical
465 eruption (section C), which is broadly consistent with tephra preceding the lagged stratospheric
466 sulfate from a distal eruption, although they precede this peak by less than half a year, whereas
467 tephra from tropical eruptions typically arrives near the onset of sulfate deposition (Plunkett et
468 al., 2023). Thus, the deposition of the T2 shards is broadly compatible in timing with the
469 tropical eruption, whereas that of the T1 shards is less consistent with the initial NH
470 extratropical eruption; nevertheless, depositional timing alone cannot establish the source of
471 either glass population, and their potential volcanic sources are therefore assessed below in
472 light of their geochemistry.

473 To identify the source volcanoes of the ice-core shards, we compared their glass
474 geochemical compositions with those of volcanic products from documented eruptions in the
475 tropics (30° S–30° N) and the NH extratropics (30° N–90° N) for the period 1000–1200 CE.
476 These eruption records were compiled from the existing literature, the Large Magnitude
477 Explosive Volcanic Eruptions (LaMEVE) database (Crosweller et al., 2012), and the
478 Smithsonian Institution’s Global Volcanism Program (GVP) catalog. We compiled events with
479 $VEI \geq 4$ (Table 1 and Fig. 1a) or with currently unassigned VEI ratings (Table S3), because
480 distant volcanic eruptions must be sufficiently explosive to enable the long-range transport of
481 ash to Greenland (Plunkett and Pilcher, 2018; Plunkett et al., 2022). An exception was made
482 for Icelandic eruptions (Table 2), which were evaluated regardless of their VEI or
483 documentation status because Iceland’s proximity allows tephra from even low-magnitude
484 eruptions to reach Greenland and because past events may be under-recorded in geological



485 archives despite active Icelandic volcanism throughout the late Holocene (Thordarson and
486 Larsen, 2007; Óladóttir et al., 2011; Sæmundsson et al., 2020). The wide time span was adopted
487 to accommodate the chronological uncertainty of the events; however, eruptions with robust
488 age constraints were excluded from the comparison if their ages did not overlap with the 1108–
489 1110 CE interval. Geochemical data were available for 70 of the 77 compiled records, and these
490 70 records formed the basis of the comparison. For each record, we prioritized geochemical
491 data from tephra deposits attributed to the corresponding eruptions; where such data were
492 unavailable, we used compositions of eruptive products representative of the volcano. Among
493 these records, high-magnitude events ($VEI \geq 4$), including the VEI 5 eruptions of Mashu (1125
494 ± 95 CE), Asama (1108 CE), and Ubinas (1082 ± 82 CE), were considered the most plausible
495 candidates.

496 As shown in **Figs. 4 and S4**, the ice-core glass compositions are clearly distinct from those
497 of the high-magnitude events, as well as those of other eruptions in the extratropics (Alaska,
498 the Azores, Europe, Japan, Kamchatka, the Kurils, and North America) and the tropics.
499 Therefore, these eruptions, including the 1108 CE Asama eruption previously proposed as a
500 source (Guillet et al., 2020), can be excluded as sources of the analyzed glass shards. Instead,
501 the ice-core glass geochemistry shows the greatest affinity with basaltic materials from
502 Icelandic volcanic systems (**Table 2** and **Fig. 5**). Among these basaltic Icelandic candidates,
503 volcanoes characterized by relatively low (Þeistareykir and Langjökull) or high
504 (Vestmannaeyjar) alkali contents, high TiO_2 and FeO (Katla), high FeO and low MgO
505 (Kverkfjöll), high CaO (Brennisteinsfjöll), or FeO/ TiO_2 ratios of >6 (Bárðarbunga–Veidivötn,
506 Svartsengi, and Krýsuvík; Óladóttir et al., 2011) can also be ruled out (**Table 2** and **Fig. 5**).

507 Of the remaining candidates, the T1 shards show geochemical affinities with Grímsvötn
508 and Hengill compositions. The T2 shards are most similar to Hengill products, although they
509 have higher MgO and CaO contents than Hengill products at comparable SiO_2 contents (**Fig.**
510 **5**). These affinities raise the possibility that the T1 shards originated from Grímsvötn and the
511 T2 shards from Hengill, or that both populations originated from Hengill. However, previous
512 studies of Icelandic volcanic activity during the Holocene reported that no Grímsvötn eruptions
513 occurred between the 1104 and 1158 CE Hekla eruptions (Óladóttir et al., 2011) and that the
514 most recent known eruption of Hengill occurred approximately 2000 years ago (Sæmundsson
515 et al., 2020). Therefore, if the shards originated from Grímsvötn and/or Hengill, they may



516 represent under-recorded eruptions or material remobilized from older deposits, although
517 remobilization of Hengill material is unlikely because the Na₂O contents of the analyzed shards
518 show no evidence of the alteration expected after such prolonged exposure. By contrast, two
519 Grímsvötn eruptions are recorded shortly before the 1104 CE Hekla eruption (Óladóttir et al.,
520 2011), indicating the possibility that the T1 shards result from remobilization of these older
521 deposits.

