

Cooling and societal change during the Late Antique Little Ice Age from 536 to around 660 AD

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Radiocarbon dating and tree-ring detrending

Standard radiocarbon dating procedures (31, 32) using the compact MICADAS radiocarbon facility (33, 34), a tandem accelerator mass spectrometry (AMS), at ETH Zurich (35, 36), were applied to obtain all radiocarbon measurements. The inter-ring specific isotopic measurements reveal a rapid ^{14}C increase of 1.5% in atmospheric radiocarbon concentration from AD 774-775, as well as a >1.8% rise between AD 773 and 776 (Fig. S3). The annual-precise timing of the observed radiocarbon peak is within the 95% confidence interval and in agreement with independent high-resolution ^{14}C data from Austria, Germany, Japan and Switzerland (24, 37-40).

To maintain inter-annual (high-frequency) variability from the raw TRW measurement series, individual cubic smoothing splines with 50% frequency-response cutoff at 32 years were used (41). This detrending approach has been shown to robustly preserve growth extremes while eliminating background noise on inter-decadal and longer time-scales. Annual TRW indices were calculated as residuals after the application of a data adaptive power-transformation applied on the raw measurement series (42). A bi-weight robust mean was used to generate TRW chronologies of high-frequency variability (Fig. S4). To maintain low-frequency information on time-scales exceeding those of the individual series lengths (43),

the Regional Curve Standardization (RCS) method was independently applied on the living and relict subsets (5, 44). The resulting (horizontal) RCS chronologies (Fig. S2E) were annually weighted by their sample replication and averaged (5). Uncertainty in the Altai composite is strongly inflated before ~600 CE and again in the 20th century (Fig. S2E), when the mean TRW chronology tends to lie near the lower error bar. The recent discovery of archaeologically excavated funerary structures and fence poles from the Scythian and Türk empires (45), as well as subfossil wood from glacier retreats (46), suggests the potential for improving sample replication back in time. The even distribution of an overall high sample replication suggests that the new Altai TRW chronology constitutes a central Asian equivalent to the dendroclimatological evidence from the European Alps (5). Note that even though the horizontal split approach may reduce some longer-term biases in the chronology, it can possibly also diminish lower frequency information from the shorter subsets (43). This approach was, however, also applied for the Alpine dataset (5), thus allowing straightforward comparison of the two TRW chronologies (Fig. 1A).

Climate response patterns

To avoid regression-based variance reduction in the proxy reconstruction model (47), the new Altai TRW RCS larch chronology was scaled (1900-2003) against JJA temperature means averaged over the 45-50°N and 85-90°E region in central Asia (48). This ‘composite plus scaling’ procedure, i.e. the adjustment of proxy mean and variance, is the simplest amongst various calibration techniques and least prone to variance underestimation (5, 47, 49). A slightly higher first-order autocorrelation coefficient of the Alpine climate (0.35 vs. 0.26) reflects the ‘redder’ spectra of a more ‘maritime’ vs. ‘continental’ climate regime (50, 51). Both the Alpine and Altai TRW records are positively correlated with 20th century JJA temperatures (Fig. S7A-B), possessing significant year-to-year and multi-decadal agreement ($r = 0.74$ and 0.49 ; 1900-2003). The year-to-year and multi-decadal proxy-climate agreement

($r = 0.74$ and 0.63) is particularly strong during the second half of the 20th century (1950-2003), when the quality and quantity of the instrumental measurements across central Asia have improved (52). The overall better fit of the Alpine data is further expressed by trend-free annual residuals between the scaled TRW indices and the JJA temperatures (Fig. S7C-D). Low-frequency summer temperature variability is well captured by the TRW chronologies from both regions ($r = 0.97$ and 0.90 after 20yr low-pass filtering; 1950-2003). The high- to low-frequency agreement between the TRW proxy and instrumental JJA target time-series refutes any possible ‘divergence issue’ (53). Our results are also in line with previous findings from the Russian Altai that emphasized dendroclimatological multi-parameter approaches to reconstruct regional climate variability (54, 55). Since TRW-based reconstructions might overestimate abrupt climate perturbations with increased thermal inertia, due to inflated biological memory (56, 57), this study, among many others, underlines the importance of alternative dendrochronological and wood anatomical parameters to better assess the frequency and severity of past temperature extremes (57), thus stressing the need for follow-up projects.

Positive feedback loops

Distinct cooling of the LALIA occurred during a period of relatively cooler mean summer temperatures compared with the warmer (previous) Roman and (following) medieval times. In this regard, it is important to note that the persistence of the LALIA was not solely forced by the direct thermal shock following the extreme volcanic eruptions (14), but prominently amplified and sustained by internal climatic processes including ocean – sea ice and ocean-atmosphere feedbacks (58, 59). For instance, (58, 59) investigated the ocean response in the aftermath of strong volcanic eruptions in climate model simulations covering the last pre-industrial millennium (850-1849 CE). They identified a lagged effect that includes changes in the North Atlantic Subpolar Gyre and the Atlantic Meridional Overturning Circulation

