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Timing and climate forcing of volcanic eruptions for the past 2,500 years

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1 **Timing and climate forcing of volcanic eruptions for the past 2,500 years**

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29 **Volcanic eruptions contribute to climate variability, but quantifying these contributions has been**
30 **limited by inconsistencies in the timing of atmospheric volcanic aerosol loading determined from**
31 **ice cores and subsequent cooling from climate proxies such as tree rings. Using new records of**
32 **atmospheric aerosol loading developed from high-resolution, multi-parameter measurements**
33 **from an array of Greenland and Antarctic ice cores as well as distinctive age markers to constrain**
34 **chronologies, we resolve these inconsistencies. Here we show that large eruptions in the tropics**
35 **and high latitudes were primary drivers of interannual-to-decadal temperature variability in the**
36 **Northern Hemisphere during the past 2,500 years. Overall, cooling was proportional to the**
37 **magnitude of volcanic forcing and persisted for up to ten years after some of the largest eruptive**
38 **episodes. Our revised timescale now firmly implicates volcanic eruptions as catalysts in the major**
39 **6th century pandemics, famines, and socioeconomic disruptions in Eurasia and Mesoamerica while**
40 **allowing multi-millennium quantification of climate response to volcanic forcing.**

41 Volcanic eruptions are primary drivers of natural climate variability – as sulfate aerosol injections
42 into the stratosphere shield the Earth’s surface from incoming solar radiation, leading to short-term
43 cooling at regional-to-global scales¹. Temperatures during the past 2,000 years have been
44 reconstructed at regional², continental³, and global scales⁴ using proxy information from natural
45 archives. Tree-ring-based proxies provide the vast majority of climate records from mid- to high-
46 latitude regions of (predominantly) the Northern Hemisphere (NH), whereas ice-core records (e.g.,
47 $\delta^{18}\text{O}$) represent both polar regions³. Climate forcing reconstructions for the Common Era (CE) –
48 including solar (e.g., ^{10}Be)⁵ and volcanic (e.g., sulfate)^{6,7} activity – mostly derive from ice-core
49 proxies. Any attempt to attribute reconstructed climate variability to external volcanic forcing – and
50 distinguish between response, feedback, and internal variability of the climate system – requires ice-
51 core chronologies that are synchronous with those of other climate reconstructions. In addition,
52 multi-proxy climate reconstructions²⁻⁴ derived from ice cores and other proxies such as tree rings will
53 have diminished high-to-mid-frequency amplitudes if the individual climate records are on different
54 timescales. Magnitudes and relative timing of simulated NH temperature response to large volcanic

55 eruptions are in disagreement with reconstructed temperatures obtained from tree rings^{8,9}, but it is
56 unclear to what extent this model/data mismatch is caused by limitations in temperature
57 reconstructions, volcanic reconstructions, or implementation of aerosol forcing in climate models⁹⁻¹¹.
58 The hypothesis of chronological errors in tree-ring-based temperature reconstructions^{8,9} offered to
59 explain this model/data mismatch has been tested and widely rejected¹¹⁻¹⁴, while new ice-core
60 records have become available providing different eruption ages^{15,16} and more precise estimates of
61 atmospheric aerosol mass loading¹⁷ than previous volcanic reconstructions. Using documented¹⁸ and
62 previous ice-core-based eruption ages¹⁶, strong summer cooling following large volcanic eruptions
63 has been recorded in tree-ring-based temperature reconstructions during the 2nd millennium CE with
64 a one-to-two year lag similar to that observed in instrumental records after the 1991 Pinatubo
65 eruption¹⁹. An apparent seven-year delayed cooling observed in individual tree-ring series relative to
66 Greenland ice-core acidity peaks during the 1st millennium CE, however, suggests a bias in existing
67 ice-core chronologies^{20,21}. Using published ice-core chronologies, we also observed a seven-year
68 offset between sulfate deposition in North Greenland and the start of tree-ring growth reduction in
69 a composite of five multi-centennial tree-ring summer temperature reconstructions (“N-Tree”) from
70 the NH between 1 and 1000 CE (Methods), whereas tree-ring response was effectively immediate
71 for eruptions occurring after 1250 CE (Fig. 1a).

72 **Precise time marker across hemispheres.** Independent age markers to test the accuracy of tree-ring
73 and ice-core chronologies recently have become available with the detection of abrupt enrichment
74 events in the ¹⁴C content of tree rings. Rapid increases of atmospheric ¹⁴C were first identified in
75 individual growth increments of cedars from Japan between 774 and 775 CE²² and between 993 and
76 994 CE²³. The presence and timing of the event in 775 CE (henceforth, 775 event) has been
77 reproduced by all radiocarbon measurements performed on tree rings at annual (or higher)
78 resolution – including tree cores from Germany²⁴, the Alps¹², the Great Basin²⁵ (USA), and Siberia²⁵.
79 Recent identification of the same 775 CE event in kauri wood samples from New Zealand in the
80 Southern Hemisphere (SH) demonstrates the global extent of the rapid ¹⁴C increase and provides

81 further constraints on the event's exact timing (March 775 ± 6 months) due to the asynchronous SH
82 growing season²⁶. While the cause of the 775 and 994 events is still debated^{22,24,27}, we expect that
83 they might also produce an excess of cosmogenic ^{10}Be through the interaction of high-energy
84 particles with atmospheric constituents^{28,29}. Since both of these radionuclides are incorporated
85 rapidly into proxy archives via aerosol deposition (^{10}Be in ice cores) and photosynthesis ($^{14}\text{CO}_2$ in tree
86 rings), isotopic anomalies caused by these extraterrestrial events provide a global age marker to
87 directly link ice-core records to tree-ring chronologies²⁷. The latter serve as an absolute and precise
88 age marker, verified (at least since 775 CE) by the coherence of the rapid increase in ^{14}C in all tree-
89 rings records for which high-resolution radiocarbon analyses were performed, including those
90 speculated to be at risk of missing rings⁸. We measured ^{10}Be concentrations at approximately annual
91 resolution in four ice cores – NEEM-2011-S1, TUNU2013, and NGRIP in Greenland, and the West
92 Antarctic Ice Sheet Divide Core (WDC) – over depth ranges encompassing the year 775 as dated in
93 existing ice-core chronologies in order to provide a direct, physically-based test of any dating bias in
94 these chronologies (Fig. 1, Extended Data Fig. 1, Methods, Supplementary Data S1). Both polar ice
95 sheets contain ^{10}Be concentrations exceeding the natural background concentration ($>150\%$; 6σ) for
96 one-to-two consecutive years, compatible with the 775 CE event observed in tree rings. Using the
97 original ice-core age models^{16,30}, the ages of the ^{10}Be maxima in NEEM-2011-S1, NGRIP, and WDC are
98 768 CE, offset by 7 years from the tree-ring event. A further ^{10}Be anomaly measured in NEEM-2011-
99 S1 at 987 CE, compatible with the 994 CE event in tree rings, suggests a chronological offset was
100 present by the end of the first millennium CE (Fig. 1). Several different causes may have contributed
101 to the offset (see a summary in the Methods section), among which is the use of a previous dating
102 constraint³⁰ for Greenland where volcanic fallout in the ice was believed to indicate the historic (79
103 CE) eruption of Vesuvius.

