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Supplementary information

Abrupt Common Era hydroclimate shifts drive west Greenland ice cap change

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Abrupt Common Era hydroclimate shifts drive west Greenland ice cap change

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Supplementary Material

Contents:

1. Geodetic measurements

We generated a digital elevation model for the ice cap using photogrammetric image data from the Worldview-1 satellite, collected 9 September, 2012. The imagery was converted to a Digital Elevation Model (DEM; Fig. 2a) using the Ames Stereo Pipeline software package[1]. Spot checks against field-collected GPS data suggest that the DEM is accurate to within 1-2 m over the ice sheet.

2. Ice-penetrating radar measurements

We conducted a ground-based ice-penetrating radar (IPR) survey on the south face of the ice cap during April 2014 using a monopulse radar system with 7-MHz loaded-dipole antennas. The survey followed four flowlines and two contour lines spanning an area of around 1x3 km that appeared crevasse-free on satellite imagery. Towards the end of the survey, a connector failed on the radar system, leaving us with data on three out of the four flow lines. Additional data were collected by NASA's Operation Ice Bridge which overflew the ice cap three times between 2014 and 2016 and measured ice thickness with the MCCORDS radar system. Radargrams from both systems showed clear bed and surface returns, but no visible internal stratigraphy. On the radargrams from our IPR survey, and on the level-1b IceBridge data products, we digitized the surface and bed returns and calculated the ice thickness based on the difference between the two using a wave velocity of 168 m μs^{-1}. Integrating the porosity from density measurements in the ice core from the SE flank of the ice cap gave an equivalent air column of 11.3 m, which implies a firn-air correction of 5 m for ice-thickness estimates.

We converted the ice-thickness measurements into estimates of the bed elevation using our surface digital elevation model (Fig. 2a). The measurements reflect a very smooth domed structure beneath the central part of the ice cap. Because our data distribution was somewhat irregular, and because the structure of the height measurements appeared very smooth, we interpolated bed-elevation between our measurements using a second-degree polynomial fit to the bed elevations: $Z_p = 72.8x + 9.5y - 59.4xy - 20.7x^2 - 33.1y^2$. Here, x and y are offsets (km) relative to 70.489˚N and 52.263˚W (the point we eventually chose as our core location) in the Greenland polar stereographic projection. This polynomial matches our bed-elevation estimates with a root-mean-squared error of 1.5 m which is well within the expected error bounds for the data.

3. Ground-penetrating radar measurements

The 2014 radar survey team also carried a GSSI ground-penetrating radar (GPR) operating at 400 MHz. We processed the data with a bandpass filter with cutoff frequencies of 150 and 750 MHz. The data showed strongly disturbed from the surface to the lower limit of the measurements, at around 17 m depth. We found one layer that appeared to be marginally continuous on a path from the summit to the east along the ridge, and back along one of the flowlines to the summit. The depth to this layer varied from around 1.6 m at the summit to around 3.7 m at the eastern edge of the survey. Other layers are apparently continuous over shorter distances in other parts of the survey, but they were sufficiently disturbed that we did not feel that picking their depths would add useful information to the data.

4. Final ice core site selection

At the time that we needed to pick a core location, we had a high-resolution photogrammetric DEM of the ice cap, the 2014 ice-thickness measurements from IPR, and layering measurements from GPR. At the time, we interpreted the layer in the GPR measurements as an isochrone, which implied that the accumulation at the summit was about half that on the SE flank of the ice cap. Under this interpretation, the accumulation gradient was sharpest in within 100 m of the summit, and the accumulation rate was relatively constant beyond this. Although we did not have ice-thickness measurements on the flowline for which we had GPR thickness measurements, the close agreement between the polynomial fit and the bed-elevation measurements suggested that the bed elevation should be well constrained at a point around 250 m downhill from the summit. By not drilling exactly at the summit, we expected to avoid potential disturbances in the layering caused by small-scale orographic accumulation variations, and to avoid some potential disturbances caused by divide migration in the past. Our final selected location was 70.489˚N, 52.263˚W, for which we estimated the ice thickness, based on our bed and surface models, to be 153 m, with an uncertainty of 5-10 m based on the bandwidth of the ground based IPR and the rise time of the bed return in the airborne radar data. The ice-equivalent thickness, equal to the measured thickness minus the equivalent air column height calculated from the density profile, is 142 m_{ice}.