522 Furthermore, regardless of whether the shards derived directly from under-recorded
523 eruptions or from remobilized older deposits, these potential Icelandic sources are unlikely to
524 account for the stratospheric sulfate signal during the 1108–1110 CE episode (see Sect. 3.1). If
525 the shards derived from contemporaneous eruptions, the absence of corresponding tephra
526 layers in Icelandic records may suggest that these eruptions were relatively weakly explosive
527 and contributed little sulfur to the stratospheric overworld. Even the much larger 1783 CE Laki
528 eruption (VEI 4), which released approximately 60 Tg S into the atmosphere, did not produce
529 the characteristic MIF signal in Greenland ice cores (Lanciki et al., 2012). This implies that the
530 cryptotephra and the stratospheric sulfate likely derive from different volcanic sources.
531 Accordingly, while no direct geochemical link between the ice-core shards and the high-
532 magnitude eruptions was found, these eruptions cannot be ruled out as potential sources of the
533 stratospheric sulfate. The absence of glass shards from these eruptions may reflect limited
534 tephra transport from the source volcanoes to Greenland due to atmospheric circulation
535 (Lacasse, 2001; Monteath et al., 2025; Shin et al., 2025). Further investigation is required to
536 identify the source volcanoes of the stratospheric sulfur-injecting eruptions.

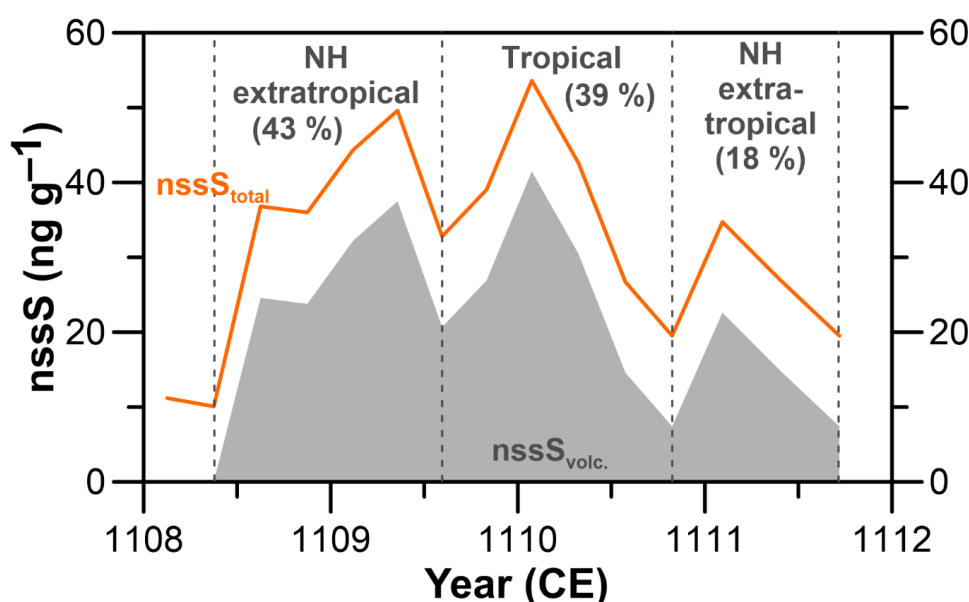
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538 3.3. Revised volcanic stratospheric sulfur injection for the 1108–1110 CE volcanic episode

539 The VSSI for the 1108–1110 CE volcanic episode was previously estimated at 19.2 ± 4.7
540 Tg S (1σ), based on the assumption that the episode originated from a single tropical eruption
541 and that all volcanic sulfate deposited on polar ice sheets was transported exclusively through
542 the stratospheric overworld (Toohey and Sigl, 2017). Here we revise the VSSI to account for
543 the multi-eruption character of the episode and the stratospheric sulfate fraction of each event
544 (see Sect. 2.4). Based on the previously reported monthly resolved NGRIP1 record (Sigl et al.,
545 2015; Sinnl et al., 2022), the initial NH extratropical eruption, the subsequent tropical eruption,
546 and the minor NH extratropical eruption in early 1111 CE account for 43 %, 39 %, and 18 %