104 **Revised ice-core chronologies.** Given the detection of a bias in existing ice-core chronologies, we
105 developed new timescales prior to the 1257 Samalas eruption³¹ using highly resolved, multi-
106 parameter aerosol concentration records from three ice cores: NEEM-2011-S1, NEEM, and WDC. We

107 used the StratiCounter program, an automated, objective, annual-layer detection method based on
108 Hidden Markov Model (HMM) algorithms³² (Methods). For NEEM-2011-S1, obtained confidence
109 intervals for the layer counts allowed us to improve the dating further by constraining the timescale
110 using the 775 CE ¹⁰Be anomaly and three precisely dated observations of post-volcanic aerosol
111 loading of the atmosphere (Fig. 2, Extended Data Tables 1-3, Methods, Supplementary Data S2). We
112 evaluated the accuracy of our new chronologies (“WD2014” for WDC and “NS1-2011” for NEEM) by
113 comparison to (1) an extensive database of historical volcanic dust veil observations (Extended Data
114 Fig. 2, Methods, Supplementary Data S2), (2) ice-core tephra evidence (Methods), and (3) the 994 CE
115 event (Methods, Fig. 2). Using the new timescales, we found large sulfate signals in Greenland (e.g.
116 682, 574, 540 CE) between 500 and 2000 CE that frequently occurred within ± 1 year from
117 comparable – and independently dated – signals in Antarctica. These bipolar signals can now be
118 confidently attributed to large tropical eruptions (Fig. 2). Back to 400 BCE, other large sulfate peaks
119 (e.g., 44 BCE) were synchronous to within ± 3 years (Fig. 2). We conclude that the revised ice-core
120 timescales are accurate to within less than ± 5 years during the past 2,500 years based on combined
121 evidence from radionuclides, tree rings, tephra analyses, and historical accounts. Compared to the
122 previous chronologies, age models differ prior to 1250 CE by up to 11 years (GICC05, Greenland) and
123 14 years (WDC06A-7, Antarctica) (Extended Data Fig. 3).

124 **History of volcanic forcing.** Employing our revised timescales and new high-resolution, ice-core
125 sulfur measurements, we developed an extended reconstruction of volcanic aerosol deposition since
126 early Roman times for both polar ice sheets from which we then estimated radiative forcing using
127 established transfer functions¹⁵ (Fig. 3, Methods, Supplementary Data S3-S5). This forcing series is
128 characterized by improved dating accuracy, annual resolution, and a larger number of ice-core
129 records in the Antarctic ice-core sulfate composite¹⁷ than previous reconstructions^{6,7}. It spans 2,500
130 years, allowing investigation of climate-volcano linkages more accurately and earlier than with
131 previous reconstructions. It also provides a perspective on volcanic influences during major historical
132 epochs, such as the growth of Roman imperial power and subsequent decline of the migration

133 period in Europe – times of (1) demographic and economic expansion as well as relative societal
134 stability and (2) political turmoil and population instability, respectively³³. With improved dating and
135 lower volcanic-sulfate detection limits from the Antarctic array¹⁷, we were able to detect, estimate,
136 and attribute volcanic aerosol loading and forcing from 283 individual eruptive events during this
137 period (Fig. 3). We attributed about 50% to mid-to-high latitudes NH sources, while 81 were
138 attributed to tropical eruptions (having synchronous sulfate deposition on both polar ice sheets).
139 These tropical volcanic eruptions contributed 64% of total volcanic forcing throughout the period,
140 with five events exceeding the sulfate loading from Tambora, 1815 (Fig. 3, Extended Data Table 4).
141 Events in 426 BCE and 44 BCE rival the great 1257 CE Samalas eruption (Indonesia) as the largest
142 sulfate producing eruptions during this time. These two earlier events have not been widely
143 regarded as large tropical eruptions with potential for strong climate impact²⁰, due to the lack of
144 complete and synchronized sulfate records from Greenland and Antarctica. We base the claim that
145 these two eruptions were tropical in origin and caused significant radiative perturbations on the
146 observation that ice cores from Greenland and Antarctica record coeval (within their respective age
147 uncertainties) and exceptionally high volcanic sulfate concentrations. Both these events were
148 followed by strong and persistent growth reduction in tree-ring records³⁴ (Fig. 2) as typically
149 observed after large tropical eruptions during the Common Era (Fig. 3).

150 **Post-volcanic summer cooling.** Superposed epoch analyses (Methods) performed on the “N-Tree”
151 composite record centered on the largest volcanic signals between 500 BCE and 1250 CE as well as
152 between 1250 and 2000 CE, show a clear post-volcanic cooling signal. For both periods, maximum
153 tree-ring response lagged the date of initial increase of sulfate deposition by one year (Fig. 2),
154 consistent with the response observed if using only historically documented eruptions with secure
155 dating for the past 800 years¹⁸. The sharp and immediate (i.e., ≤ 1 year lag) response of tree growth
156 to the ice-core volcanic signal throughout the past 2,500 years further corroborates the accuracy of
157 our new ice-core timescales (Extended Data Fig. 4). Of the 16 most negative tree-growth anomalies
158 (i.e., coldest summers) between 500 BCE and 1000 CE, 15 followed large volcanic signals – with the

159 four coldest (43 BCE, 536, 543, and 627 CE) occurring shortly after several of the largest events
160 (Extended Data Tables 4, 5). Similarly, the coldest summers in Europe during the Common Era³ were
161 associated with large volcanic eruptions (Extended Data Table 5). Reduced tree growth after volcanic
162 eruptions also was prominent in decadal averages of the “N-Tree” composite. All 16 decades with
163 the most reduced tree growth for our 2,500-year period followed large eruptions (Fig. 3, Extended
164 Data Table 5). In many cases, such as the coldest decade from 536 to 545 CE³, sustained cooling was
165 associated with the combined effect of several successive volcanic eruptions.

166 Strong post-volcanic cooling was not restricted to tropical eruptions; it also followed NH
167 eruptions (Fig. 4), with maximum cooling in the year of volcanic-sulfate deposition. In contrast to the
168 average of the 19 largest CE tropical eruptions, however, the NH-only eruptions did not give rise to
169 any significant long-term cooling effect (Fig. 4). Persistence of implied post-volcanic cooling following
170 the largest tropical eruptions is strongly expressed in temperature reconstructions based on tree-
171 ring width measurements (e.g., those from the Alps), with recovery times of more than 10 years.
172 Persistent cooling, with temperature reduction significantly below the pre-eruption baseline for six
173 consecutive years, also is observed in temperature reconstructions based on maximum latewood
174 density (e.g., those from Northern Scandinavia), the preferred proxy to quantify volcanic cooling
175 contributions on climate due to less marked biological memory effects³⁵ (Fig. 4). These findings
176 indicate that eruption-induced climate anomalies following large tropical eruptions may last longer
177 than is indicated in many climate simulations (<3–5 years)^{9,36,37} and that potential positive feedbacks
178 initiated after large tropical eruptions (e.g., sea-ice feedbacks) may not be adequately represented in
179 climate simulations^{38,39}. The five-year averaged (lag 0 to lag 4 years) cooling response over three NH
180 regions (Methods) following the 19 largest CE tropical eruptions was -0.6 ± 0.2 °C (2 standard error
181 of the mean (SEM)), that of large NH eruptions was -0.4 ± 0.4 °C with strongest cooling induced in
182 the high latitudes. Overall, cooling was proportional to the magnitude of volcanic forcing, with
183 stratospheric sulfate loading exceeding that of Tambora inducing the strongest response of $-1.1 \pm$
184 0.6 °C (Figs. 3, 4).