5. Nuussuaq ice cap strain inversion

We formed an initial estimate of the age-depth scale for our core based on a thinning function, $C(t)$, that gives the ratio of the ice-equivalent thickness of a layer deposited in year t to its original thickness when it was deposited at the surface, and a time-variable layer thickness, $\theta_0(t)$. In this model, the thickness of any layer found in the core is:

$$\theta_c(t) = C(t)\theta_0(t) \quad . \qquad \textbf{(S1)}$$

The depth for the age-depth scale for time t is found by adding the thicknesses of all layers with ages less than t. The thinning function $C(t)$ is estimated based on a one-dimensional flow model for a constant accumulation rate, in which the vertical strain rate, $\dot{\epsilon}(z)$, at any height z above the ice cap bed $0 \leq z \leq H$ (where H is the surface height) is proportional to the horizontal velocity, $u(z)$. If the basal velocity is zero, and if the ice temperature does not vary strongly with depth, then integrating Glen's flow law[2–4] from the bed to height z gives:

$$u(z) = u_H \left[1 - \left(1 - \frac{z}{H}\right)^{n+1}\right], \qquad \textbf{(S2)}$$

where the creep exponent n = 3 is prescribed following convention[2]. If $\dot{\epsilon}(z)$ is proportional to $u(z)$, then

$$\dot{\epsilon}(z) = -\dot{\epsilon}_H \frac{u(z)}{u_H} \, . \qquad \textbf{(S3)}$$

where $\dot{\epsilon}_H$ and u_H denote the vertical strain rate and horizontal velocity at the ice cap surface, respectively. The integral of the vertical strain rate from the bed to the surface gives the vertical velocity, w, at the surface[2,4], equal to:

$$w_H = -\dot{\epsilon}_H \left(H \frac{n+1}{n+2}\right), \qquad \textbf{(S4)}$$

or, solving for $\dot{\epsilon}_H$ gives the surface vertical strain rate:

$$\dot{\epsilon}_H = -\frac{w_H}{H}\frac{n+2}{n+1} \ . \quad \textbf{(S5)}$$

This expression is similar to that for the mean strain rate (equal to the accumulation rate divided by the ice thickness) but because the strain rate near the bed is small, the surface strain rate must be larger than the mean. The final expression for the vertical strain rate as a function of height above the bed is then:

$$\dot{\epsilon}(z) = \dot{\epsilon}_H \left[1 - \left(1 - \frac{z}{H}\right)^{n+1}\right] \ . \quad \textbf{(S6)}$$

The vertical velocity at any depth is found by integrating $\dot{\epsilon}(z)$ from the bed to z:

$$w(z) = -\dot{\epsilon}_H H \left[1 - \frac{1}{n+2} - \left(\mu - \frac{1}{n+2}\ \mu^{n+2}\right)\right] \ . \quad \textbf{(S7)}$$

Here, $\mu = 1 - \frac{z}{H}$. An idealized steady-state age-depth scale for time t can be found by integrating the position of a particle as it traverses the vertical-velocity field from the surface at time - t to t = 0 using the vertical-velocity depth profile (**S7**). The thinning function $C(t)$ is the exponential of $\dot{\epsilon}_H$, integrated as a function of time along a layer's path from H to its final depth, z:

$$C(t) = exp\left[\int_{-t}^{0} \dot{\epsilon}(z(t))dt\right] \ . \quad \textbf{(S8)}$$

Age-depth scales calculated using this scheme depend on the initial layer thickness (i.e., accumulation rate) and the total ice thickness. If the accumulation rate varies in time, then during the time between a layer's deposition at the surface and its recovery in an ice core, the ice cap thickness will change, and the thinning rate will in general be different from that given by **S8** for constant w_H. However, as long as the thickness variation over the lifetime of a layer is small, we expect that layer thicknesses calculated with **S8** will not differ substantially from those of a steady-state model with w_H equal to the temporal mean of the accumulation rate, $\dot{b}_0$. Under this assumption, we calculate an age-depth scale from **S1** using $C(t)$ from **S8** and $w_H = -\dot{b}_0$ in **S5**. Under this assumption, we expressed the age-depth scale, $\delta(t)$, as a linear function of $\dot{b}(z(t))$ and a non-linear function of $\dot{b}_0$ and the ice cap "reference" thickness, H_0:

$$\delta(t) = \dot{b}_0 \sum_{i=0}^{t} s_i C(t_i) \ . \quad \textbf{(S9)}$$

Here s_i is a set of scaling values quantifying the ratio between the original annual layer thickness, $\dot{b}_i$, and $\dot{b}_0$ for each year i of the model. Age constraint depths specify a few of the estimated 1935 annual layers in the core, but to calculate $\delta(t)$ we need to find estimates of s_i for the remaining layers. Alone, **(S9)** does not allow a unique solution for these quantities for all $\dot{b}_0$ and H, so we look for a set of solutions that give the least-complex accumulation histories that match the measured layer dates to within their estimated age-constrained depth uncertainties (σ_k). These solutions minimize the cost function:

$$J = \sum_{k=1}^{n} \left(\frac{\delta(t_k, \dot{b}_0, H_0) - \delta_{0,k}}{\sigma_k}\right)^2 + \lambda \sum_{i=0}^{t} (s_i - 1)^2 \ . \quad \textbf{(S10)}$$

The scalar λ is a weight value that specifies the relative importance of the layer thicknesses and the data misfits in determining J. Minimizing the first term improves the match between δ_t at the n age-constrained depths ($\delta_{0,k}$), minimizing the second term yields a simpler solution; minimizing both at the same time gives solutions that balance between the two, the particulars of the solution depending on the choice of λ.

The n age-constrained depths only provides information about the thickness of groups of layers, not on individual layers, so to further simplify the problem, we assume that s is constant between pairs of depth picks, thus specifying a mean accumulation rate for the interval t_k to t_{k+1}. With this simplification, **(S10)** becomes:

$$J = \sum_{k=1}^{n} \left(\frac{\delta(t_k, \dot{b}_0, H_0) - \delta_{0,k}}{\sigma_k} \right)^2 + \lambda \sum_{i=0}^{k} (t_{k+1} - t_k)(s_k - 1)^2 \ . \quad \textbf{(S11)}$$

Here s_k is the constant scaling value applied to all of the layer thicknesses between two measured layer ages. To express this equation as a set of matrix multiplications, we use the notation:

$$J = (\boldsymbol{Gs} - \delta_0)^T \boldsymbol{\Sigma}^{-1} (\boldsymbol{Gs} - \delta_0)^T + \lambda(\boldsymbol{s} - 1)^{\boldsymbol{T}} \boldsymbol{W} (\boldsymbol{s} - 1) \ , \quad \textbf{(S12)}$$

with bolded-capitol denoting a matrix, bolded-lowercase a vector and unbolded-lowercase a scalar. Here,

$$\boldsymbol{G} = \begin{bmatrix} g_{11} & 0 & \cdots & 0 & \cdots & 0 \\ g_{21} & g_{22} & \cdots & 0 & \cdots & 0 \\ \vdots & \vdots & \ddots & \vdots & \ddots & \vdots \\ g_{k1} & g_{k2} & \cdots & g_{kk} & \cdots & 0 \\ \vdots & \vdots & \ddots & \vdots & \ddots & \vdots \\ g_{n1} & g_{n2} & \cdots & g_{nk} & \cdots & g_{nn} \end{bmatrix}, \quad \textbf{(S13)}$$

wherein,

$$g_k = \dot{b}_0 \int_{t_{k-1}}^{t_k} C(t, \dot{b}_0, H_0) dt \ , \quad \textbf{(S14)}$$

That is, $\boldsymbol{G}$ is a lower-triangular n x n matrix comprised of idealized ice equivalent thicknesses (predicted ice equivalent thicknesses under some combination of prescribed steady-state $\dot{b}_0$ and H_0) between the surface and n age-constrained depths. The n x 1 vector $\boldsymbol{s}$ consists of the n mean accumulation rate scale values for intervals t_k to t_{k+1}. $\boldsymbol{\Sigma}$ is a diagonal matrix consisting of the n squared depth uncertainty values (σ^2). In the second term, $\boldsymbol{W}$ is a diagonal matrix consisting of δ_t = $[(t_1 - 0) \ldots (t_k - t_{k-1}) \ldots (t_n - t_{n-1})]$.