547 of the total Greenland sulfate deposition, respectively (**Fig. 6**). Applying these depositional
548 fractions together with the stratospheric sulfate fractions adopted for the three eruptions (37 %,
549 100 %, and 0 %, respectively; see **Sect. 3.1**) yields a total VSSI of 9.0 ± 1.4 Tg S, comprising
550 1.1 ± 0.5 Tg S for the initial NH extratropical eruption and 8.0 ± 0.9 Tg S for the tropical
551 eruption (values do not sum exactly because of rounding).



552

553 **Figure 6.** Sub-annually resolved total (orange line) and volcanic (grey shading) non-sea-salt
554 (nss) sulfur concentrations in the NGRIP1 ice core between 1108 and 1112 CE ([Sigl et al., 2015](#);
555 [Sinnl et al., 2022](#)). Volcanic concentrations were obtained by subtracting the non-volcanic
556 background (see **Sect. 2.4**). Vertical dashed lines bound the depositional intervals attributed to
557 the initial NH extratropical eruption (first peak, early 1109 CE), the subsequent tropical
558 eruption (second peak, early 1110 CE), and the minor NH extratropical eruption (third peak,
559 early 1111 CE). Percentages indicate the fraction of the integrated volcanic sulfur in NGRIP1
560 attributed to each eruption, which was used to apportion the mean volcanic sulfate deposition
561 on the Greenland Ice Sheet among the three eruptions (14.9, 13.6, and 6.1 kg km⁻²,
562 respectively). All volcanic sulfate deposition on the Antarctic Ice Sheet (10.3 kg km⁻²) was
563 attributed to the tropical eruption (see **Sect. 2.4**).



564 The bipolar ice-core arrays used for the VSSI calculation in this study (see **Sect. 2.4**) differ
565 slightly from those of [Toohey and Sigl \(2017\)](#), who incorporated three Greenland records
566 (NGRIP1, NEEM-2011-S1, and GISP2) and 15 Antarctic records (WDC06A, SP1, SP4,
567 DML05, DML07, B40, NUS08-5, NUS07-2, NUS07-5, NUS07-7, DF01, DFS10, DSS,
568 EDC96, and DA2005). Specifically, we incorporate new records (TUNU2013, TUNU2022A,
569 TUNU2022B, SPC14, and Taylor Dome), exclude DA2005, which is not publicly available,
570 and NUS07-2 and DSS, which show no clear 1108–1110 CE volcanic sulfate peaks (**Fig. S2**),
571 likely due to site-specific chemical signal variability ([Wolff et al., 2005](#); [Gautier et al., 2016](#)),
572 and apply the most recent ice-core chronologies. Nevertheless, when our methodology is
573 applied to the 1108–1110 CE episode under the same assumptions as [Toohey and Sigl \(2017\)](#)
574 (a single tropical eruption and stratospheric overworld transport only; see above), we obtain a
575 VSSI of 15.0 ± 2.4 Tg S, which overlaps within uncertainty with their previous estimate (19.2
576 ± 4.7 Tg S). This suggests that the substantial reduction in our final estimate of 9.0 ± 1.4 Tg S
577 is driven mainly by the refined multi-eruption parameters rather than by methodological or
578 data-source differences. Our findings therefore suggest that the stratospheric sulfur loading
579 associated with this episode has been overestimated in the current forcing reconstructions,
580 highlighting the value of sulfur isotope analysis for accurately characterizing the eruptive
581 nature of past volcanic events.