185 **Global climate anomalies in 536-550 CE.** Our new dating allowed us to clarify long-standing debates
186 concerning the origin and consequences of the severe and apparently global climate anomalies
187 observed from c.536–550 CE, which began with recognition of the “mystery cloud” of 536 CE⁴⁰
188 observed in the Mediterranean basin. Under previous ice-core dating, it has been argued that this
189 dust veil corresponded to an unknown tropical eruption dated 533–534 CE (± 2)⁴¹. Using our revised
190 timescales, we found at least two large volcanic eruptions around this period (Fig. 5). A first –
191 apparently NH, eruptive episode in 535 or early 536 CE – injected large amounts of sulfate and ash
192 into the atmosphere. Geochemistry of tephra filtered from the NEEM-2011-S1 ice core at a depth
193 corresponding to 536 CE indicated multiple North American volcanoes as likely candidates for a
194 combined volcanic signal (Extended Data Fig. 5, Methods, Supplementary Data S5). Historical
195 observations (Extended Data Table 3) identified atmospheric dimming as early as March 24, 536,
196 lasting up to 18 months. The summer of 536 CE appeared exceptionally cold in all tree-ring
197 reconstructions in the extra-tropical NH from N-America³⁴, over Europe^{35,42,43} to Asia⁴⁴. Depending
198 upon reconstruction method, European summer temperatures in 536 CE dropped 1.6-2.5°C relative
199 to the previous 30-year average³. A second eruptive episode in 539 or 540 CE, identified in both
200 Greenland and Antarctica ice-core records and hence likely tropical in origin, resulted in up to 10%
201 higher global aerosol loading than the Tambora 1815 eruption reconstructed from our bipolar
202 sulfate records. Summer temperatures consequently dropped again, by 1.4-2.7°C in Europe in 541
203 CE³, and cold temperatures persisted in the NH until almost 550 CE^{3,33,34,42} (Figs. 2, 3, 5). This
204 provides a notable environmental context to widespread famine and the great Justinian Plague 541-
205 543 CE that was responsible for decimating populations in the Mediterranean and potentially
206 China^{45,46}. While certain climatic conditions (e.g., wet summers) have been linked to plague
207 outbreaks in the past⁴⁷, a direct causal connection of these two large volcanic episodes and
208 subsequent cooling to crop failures and outbreaks of famines and plagues is difficult to prove³³.
209 However, the exact delineation of two of the largest volcanic signals – with exceptionally strong and
210 prolonged NH cooling; written evidence of famines and pandemics; as well as socio-economic

211 decline observed in Mesoamerica (“Maya Hiatus”⁴⁸), Europe, and Asia – supports the idea that the
212 latter may be causally associated with volcanically-induced climatic extremes.

213 Detailed study of major volcanic events during the 6th century (Fig. 5) and an assessment of post-
214 volcanic cooling throughout the past 2,500 years using stacked tree-ring records and regional
215 temperature reconstructions (Fig. 4, Extended Data Fig. 4) demonstrated that large eruptions in the
216 tropics and high latitudes were primary drivers of interannual-to-decadal NH temperature variability.
217 The new ice-core chronologies imply that previous multi-proxy reconstructions of temperature that
218 include ice-core records²⁻⁴ have diminished high-to-mid-frequency amplitudes and must be updated
219 to accurately capture the timing and full amplitude of paleoclimatic variability. By creating a volcanic
220 forcing index independent of but consistent with tree-ring-indicated cooling, we provide an essential
221 step to advance understanding of external forcing on natural climate variability during the past 2,500
222 years. With the expected detection of additional rapid $\Delta^{14}\text{C}$ enrichment events from ongoing efforts
223 in annual resolution ¹⁴C tree-ring analyses⁴⁹, there will be future opportunities to further constrain
224 ice-core dating throughout the Holocene and develop a framework of precisely dated, globally
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333

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379 **Figure 1 | Annual ^{10}Be ice-core records and post-volcanic cooling from tree rings under existing ice-**
380 **core chronologies. a)** Superposed epoch analysis for the seven largest volcanic signals in NEEM-
381 2011-S1 between 78 and 1000 CE and for the 10 largest eruptions between 1250 and 2000 CE,
382 respectively¹⁶. Shown are growth anomalies from a multi-centennial, temperature-sensitive tree-ring
383 composite (N-Tree^{42,43,76-78}, Methods) 10 years after the year of volcanic sulfate deposition at NEEM
384 ice core site in Greenland (GICC05 timescale), relative to the level five years prior to sulfate
385 deposition; **b)** annually resolved ^{10}Be concentration records from WDC, TUNU2013, NGRIP, and

386 NEEM-2011-S1 ice cores on their original timescales and annually resolved $\Delta^{14}\text{C}$ series from tree-ring
387 records between 755-795 CE^{22,24}, with arrows representing the suggested time shifts for
388 synchronization; error bars are 1σ measurement uncertainties; estimated relative age uncertainty
389 for TUNU2013 at this depth interval from volcanic synchronization with NEEM-2011-S1 is ± 1 year; **c)**
390 annually resolved ^{10}Be concentration record from NEEM-2011-S1 ice core on its original timescale
391 and annually resolved $\Delta^{14}\text{C}$ series from tree rings between 980-1010 CE²³; error bars are 1σ
392 measurement uncertainties.

393 **Figure 2 | Re-dated ice-core, non-sea-salt sulfur records from Greenland and Antarctica in relation**
394 **to growth anomalies in the N-Tree composite. a)** Upper panel: Ice-core, non-sea-salt sulfur (nssS)
395 records from Greenland (NEEM, NEEM-2011-S1) on the NS1-2011 timescale between 500 BCE and
396 1300 CE, with identified layer of Tianchi tephra⁶⁷ highlighted (orange star). Calendar years are given
397 for the start of volcanic sulfate deposition. Events used as fixed age markers to constrain the dating
398 (i.e., 536, 626, 775, 939, and 1258) are indicated (purple stars). Annually resolved ^{10}Be concentration
399 record (green) from NEEM-2011-S1 encompassing the two $\Delta^{14}\text{C}$ excursion events in trees from 775
400 and 994; middle panel: tree-ring growth anomalies (relative to 1000-1099 CE) for the N-Tree
401 composite^{42,43,76-78}; lower panel: nssS records from Antarctica (WDC, B40) on the WD2014 timescale
402 and annually resolved ^{10}Be concentrations from WDC; **b)** superposed epoch analysis for 28 large
403 volcanic signals during the past 2,500 years. Tree-ring growth anomalies relative to the timing of
404 reconstructed sulfate deposition in Greenland (NS1-2011) are shown for 1250 to 2000 CE and 500
405 BCE to 1250 CE.