For a given value of λ, J is minimized by $\boldsymbol{s}$ for which $\frac{dJ}{ds} = 0$; that is, the least squares solution to **(S14)**. We assume that if the error estimates for our layer picks are well prescribed and the errors Gaussian normal, the first term of **(S12)** for the true age-depth relationship will have a χ^2-distribution with $n - 1$ degrees of freedom. We solved **(S12)** for $\boldsymbol{s}$ for different values of λ, which gave us a range of χ^2 values. We identified the $\boldsymbol{s}$ that corresponds to $\chi^2 = n - 1$ to be the locally optimum model: the simplest model that matches the data to within the specified error tolerances under prescribed $\dot{b}_0$ and H_0.

We summarize the mapping between data and model in Eq. 1 of the main text as $T(\boldsymbol{m}, \dot{b}_0, H_0)$, which is equivalent to $\boldsymbol{Gs}$ for $\boldsymbol{G}$ evaluated for $\dot{b}_0$ and H_0. Likewise, $\boldsymbol{m}$ is equivalent to $\boldsymbol{s}\dot{b}_0$.

Table S1: Age-constraint depths identified for the NU core, with attribution to the event horizon, and parameters used to identify it.

Year (CE)	Depth (m w. eq.)	Uncertainty (1σ yrs)	Tie attribution	Parameter(s) used	Citation
169.5	114.63	50	Volcanic eruption - ?	ECM, Cond.	[5]
269.5	113.83	15	Volcanic eruption - ?	ECM, S, Cond.	[5,6]
281.5	113.62	10	Volcanic eruption - ?	ECM, Cond.	[5]
424.5	112	5	Volcanic eruption - ?	ECM, S, Cond.	[5,6]
536.5	110.8	5	Volcanic eruption - El Chichon, Mexico	ECM, Cond., S, Pb	[5,6]
574.5	110.37	5	Volcanic eruption - El Salvador	Cond., S, Pb	[5]
626.5	109.77	5	Volcanic eruption - ?	ECM, Cond., S, Pb	[5]
682.5	108.97	5	Volcanic eruption - Jombolok, Central Asia	ECM, Cond., S, Pb	[5]
750.5	107.59	2	Volcanic eruption - ?	S, Cond., Part.	[5]
777.5	106.91	3	Pb - step function	Pb	[6]
853.5	105.45	1	Volcanic eruption - ?	Pb	[5]
879.5	104.93	2	Volcanic eruption - ?	S	[5]
903.5	104.41	1	Volcanic eruption - ?	S	[5]
940.5	103.51	1	Volcanic eruption - Eldgjá, Iceland	S, Pb, Cond.	[5]
977.5	102.51	2	Volcanic eruption(s) - ?	S, Pb	[5,6]
1020.5	101.41	2	Volcanic eruption - ?	S, Pb, Cond.	[5]
1028.5	101.23	1	Volcanic eruption - ?	Part., S, Pb, Cond.	[5]
1110.5	98.81	5	Volcanic eruption - Mt. Asama, Japan	S	[5,6]
1145.5	97.815	3	Pb - step function	Pb	[6]
1159.5	97.27	3	Pb - step function	Pb	[6]
1172.5	96.69	3	Volcanic eruption - ?	S	[5]
1182.5	96.33	3	Volcanic eruption - ?	S, Pb, Cond.	[5]
1192.5	96.01	3	Volcanic eruption - ?	S, (Pb), Cond.	[5]
1201.5	95.75	2	Volcanic eruption - ?	S, Pb, Cond.	[5]
1222.5	94.81	2	Pb-depositional horizon	Pb	[6]
1231.5	94.53	3	Volcanic eruption - Lipari, Italy	S, Pb	[5,6]
1259.5	93.353	3	Volcanic eruption - Samalas, Indonesia	S	[5]
1288.5	92.15	5	Volcanic eruption - ?	S, Pb	[5,6]
1330.5	89.938	2	Volcanic eruption - Mt. Etna, Italy	S, Pb, Cond.	[5]
1345.5	89.09	2	Volcanic eruption - Popocatepetl, Mexico	S, Pb, Cond.	[5]
1358.5	88.51	3	Volcanic eruption - ?	Pb, S	[5,6]
1391.5	86.71	3	Volcanic eruption - Hekla, Iceland	Pb, S	[5,6]
1442.5	84.19	3	Volcanic eruption - ?	S	[6]
1477.5	82.15	2	Volcanic eruption - Sangeang Api, Indonesia	Pb, S, Cond.	[5]
1502.5	80.85	3	Volcanic eruption - ?	S, Cond.	[5]
1512.5	80.05	3	Volcanic eruption - ?	S, Pb, Cond.	[5]
1537.5	78.43	3	Volcanic eruption - ?	S, Cond.	[5]
1554.5	76.85	3	Volcanic eruption - ?	S, Pb, Cond.	[5]