582

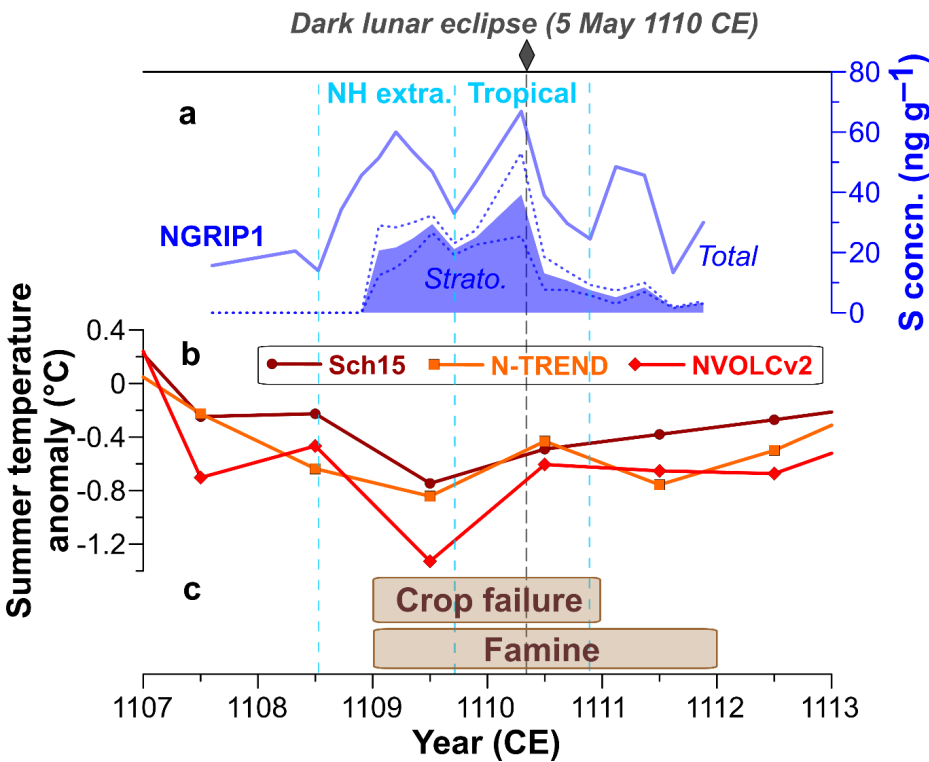
583 **3.4. Timing of volcanic forcing for the 1108–1110 CE volcanic episode**

584 The 1108–1110 CE volcanic episode has been linked to severe NH summer cooling in
585 1109 CE, societal disruption in Western Europe, and the dark total lunar eclipse observed in
586 May 1110 CE (**Figs. 1b and 7**; [Guillet et al., 2020](#)). Our high-resolution sulfur isotope data
587 from the NGRIP1 samples provide timing constraints on the volcanic forcing associated with
588 this episode. Note that caution is required when interpreting this timing, given the 3-year
589 uncertainty of the GICC21 chronology during this period ([Sinnl et al., 2022](#)). Nevertheless, the
590 robust synchronization of the 1108–1110 CE volcanic signal across well-constrained bipolar
591 chronologies ([Sigl et al., 2015](#); [McConnell et al., 2018](#); [Sinnl et al., 2022](#)) and independent
592 historical evidence for this eclipse ([Guillet et al., 2020, 2023](#)) support the accuracy of the
593 chronology over this interval.



594 As shown in **Fig. 7a**, sulfate deposition from the initial NH extratropical eruption spans
595 from mid-1108 to late 1109 CE. This signal is characterized by sulfate transport via the T/LMS
596 during the second half of 1108 CE, followed by a transition to a stratospheric sulfate signature.
597 In contrast, stratospheric sulfate deposition from the subsequent tropical eruption occurs
598 between late 1109 and late 1110 CE, peaking in early 1110 CE. Notably, the most pronounced
599 NH summer cooling occurred in 1109 CE (**Fig. 7b**). Superposed epoch analyses of tree-ring
600 temperature reconstructions indicate that maximum cooling following NH extratropical
601 eruptions tends to occur in the year of volcanic sulfate deposition, whereas tropical eruptions
602 generally produce a more delayed cooling response (Sigl et al., 2015). This timing suggests
603 that the initial NH extratropical eruption in 1108 CE likely played a major role in driving the
604 cooling.

605 This finding is consistent with recent studies indicating that extratropical eruptions can
606 generate disproportionately large NH cooling anomalies for a given stratospheric sulfur loading
607 (Toohey et al., 2019; Burke et al., 2023). Several of the coldest NH summers reconstructed for
608 the Common Era have also been associated with extratropical eruptions (Burke et al., 2023).
609 Although stratospheric aerosol simulations indicate stronger NH radiative forcing from
610 extratropical eruptions, differences in radiative forcing alone do not fully explain the enhanced
611 cooling sensitivity inferred from tree-ring records and revised stratospheric sulfur loadings
612 (Sigl et al., 2015; Toohey et al., 2019; Burke et al., 2023). This disproportionate cooling
613 response could be explained by feedbacks in the climate system that are triggered at high
614 latitudes (such as sea-ice feedbacks) or a hitherto unappreciated role of T/LMS aerosols in
615 climate forcing (e.g. through aerosol-cloud interactions; Malavelle et al., 2017; Chen et al.,
616 2022). Aerosols injected into the T/LMS are currently not explicitly represented in most
617 volcanic forcing reconstructions, which generally attribute volcanic sulfate deposition to
618 stratospheric loading (e.g., Gao et al., 2007; Toohey and Sigl, 2017). Further research is needed
619 to quantify the climatic effects of these lower-altitude aerosols.