406 **Figure 3 | Global volcanic aerosol forcing and NH temperature variations for the past 2,500 years.**
407 **a)** 2,500 year record of tree-growth anomalies (N-Tree^{42,43,76-78}; relative to 1000-1099 CE) and
408 reconstructed summer temperature anomalies for Europe and Arctic³; 40 coldest single years and
409 12 coldest decades based on N-Tree are indicated; **b)** reconstructed global volcanic aerosol forcing
410 from bipolar sulfate composite records from tropical (bipolar), NH, and SH eruptions. Total (i.e., time

411 integrated) forcing values are calculated by summing the annual values for the duration of volcanic
412 sulfur deposition. 40 largest volcanic signals are indicated, and ages are given for events
413 representing atmospheric sulfate loading exceeding Tambora 1815.

414 **Figure 4 | Post-volcanic cooling.** Superposed composites (time segments from selected periods in
415 the Common Era positioned so that the years with peak negative forcing are aligned) of the JJA
416 temperature response to: **a)-c)** 24 largest eruptions (>Pinatubo 1991) for three regional
417 reconstructions in Europe^{3,35,42}; **d)-f)** 19 largest tropical eruptions; **g)** five largest NH eruptions; **h)**
418 eruptions with negative forcing larger than Tambora 1815 for Northern Europe and **i)** for Central
419 Europe (note the different scale for g-i); shown are JJA temperature anomalies (°C) for 15 years after
420 reconstructed volcanic peak forcing, relative to the five years before the volcanic eruption. Dashed
421 lines present two times the standard error of the mean (2 SEM) of the temperature anomalies
422 associated with the multiple eruptions. 5-year average post-volcanic temperatures are shown for
423 each reconstruction (lag 0 to lag +4 yrs, gray shading).

424 **Figure 5 | Volcanism and temperature variability during the Migration Period (500-705 CE).** Upper
425 panel: Ice-core non-sea-salt sulfur (nssS) records from Greenland (NEEM-2011-S1, TUNU2013).
426 Insets give translations of historical documents describing observation of post-volcanic atmospheric
427 effects in 536-537 and 626-627 CE. Calendar years for five large eruptions are given for the start of
428 volcanic sulfate deposition; middle panel: Summer temperature anomalies for Europe³, and
429 reconstructed N-Tree growth anomalies and occurrence of frost rings in North American bristlecone
430 pine tree-ring records; lower panel: nssS records from Antarctica (WDC, B40) on the WD2014
431 timescale; attribution of the sulfur signals to bipolar, NH, and SH events based on the timing of
432 deposition on the two independent timescales is indicated by shading.

433 **Methods**

434 **Ice cores.** This study included new and previously described ice-core records from five drilling sites
435 (Extended Data Fig. 1, Supplementary Data S1). The upper 577 m of the 3,405 m WAIS Divide (WDC)

436 core from central West Antarctica and a 410 m intermediate-length core (NEEM-2011-S1) drilled in
437 2011 close to the 2,540 m North Greenland Eemian Ice Drilling (NEEM)⁵⁰ ice core previously have
438 been used to reconstruct sulfate deposition in both polar ice sheets¹⁶. These coring sites are
439 characterized by relatively high snowfall ($\sim 200 \text{ kg m}^{-2} \text{ yr}^{-1}$) and have comparable elevation, latitude,
440 and deposition regimes. WDC and NEEM-2011-S1 provided high-resolution records that allowed
441 annual-layer dating based on seasonally varying impurity content¹⁶. New ice-core analyses included
442 the upper 514 m of the main NEEM core used to extend the record of NEEM-2011-S1 to cover the
443 past 2,500 years, as well as B40 drilled in 2012 in Dronning Maud Land in East Antarctica and
444 TUNU2013 drilled in 2013 in Northeast Greenland – both characterized by lower snowfall rates (~ 70 -
445 $100 \text{ kg m}^{-2} \text{ yr}^{-1}$). Volcanic sulfate concentration from B40 had been reported previously for the past
446 2,000 years¹⁷, but we extended measurements to 200 m depth to cover the past 2,500 years.

447 **High-resolution, ice-core aerosol analyses.** Ice-core analyses were performed at the Desert
448 Research Institute (DRI) using 55 to 100 cm long, longitudinal ice core sections (33 x 33 mm wide).
449 The analytical system for continuous analysis included two Element2 (Thermo Scientific) high-
450 resolution inductively coupled plasma mass spectrometers (HR-ICP-MS) operating in parallel for
451 measurement of a broad range of ~ 35 elements; an SP2 (Droplet Measurement Technologies)
452 instrument for black carbon (BC) measurements; and a host of fluorimeters and spectrophotometers
453 for ammonium (NH_4^+), nitrate (NO_3^-), hydrogen peroxide (H_2O_2), and other chemical species. All
454 measurements were exactly co-registered in depth, with depth resolution typically less than 10-15
455 mm⁵¹⁻⁵³. We corrected total sulfur (S) concentrations for the sea-salt-sulfur contribution using sea-
456 salt-Na concentrations¹⁶. Measurements included TUNU2013 and NEEM (400-515 m) in Greenland,
457 and B40 in Antarctica (Extended Data Fig. 1). Gaps (i.e., ice not allocated to DRI) in the high-
458 resolution sulfur data of the NEEM core were filled with ~ 4 cm resolution discrete sulfate
459 measurements using fast ion-chromatography techniques⁵⁴ performed in the field between 428 and
460 506 m depth.

461 Independent analyses of the upper part of the NEEM main core were performed in the field
462 using a continuous flow analysis (CFA) system⁵⁵ recently modified to include a new melter head
463 design⁵⁶. Ca^{2+} , NH_4^+ , and H_2O_2 were analyzed by fluorescence spectroscopy; Na^+ and NO_3^- by
464 absorption spectroscopy; conductivity of the meltwater by a micro flow cell (Amber Science Inc.);
465 and a particle detector (Abakus, Klotz) was used for measuring insoluble dust particle concentrations
466 and size distribution⁵⁷. Effective depth resolution typically was better than 20 mm. Measurements
467 were exactly synchronized in depth using a multicomponent standard solution; the accuracy of the
468 depth assignment for all measurements typically was better than 5 mm.