1569.5	75.73	3	Volcanic eruption - Tambora, Indonesia(?)	Pb, S	[5,6]
1585.5	74.31	3	Volcanic eruption – (?)	S, Pb, Cond.	[5]
1601.5	72.838	1	Volcanic eruption - Huaynaputina, Peru	S, Pb, Part., Cond.	[5]
1642.5	69.21	1	Volcanic eruption - Parker, Phillipines	S, Pb, Part., Cond.	[5]
1667.5	67.01	1	Volcanic eruption - Mt Tarumae, Japan	S, Pb, Part., Cond.	[5]
1696.5	64.35	1.5	Volcanic eruption - Sabancaya, Peru	S, Pb, Part., Cond.	[5]
1739.5	59.95	1	Volcanic eruption - Mt. Tarumae, Japan	S, Pb, Part., Cond.	[5]
1766.5	56.65	1	Volcanic eruption - Hekla, Iceland	S, Pb, Part., Cond.	[5]
1783.5	54.43	0.5	Volcanic eruption - Laki, Iceland	S, Pb, Ti, Part., Cond.	[5,7]
1816.5	48.634	1	Volcanic eruption - Tambora, Indonesia	S, Pb, Ti, Part., Cond.	[5,7]
1836.5	45.47	1	Volcanic eruption - Cosigüina, Nicaragua	S, Pb, Ti, Part., Cond.	[5,7]
1863.5	41.19	1	Volcanic eruption - Makian, Indonesia	S, Pb, Ti, Part., Cond.	[5,7]
1873.5	39.27	0.5	Volcanic eruption - Grímsvötn, Iceland	S, Pb, Ti, Part., Cond.	[5,7]
1884.5	37.01	1	Volcanic eruption - Krakatoa, Indonesia	S, Pb, Ti, Part., Cond.	[5,7]
1912.5	30.67	1	Volcanic eruption - Novarupta, AK	S, Pb, Ti, Part., Cond.	[5,7]
1955.5	19.95	1	Radiogenic (bomb horizon)	^{239}Pu	[8]
1962.5	17.81	1	Radiogenic (bomb horizon)	^{239}Pu	[8]
2015.42	0	0	Core top	N/a	~

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References

1. Shean, D. E. *et al.* An automated, open-source pipeline for mass production of digital elevation models (DEMs) from very-high-resolution commercial stereo satellite imagery. *ISPRS J. Photogramm. Remote Sens.* **116**, 101–117 (2016).
2. Cuffey, K. & Paterson, W. *The Physics of Glaciers, Fourth Edition*. (Academic Press, 2010).
3. Nye, J. F. Correction Factor for Accumulation Measured by the Thickness of the Annual Layers in an Ice Sheet. *J. Glaciol.* **4**, 785–788 (1963).
4. Dansgaard, W. & Johnsen, S. J. A Flow Model and a Time Scale for the Ice Core from Camp Century, Greenland. *J. Glaciol.* **8**, 215–223 (1969).
5. Sigl, M. *et al.* Timing and climate forcing of volcanic eruptions for the past 2,500 years. *Nature* **523**, 543 (2015).
6. McConnell, J. R. *et al.* Pervasive Arctic lead pollution suggests substantial growth in medieval silver production modulated by plague, climate, and conflict. *Proc. Natl. Acad. Sci.* **116**, 14910 LP – 14915 (2019).
7. McConnell, J. R. & Edwards, R. Coal burning leaves toxic heavy metal legacy in the Arctic. *Proc. Natl. Acad. Sci.* **105**, 12140 LP – 12144 (2008).
8. Arienzo, M. M. *et al.* A Method for Continuous 239Pu Determinations in Arctic and Antarctic Ice Cores. *Environ. Sci. Technol.* **50**, 7066–7073 (2016).