620

621 **Figure 7.** Comparison of volcanic sulfur deposition, NH summer temperature anomalies, and
622 documented societal impacts between 1107 and 1113 CE. (a) Total (solid line) and stratospheric
623 (shading) sulfur concentrations from the NGRIP1 ice core, as in **Fig. 2d**; dotted lines represent
624 $\pm 1\sigma$ of the stratospheric sulfate fraction estimated from Monte Carlo simulations. (b) NH
625 summer temperature anomalies (relative to the 1961–1990 mean) reconstructed from tree rings
626 (Sch15: dark-red line (Schneider et al., 2015); N-TREND: orange line (Wilson et al., 2016);
627 NVOLCv2: red line (Guillet et al., 2020)). (c) Periods of documented crop failure (1109–1110
628 CE) and famine (1109–1111 CE) in Western Europe compiled from contemporary narrative
629 sources (Guillet et al., 2020). Cyan dashed lines bound the sulfate deposition intervals
630 attributed to the initial NH extratropical eruption and the subsequent tropical eruption (see **Fig.**
631 **6** and **Sect. 3.1**), and the vertical dark-grey dashed line denotes the 5 May 1110 CE dark total
632 lunar eclipse (Guillet et al., 2020).

633

634 4. CONCLUSION

635 This study provides a reassessment of the 1108–1110 CE volcanic episode, a period
636 associated with pronounced climatic anomalies and societal disruption, by integrating new
637 high-resolution sulfur isotope records and cryptotephra geochemistry obtained from Greenland



ice cores. Our sulfur isotope records show that this episode comprises an NH extratropical eruption beginning in mid-1108 CE, a subsequent tropical eruption, and a minor NH extratropical eruption in early 1111 CE, with the first two events injecting sulfur into the stratosphere. Cryptotephra analysis identifies basaltic glass shards geochemically consistent with Icelandic sources; however, this geochemical evidence does not provide definitive provenance constraints on the two stratospheric sulfur-injecting eruptions. Accordingly, high-magnitude candidate eruptions cannot be ruled out as potential sources of the stratospheric sulfate, and further work is required to identify the source volcanoes. Based on the new isotopic constraints, we revise the existing VSSI for the 1108–1110 CE episode to 9.0 ± 1.4 Tg S, comprising 1.1 ± 0.5 Tg S for the initial NH extratropical eruption and 8.0 ± 0.9 Tg S for the tropical eruption. This total is approximately 53 % lower than the previous estimate of 19.2 ± 4.7 Tg S, indicating that the stratospheric sulfur loading associated with this episode has been overestimated in current volcanic forcing reconstructions. Our timing constraints further show that the most pronounced NH summer cooling in 1109 CE was likely driven by the initial NH extratropical eruption. We suggest that sulfate aerosols residing in the T/LMS may play a major role in driving NH summer cooling, but their contribution to the climate response through aerosol-cloud interactions is typically excluded from volcanic forcing estimates. These results provide a more accurate representation of volcanic forcing over this time period for use in climate model simulations.

657



658 **DATA AVAILABILITY**

659 The data that support the findings of this study are provided in the Supplement.

660

661 **AUTHOR CONTRIBUTIONS**

662 WH and AB designed the project. AS and MS provided the NGRIP1 ice-core samples. MS,
663 JRM, NJC, and AB provided the TUNU2013 ice-core samples. FB, PS, and AB measured the
664 sulfur isotopes. SR, FB, and WH performed the glass shard analysis. SR and IG calculated the
665 revised VSSI. SR did the data analysis and wrote the original manuscript, with contributions
666 from WH, IG, MS, JRM, NJC, and AB to manuscript revision. All authors have read and
667 approved the paper.

668

669 **COMPETING INTERESTS**

670 At least one of the (co-)authors serves as editor for the special issue to which this paper belongs.

671

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