469 **High-resolution measurements of ^{10}Be in ice cores using accelerator mass spectrometry (AMS).**

470 Samples from the NEEM-2011-S1, WDC, NGRIP, and TUNU2013 ice cores encompassing the time
471 period of the $\Delta^{14}\text{C}$ anomalies from tree-ring records^{12,22-25} were used for ^{10}Be analysis
472 (Supplementary Data S1). NEEM-2011-S1 and WDC were sampled in exact annual resolution, using
473 the maxima (minima in WDC) of the annual cycles of Na concentrations to define the beginning of
474 the calendar year¹⁶. NGRIP was sampled at a constant resolution of 18.3 cm providing an age
475 resolution of about one year. Similarly, TUNU2013 was sampled in quasi-annual resolution according
476 to the average annual-layer thickness expected at this depth based on prior volcanic synchronization
477 to NEEM-2011-S1. The relative age uncertainty for TUNU2013 with respect to the dependent NEEM-
478 2011-S1 chronology at this depth is assumed to be ± 1 year at most given a distinctive match for
479 selected volcanic trace elements in both ice core records (752-764 CE, NS1-2011 timescale). Sample
480 masses ranged between 100 and 450 g, resulting in median overall quantification uncertainties of
481 less than 4–7%. The $^{10}\text{Be}/^9\text{Be}$ ratios of samples and blanks were measured relative to well-
482 documented ^{10}Be standards¹³ by AMS at Purdue's PRIME laboratory (WDC, NEEM-2011-S1,
483 Tunu2013) and Uppsala University (NGRIP)^{58,59}. Results were corrected for an average blank $^{10}\text{Be}/^9\text{Be}$
484 ratio, corresponding to corrections of 2–10% of the measured $^{10}\text{Be}/^9\text{Be}$ ratios.

485 **Annual-layer dating using the StratiCounter algorithm.** For annual-layer interpretation, we used
486 DRI's broad-spectrum aerosol concentration data from WDC (188–577 m), NEEM-2011-S1 (183–411

487 m), and NEEM (410–515 m), as well as NEEM aerosol concentration data (183–514 m) from the field-
488 based CFA system. The original timescale for NEEM-2011-S1 was based on volcanic synchronization
489 to the NGRIP sulfate record on the GICC05 timescale and annual-layer interpretation between the
490 volcanic age markers, while WDC previously was dated by annual-layer counting¹⁶.

491 Parameters with strong intra-annual variability included tracers of sea salt (e.g., Na, Cl, Sr),
492 dust (e.g., Ce, Mg, insoluble particle concentration), and marine biogenic emissions such as non-sea-
493 salt sulfur (nssS). Tracers of biomass-burning emissions, such as BC, NH_4^+ , and NO_3^- , also showed
494 strong seasonal variations in deposition during pre-industrial times^{16,60,61}. Datasets used for annual
495 layer interpretation are provided in Extended Data Table 1. For NEEM-2011-S1, the final database
496 used for annual-layer dating included 13 parameters and the ratio of nssS/Na. For WDC, the final
497 database included five parameters and the ratio of nssS/Na. For NEEM (410-515 m depth), the final
498 database included eight parameters (Na^+ , Ca^{2+} , NH_4^+ , H_2O_2 , NO_3^- , conductivity, insoluble particle
499 concentrations, and ECM⁶²) from the field-based measurements and eleven parameters (Na, Cl, Mg,
500 Mn, Sr, nssS, nssS/Na, nssCa, BC, NO_3^- , NH_4^+) from the DRI system.

501 We focused here on the time period prior to the large volcanic eruption of Samalas in 1257
502 CE³¹, clearly detectable as an acidic peak in both ice-core records, and consequently started annual-
503 layer counting of NEEM-2011-S1, NEEM, and WDC at the depth of the corresponding sulfur signal.
504 For the time period 1257 CE to present, ice-core chronologies were constrained by numerous
505 historic eruptions and large sulfate peaks showing a strong association to Northern Hemisphere (NH)
506 cooling events as indicated by tree-ring records¹⁶.

507 We applied the StratiCounter layer-detection algorithm³² to the multi-parameter aerosol
508 concentration records (n=14 for NEEM-2011-S1; n=6 for WDC; n=8 for NEEM <410 m; n=19 for NEEM
509 >410 m) to objectively determine the most likely number of annual layers in the ice cores along with
510 corresponding uncertainties. The StratiCounter algorithm is based on statistical inference in Hidden
511 Markov Models (HMMs), and it determines the maximum likelihood solution based on the annual

512 signal in all aerosol records in parallel. Some of these displayed a high degree of similarity, so we
513 weighted these records correspondingly lower. The algorithm was run step-wise down the core,
514 each batch covering approximately 50 years, with a slight overlap. All parameters for the statistical
515 description of a mean layer and its inter-annual variability in the various aerosol records were
516 determined independently for each batch as the maximum likelihood solution. The algorithm
517 simultaneously computes confidence intervals for the number of layers within given sections,
518 allowing us to provide uncertainty bounds on the number of layers between selected age-marker
519 horizons (Extended Data Table 2).

520 Annual-layer detection in the NEEM main core below 410 m was made more difficult by
521 frequent occurrence of small gaps in the two independent high-resolution aerosol data sets.
522 Depending on the parameter, data gaps from the CFA field measurements accounted for up to 20%
523 of the depth range between 410 and 515 m, but the combined aerosol records from both analyses
524 provided an almost complete aerosol record with 96% data coverage. As this was the first time that
525 the StratiCounter algorithm was used simultaneously on data records from two different melt
526 systems, with different characteristics and lack of exact co-registration, we also manually
527 determined annual layers below 410 m using the following approaches: one investigator used Na
528 and nssCa concentrations and the ratio of nssS/Na (from DRI analysis) as well as Na^+ and insoluble
529 particle concentrations (from CFA analysis) as primary dating parameters. BC, NH_4^+ , nssS, and
530 conductivity were used as secondary dating parameters where annual-layer interpretation was
531 ambiguous. A second investigator used DRI's Na, Ca, BC, NH_4^+ and CFA Na^+ , Ca^{2+} , and NH_4^+
532 measurements as parameters. The annual-layer interpretation of the NEEM core between 410 and
533 514 m from investigator 1 was within the interpretation uncertainties of the StratiCounter output,
534 from which it differed less than a single year over the majority of this section, and it differed from
535 independently counted timescales (e.g., GICC05)⁶² by on average less than three years (Extended
536 Data Fig. 2). This set of layer counts was used for the resulting timescale.

537 **New ice-core chronologies (NS1-2011, WD2014).** We defined the depth of NEEM-2011-S1
538 containing the maximum ^{10}Be concentration as the year 775 CE. Relative to this constraint, the
539 maximum likelihood ages for three large volcanic sulfate peaks were within ± 1 year of documented
540 historical reports from early written sources of prominent and sustained atmospheric dimming
541 observed in Europe and/or the Near East (Extended Data Table 3, Supplementary Data S2).
542 Automated-layer identification for NEEM-2011-S1 was therefore constrained by tying the respective
543 ice-core volcanic signals to the corresponding absolute historically-dated ages of 536, 626, and 939
544 CE (Extended Data Table 2) – thereby creating a new ice-core timescale (NS1-2011). The volcanic
545 sulfur signal corresponding to the eruption of Samalas believed to have occurred in late 1257³¹ was
546 constrained to 1258 CE to account for several months' delay in sulfate deposition in the high
547 latitudes. Before 86 CE (the bottom depth of NEEM-2011-S1), the NS1-2011 timescale was extended
548 using the manually-derived annual-layer interpretation of the combined NEEM aerosol data-sets
549 back to 500 BCE (Fig. 2).

550 In NS1-2011 we did not attribute acid layers to the historical eruptions Vesuvius 79 and
551 Hekla 1104, due to a lack of corroborative tephra at these depths in this and a previous study⁶³.
552 Possible Vesuvian tephra was reported from the Greenland Ice Sheet Project (GRIP) ice core at 429.3
553 m depth⁶⁴, but in view of the new annual-layer dating results (Extended Data Fig. 3), we concluded
554 that this layer dates to 87/88 CE. Furthermore, volcanic sulfate deposition values for the
555 corresponding event show a strong spatial gradient over Greenland with highest values in NW
556 Greenland¹⁶ and lowest in Central and South Greenland⁶⁵, favoring the attribution of a volcanic
557 source from the high latitudes. Documentary sources (Supplementary Data S2) also suggest that the
558 main vector of ash transport following the Vesuvius 79 CE eruption was toward the eastern
559 Mediterranean⁶⁶.

560 For WDC, we do not have other sufficiently well-determined age constraints besides the
561 rapid ^{10}Be increase in 775 CE and the sulfur signal of the Samalas 1257 eruption. Therefore, no

562 additional constraints were used when creating the new ice-core timescale (“WD2014”) from the
563 StratiCounter annual-layer interpretation back to 396 BCE.

564 Depth-age information for six distinctive marker horizons in Greenland is given, and five of
565 these horizons were used to constrain NS1-2011 (Extended Data Table 3). Similarly, depth
566 information, the number of annual layers, and 95% confidence intervals between distinctive volcanic
567 marker horizons are given for NEEM, NEEM-2011-S1, and WDC, supporting attribution of these ice-
568 core signals to eruptions in the low latitudes with bipolar sulfate deposition.

569 **Evaluation of NS1-2011 using independent age information.** We evaluated timescale accuracy using
570 additional distinctive age markers not used during chronology development:

- 571 1) Tephra from the eruption of Changbaishan/Tianchi (China)⁶⁷ was detected in NEEM-2011-S1
572 in 946–947 CE, in agreement with widespread documentary evidence of an eruption in that
573 region in winter 946/47 CE⁶⁸ also supported by a high-precision ¹⁴C wiggle-match age of 946 ±
574 3 CE obtained from a tree killed during this eruption⁶⁸.
- 575 2) The rapid increase of ¹⁰Be from the 994 event occurred in NEEM-2011-S1 in 993 CE,
576 consistent with Δ¹⁴C from Japanese tree rings showing that the rapid increase in
577 radionuclide production took place between the NH growing seasons of 993 and 994 CE²³.
- 578 3) To assess the accuracy of the NS1-2011 timescale prior to the earliest age marker at 536 CE,
579 we compiled an independent time series of validation points, featuring years with well dated
580 historical reports of atmospheric phenomena associated with high-altitude volcanic dust
581 and/or aerosols (Supplementary Data S2) as known from modern observations to occur after
582 major eruptions (e.g., Krakatau, 1883). These phenomena include diminished sunlight,
583 discoloration of the solar disk, solar coronae (i.e., Bishop’s Rings), and deeply red twilights
584 (i.e., volcanic sunsets)^{69,70}. Thirty-two events met our criteria as validation points for the pre-
585 536 CE NS1-2011 timescale. For the earliest in 255 BCE, it was reported in Babylon that “the
586 disk of the sun looked like that of the moon”⁷³. For the latest in 501 CE, it was reported in

587 North China that “the Sun was red and without brilliance”⁷⁴. We found that NEEM volcanic
588 event years (including both NEEM and NEEM-2011-S1 data) occurred closely in time (i.e.,
589 within a conservative ± 3 year margin) to 24 (75.0%) of our validation points (Extended Data
590 Figure 2). To assess whether this association arose solely by chance, we conducted a Monte
591 Carlo equal means test with 1,000,000 iterations (Supplementary Data S2) and found that
592 the number of volcanic event years within ± 3 years of our validation points was significantly
593 greater than expected randomly ($p < 0.001$). A significant association was also observed
594 ($p < 0.001$) when using less conservative error margins (± 1 and ± 2 years) and when excluding
595 any historical observations with less certainty of a volcanic origin (Supplementary Data S2).
596 When placing volcanic event years on the original GICC05 timescale, we did not observe any
597 statistically significant association with our independent validation points.

598 **Potential causes of a previous ice-core dating bias.** Interpretation of annual layers in ice cores is
599 subject to accumulating age uncertainty due to ambiguities in the underlying ice-core profiles^{30,73}.
600 Bias in existing chronologies may arise from several factors, including: 1) low effective resolution of
601 some ice core measurements (NGRIP, GRIP); 2) use of only single (or few) parameters for annual-
602 layer interpretation (GRIP, Dye-3); 3) intra-annual variations in various ice-core parameters falsely
603 interpreted as layer boundaries (e.g., caused by summer melt in Dye-3)⁷⁴; 4) use of tephra believed
604 to originate from the 79 CE Vesuvian eruption⁶⁴ as a fixed reference horizon to constrain the
605 Greenland ice-core dating³⁰; 5) use of manual-layer interpretation techniques that may favor
606 interpretations consistent with *a priori* knowledge or existing chronologies (WDC)^{16,21}.

607 **Volcanic synchronization of B40, TUNU2013, and NGRIP to WDC and NEEM.** Two high-resolution
608 sulfur ice-core records (TUNU2013, Greenland and B40, Antarctica) were synchronized to NEEM-
609 2011-S1 and WDC, respectively, using volcanic stratigraphic age markers¹⁷ with relative age
610 uncertainty between the tie-points estimated to not exceed ± 2 years. The NGRIP sulfate record
611 measured at 5 cm depth resolution¹⁵ similarly was synchronized to NS1-2011 using 124 volcanic tie-
612 points between 226 and 1999 CE. During the time period with no sulfur record yet available for WDC

613 (before 396 BCE), a tentative chronology for B40 was derived by linearly extrapolating mean annual-
614 layer thickness for B40 as derived from the synchronization to WDC between the earliest volcanic
615 match points.

616 **2,500 year global volcanic forcing ice-core index.** We constructed an index of global volcanic aerosol
617 forcing by (1) re-dating and extending to 500 BCE an existing reconstruction of sulfate flux from an
618 Antarctic ice-core array¹⁷ by applying an area weighting of 80/20 between East Antarctica and West
619 Antarctica to B40 and WDC volcanic sulfate flux values, respectively; (2) compositing NGRIP and the
620 NEEM-2011-S1/NEEM sulfate flux records to a similar Greenland sulfate deposition composite back
621 to 500 BCE; (3) using established scaling functions^{6,75} to estimate hemispheric sulfate aerosol loading
622 from both polar ice-core composites; and (4) scaling global aerosol loading to the total (i.e., time
623 integrated) radiative volcanic aerosol forcing following the Tambora 1815 eruption⁷. Since the NS1-
624 2011 and WD2014 timescales are independent of each other, the timing of bipolar events had to be
625 adjusted to follow a single timescale to derive a unified global volcanic forcing series. We chose NS1-
626 2011 as the reference chronology for most of the volcanic time series because this age model was
627 constrained and validated by more stratigraphic age markers than WD2014. WD2014 was used as
628 the reference chronology only between 150 and 450 CE, because of better data quality during that
629 time period. TUNU2013 was not included in the Greenland ice-core composite because annual-layer
630 thickness variability at this site is influenced strongly by glaciological processes, leading to relatively
631 large uncertainties in atmospheric sulfur-deposition determinations.

632 **NH tree-ring composite.** Tree-ring records from certain locations reflect summer cooling (as is
633 widespread observed after volcanic eruptions) with no age uncertainty in annual ring-width dating,
634 thus allowing independent validation of ice-core timescales and the derived volcanic forcing indices.
635 However, no tree-ring-based temperature reconstructions of large spatial scales span the full 2,500
636 years represented by our new ice-core chronologies. To thus evaluate our new ice-core chronologies
637 and assess the consistency of response throughout the past 2,500 years, we compiled a composite

638 (entitled “N-Tree”) of multi-centennial tree growth records at locations where temperature is the
639 limiting growth factor. We selected available NH tree-ring records that provided a continuous record
640 of >1,500 years and showed a significant positive relationship with JJA temperatures during the
641 instrumental period (1901–2000 CE) with $p < 0.005$ (adjusted for a reduced sample size due to
642 autocorrelation of the datasets). In total, five tree-ring chronologies (three based on ring-width
643 measurements, two based on measurements of maximum latewood density) met these
644 criteria^{42,43,76-78} of which three are located in the high-latitudes of Eurasia (Extended Data Figure 1).

645 As various climatic and non-climatic parameters may influence sensitivity of tree growth to
646 temperatures during the 20th century⁷⁹⁻⁸¹, we used the time period 1000-1099 CE as a common
647 baseline for standardizing tree growth anomalies among the five chronologies and built a tree
648 growth composite record “N-Tree” (z-scores) by averaging the individual records. Correlations
649 between “N-Tree” (N=5) and the average of three regional reconstructions for the Arctic, Europe,
650 and Asia (N>275)³ between 1800 and 2000 CE are very high ($r = 0.86$, $N=201$, $p < 0.0001$), suggesting
651 that much of the large-scale variation in temperature is explained by these selected tree-ring
652 records. Three records in “N-Tree” cover the period from 138 BCE to the present, thus allowing at
653 least a qualitative assessment of the coherence of growth reduction following large volcanic
654 eruptions prior to the Common Era (Fig. 2, Extended Data Fig. 4).

655 **Temperature reconstructions.** To quantify the CE climate impact and investigate regional
656 differences, we used tree-ring-based JJA temperature reconstructions covering the past 2,000 years
657 with demonstrated strong relationship ($r \geq 0.45$; $p < 0.0001$; Extended Data Fig. 1) to instrumental JJA
658 temperature data⁸² between 1901 and 2000. For regions where this criterion was met by several
659 reconstructions (e.g., Scandinavia), we limited the analysis to the most recently updated
660 reconstruction³⁵. Three regional reconstructions from Central Europe⁴², Northern Europe³⁵, and
661 Northern Siberia (Yamal, not shown)⁷⁶ as well as a continental-scale reconstruction for Europe³ met
662 this criterion and were used to quantify the average response of summer temperature to volcanic
663 forcing during the Common Era (Figs. 3, 4).

664 **Superposed epoch analyses.** To assess tree-ring growth reduction and summer cooling following
665 large eruptions, we used superposed epoch analyses^{83,84}. We selected all volcanic eruptions (28
666 events in total, 24 CE events) with time-integrated volcanic forcing greater than -7.5 W m^{-2} (i.e.,
667 eruptions larger than Pinatubo 1991) and aligned the individual segments of “N-Tree” and regional
668 JJA temperature reconstructions relative to ice-core-indicated peak forcing. Composite response was
669 calculated for the average of the individual series (lag 0 to lag 10 or 15 years) relative to the average
670 values five years prior to individual volcanic events (lag -5 to lag -1 year). 95% confidence intervals
671 represent 2 SEM of the tree-growth (Extended Data Figure 4) and temperature anomalies (Figure 4)
672 associated with the multiple eruptions.

673 **Cryptotephra analyses of the 536 CE sample from NEEM-2011-S1.** We analyzed samples from
674 NEEM-2011-S1 for tephra between 326.73 and 328.06 m depth, corresponding to 531–539 CE (NS1-
675 2011 timescale). Samples (200 to 500 g) were filtered, and elemental composition of recovered
676 volcanic glass shards determined by electron microprobe analysis (EPMA) at Queen's University
677 Belfast using established protocols^{63,67,85} and secondary glass standards^{86,87}. Between 326.73 and
678 327.25 m, large volume samples were cut at 8 cm depth resolution ($\leq 0.5 \text{ yr}$) and with an average
679 cross section of 26 cm^2 . Between 327.25 and 328.06 m, the average cross section was 7 cm^2 and
680 depth resolution 20 cm ($\sim 1 \text{ yr}$ resolution). Tephra particles ($n \geq 17$) were isolated from a sample of ice
681 (327.17–327.25 m depth, 251 g) corresponding to the sulfate spike at 536 CE. The glass shards were
682 heterogeneous in size (20–80 μm), morphology (platey, blocky, vesicular, microlitic), and
683 geochemistry (andesitic, trachytic, rhyolitic). Individual shards had geochemical compositions that
684 share affinities with volcanic systems in the Aleutian arc (Alaska)⁸⁸, Northern Cordilleran volcanic
685 province (British Columbia)⁸⁹, and Mono-Inyo Craters area (California)^{90,91} – indicating at least three
686 synchronous eruptive events, all situated in western North America between 38 and 58°N (Extended
687 Data Fig. 5; Supplementary Data S5).

688 **Data:** Ice-core data (chemistry, including sulfur; ¹⁰Be), resulting timescales, and the volcanic forcing
689 reconstruction are provided as online supplementary material (Supplementary Data S1, S3-S5);
690 Historical documentary data is provided as Supplementary Data S2. The code for the StratiCounter
691 program is accessible at the github repository (<http://www.github.com/maiwinstrup/StratiCounter>);
692 NGRIP SO₄ data can be obtained at [http://www.iceandclimate.nbi.ku.dk/data/2012-12-](http://www.iceandclimate.nbi.ku.dk/data/2012-12-03_NGRIP_SO4_5cm_Plummet_et_al_CP_2012.txt)
693 [03_NGRIP_SO4_5cm_Plummet_et_al_CP_2012.txt](http://www.iceandclimate.nbi.ku.dk/data/2012-12-03_NGRIP_SO4_5cm_Plummet_et_al_CP_2012.txt) ; tree-ring records and temperature
694 reconstructions are from Pages-2k Consortium (Database S1, S2)
695 (<http://www.nature.com/ngeo/journal/v6/n5/full/ngeo1797.html#supplementary-information>).

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804 **Extended Data Figure 1 | Location of study sites. a)** Map showing locations of the five ice-cores
805 (WDC, B40, NEEM, NGRIP and TUNU) used in this study. Sites of temperature-limited tree-ring
806 chronologies (green)^{42,43,76-78} and sites with annual $\Delta^{14}\text{C}$ measurements from tree-rings in the 8th
807 century CE (red outline) are marked; **b)** metadata for the ice cores, tree-ring chronologies and
808 temperature reconstructions used.

809 **Extended Data Figure 2 | Volcanic dust veils from historical documentary sources in relation to**
810 **NEEM.** Time series of 32 independently-selected chronological validation points from well-dated
811 historical observations of atmospheric phenomena with known association to explosive volcanism
812 (e.g., diminished sunlight, discolored solar disk, solar corona or Bishop's Ring, red volcanic sunset) as

813 reported in the Near East, Mediterranean region, and China, prior to our earliest chronological age
814 marker at 536 CE. Black lines represent the magnitude (scale on vertical y-axes) of annual sulfate
815 deposition measured in NEEM (NEEM and NEEM-2011-S1 ice cores) from explosive volcanic events
816 on the new NS1-2011 timescale. Red crosses depict the 24 (75%) historical validation points for
817 which NEEM volcanic events occur within a conservative ± 3 year uncertainty margin. Blue crosses
818 represent the eight points for which volcanic events are not observed. The association between
819 validation points and volcanic events is statistically significantly non-random at >99.9% confidence.

820 **Extended Data Figure 3 | Timescale comparison.** Age differences of the timescales **a)** NS1-2011 and
821 GICC05 for the NEEM-2011-S1/NEEM ice cores and **b)** WD2014 and WDC06A-7 for WDC. Differences
822 before 86 CE (the bottom age of NEEM-2011-S1) deriving from the annual-layer counting of the
823 NEEM core are shown for major volcanic eruptions relative to the respective signals in NGRIP on the
824 annual-layer counted GICC05 timescale. Marker events used for constraining the annual-layer dating
825 (solid line) and for chronology evaluation (dashed lines) are indicated. Triangles mark volcanic
826 signals. Also indicated is the difference between WD2014 and the Antarctic ice-core chronology
827 (AICC2012)⁹², based on volcanic synchronization between the WDC and EDC96 ice cores.

828 **Extended Data Figure 4 | Post-volcanic suppression of tree growth.** Superposed epoch analysis for
829 large volcanic eruptions using the **a)** 28 largest volcanic eruptions; **b)** 23 largest tropical eruptions; **c)**
830 five largest NH eruptions; and **d)** eruptions larger than Tambora 1815 with respect to sulfate aerosol
831 loading. Shown are growth anomalies of a multi-centennial tree-ring composite record (N-Tree) 15
832 years after the year of volcanic sulfate deposition, relative to the average of five years before the
833 events. Dashed lines indicate 95% confidence intervals (2 SEM) of the tree-ring growth anomalies
834 associated with the multiple eruptions.

835 **Extended Data Figure 5 | Major element composition for ice core tephra QUB-1859 and reference**
836 **material.** Shown are selected geochemistry data: **(a)** SiO₂ vs. total alkali (K₂O + Na₂O); **(b)** FeO (total
837 iron oxides) vs. TiO₂; **(c)** SiO₂ vs. Al₂O₃; and **(d)** CaO vs. MgO) from 11 shards extracted from the NEEM-

838 2011-S1 ice core between 327.17 and 327.25 m depth, representing the age range 536.0–536.4 CE
839 on the new, NS1-2011 timescale. Data for Late Holocene tephra from Mono Craters (California) are
840 from the compilation by ref. 90 (all references in Methods); data for Aniakchak (Alaska) are from
841 reference material published by ref. 88; and data for the early Holocene upper Finlay tephra ,
842 believed to be from the Edziza complex in the Upper Cordilleran Volcanic province (British
843 Columbia), are from ref. 89.

844 **Extended Data Table 1 | Ice-core dating.** Parameters used for annual-layer interpretation.
845 Parameters measured by the CFA system in the field are underlined. Stratigraphic age marker used
846 to constrain annual-layer counting (*) and horizons used to evaluate the timescale (†).

847 **Extended Data Table 2 | Annual-layer results using the StratiCounter program.** Maximum-
848 likelihood number of annual layers and confidence intervals derived from annual-layer counting
849 between distinctive marker horizons and corresponding ages relative to the 775 CE ¹⁰Be event.

850 *UE: Unattributed volcanic signal and year of sulfate deposition based on final age models (negative
851 numbers are Year BCE).

852 †Year (BCE/CE) calculated from the number of annual layers relative to the fixed age marker in 775
853 CE.

854 ‡Depth has been estimated from the average depth offset between NEEM-2011-S1 and NEEM.

855 §Fixed age marker based on the ¹⁰Be maximum annual value.

856 || Section with 6 m gap in the NEEM 2011-S1 core DRI data (this section is not used for calculating
857 average age).

858 ¶ This section is based on the NEEM field CFA data, since the DRI data does not cover the entire
859 interval.

860 # Section is based on combined data set of DRI and field-measured CFA data. The number of annual
861 layers in this section from manual interpretation by investigator 1 was 383 (±7), and that of

862 investigator 2 was 393 (± 8) layers. Most of the difference between the three layer counts was
863 occurring below 480 m (i.e., before 300 BCE), where data gaps were more frequent.

864 ☆Independent age markers used to constrain annual-layer dating in a second iteration to derive the
865 final ice-core age model NS1-2011.

866 **Tephra particles were extracted from the depth range 327.17–327.25 m depth (see
867 Supplementary Data).

868 ††Unattributed volcanic signal that was previously attributed to the historic 79 CE eruption of
869 Vesuvius⁶⁴.

870 **Extended Data Table 3 | Historical documentary evidence for key volcanic eruption age markers**

871 **536-939 CE.** A comprehensive list of all sources, including translations and assessment of the
872 confidence placed in each source and its chronological information is given in Supplementary Data.

873 **Extended Data Table 4 | Large volcanic eruptions during the past 2,500 years.** Years with negative

874 numbers are before the Common Era (BCE). Tentative attribution of ice-core signals to historic
875 volcanic eruptions is based on the Global Volcanism Program volcanic eruption database⁹³. Average
876 (summer) temperature for the associated cold year is given for the average of Europe and the
877 Arctic³.

878 *Total global aerosol forcing was estimated by scaling total sulfate flux from both polar ice sheets to
879 the reconstructed total (i.e., time integrated) aerosol forcing for Tambora 1815⁷ (Methods); for high
880 latitude NH eruptions, Greenland fluxes were scaled by a factor of 0.57⁶.

881 † Unattributed volcanic events (UE) and tentative attributions for non-documented historic
882 eruptions (?) are marked.

883 **Extended Data Table 5 | Post-volcanic cooling.** Coldest years and decades (1–2000 CE, JJA

884 temperature wrt. 1901–2000) for Europe³ and years (500 BCE–1250 CE) and decades (500 BCE–2000
885 CE) with strong growth reduction in the N-Tree composite(wrt. 1000–1099). Ages of the volcanic

886 events from the ice cores reflect the start of volcanic sulfate deposition in Greenland (NS1-2011

887 timescale) with the largest 40 events indicated in bold letters and tropical eruptions underlined.

888 Years with negative numbers are before the Common Era (BCE).

889 * Latewood frost ring in bristlecone pines within ± 1 year³⁴.