(AMOC) and associated ocean heat transports. The strength and duration of volcanically forced feedbacks depended on the background state. Different model realizations can produce different responses in the climate trajectory. It should be speculated, therefore, that the real climate might also sometimes be reflected by the extreme trajectories. In any case, the ensemble mean of the cumulative effect of subsequent volcanic eruptions leads to long-lasting cooling. The cumulative effect of decadal paced volcanic eruptions was also explored (60). Based on a series of model simulations, it became evident that a mechanism combining cooling of the mixed ocean layer and a positive response of Arctic sea-ice induces a positive ice-albedo feedback. In some simulations the cold state lasts for more than 100 years (60). The cumulative effect of volcanic eruptions was also reported from a series of papers addressing empirical climate reconstructions (15, 16), suggesting that such a clustering of strong volcanic eruptions may be partly responsible for the onset of the Little Ice Age in the pre-industrial millennium.

Changes in human history

The nomadic Rouran, the people who may have come to be known as the Avars in the west (8), had effectively held sway in Mongolia since the late-4th century, but submitted to the Northern Wei Dynasty (386-534 CE), then ruling northern China, in 521. In 523 a widespread famine is reported (61), followed by frequent border crossings and renewed warfare ending with a peace agreement according to which the Rouran assisted the Northern Wei in quelling a series of rebellions in exchange for rewards. From 528 the Northern Wei dynasty entered a period of political crisis that ended in 534 with its collapse and replacement by a series of short-lived dynasties.

According to the Chinese records, after 552 the Türks revolted against the Rouran (62, 63), and replaced them as the major steppe power in Mongolia, as well as in other parts of central Asia (64). A portion of the Rouran presumably migrated westward (65). Between 562

and 567 CE the Türk kaghan Istāmi and his followers began their expansion westward from his base in the Altai (66). Before 568, the Türks entered western Eurasia, defeated the Avars and the Hephthalite Huns, and took control of the trading towns of the Sogdians, in whose interests they opened diplomatic exchanges with the Persian Empire in order to sell central Asian silk to the Persians (67 and below). In 582 the leadership of the Türk Empire split into two branches, a western and an eastern one (68, 69). The turmoil in the steppe region, including the weakening of the Rouran, their defeat and collapse at the hands of the Türks, and the Türks' expansion and later division occurred at a time of extremely low summer temperatures (Fig. 2, 3). This may explain the pressure posed by the Türks on northern China and the agreement with the northern Qi to provide large quantities of food and other goods to the Türks (70).

According to Roman sources, at least two new groups appeared on the northeastern frontiers of the empire after the middle of the 6th century. Probably in January 558, nomads whom the Romans call Avars, and who are often identified with the Rouran, sent an embassy to the imperial capital in Constantinople. They had fled their own country and come to the frontier provinces of the empire: John Malalas, *Chronographia*, 18, 125, ed. J. Thurn (Berlin: Walter de Gruyter 2000), 420.67-8, and Theophanes, *Chronographia*, A.M. 6050, ed. C. de Boor (Leipzig: Teubner, 1883-5) 1.232.6-10, clearly a more complete version of the same text; cf. E. Stein, *Histoire du Bas-Empire* (Paris: Desclée de Brouwer, 1949-59), 2.542.

Less than a decade later, a new group, identified as *Tourkoi* (Türks?) entered western Eurasia. Rebuffed by the Persians, they turned to the Romans, dispatching an initial embassy that reached Constantinople in late 568 or early 569: Menander Protector, Fg. 10, 1, ed. Blockley, *The history of Menander the Guardsman* (Liverpool: F. Cairns), 110.1-116.95.

The political upheaval continued in northern China until the unification of China under the Sui dynasty (581-618 CE) in 589. Throughout the 6th century the written records mention frequent occurrences of droughts, sometimes for long periods of time (71). In 618 this

dynasty, itself instable, was replaced by the Tang (618-907 CE). The latter had to contend with the eastern Türk Empire which had since gained dominance in Mongolia and adjacent regions and multiplied attacks on north China from 620 to 626 until it was finally conquered by the Tang in 630 (72). In 625, the contact between east Roman emperor Heraclius and a third newly arrived Asian group is documented by the two best Greek historians for this period: Nicephorus, *Chronographia*, 12 and 18, ed. C. Mango (Washington, D.C.: Dumbarton Oaks, 1990) 54.16-58.64 and 66.4-8; Theophanes, *Chronographia*, A.M. 6118, 1.315.14-16 and 315.26-316.16, who designates this group as “the Türks from the East” (tous Tourkous ek tēs heōas).

Overcoming environmental determinism and reductionism

Temperature and hydroclimatic fluctuations of interannual to multi-decadal time-scales that occurred frequently during the Holocene have been relatively small compared to those of the Pleistocene. Their amplitude, however, could still have had positive or negative effects on the agricultural carrying capacity, economy and societal organization of numerous historical civilizations. Changing climate conditions may also have triggered migrations over both shorter and longer distances. The very same climate fluctuation (e.g. warming or cooling) can have different local to regional impacts, not least by increasing the precipitation in some areas and decreasing it in others, and thus resulting, for instance, in larger agricultural productivity in some places and harvest failures in others.

The rapidly accumulating body of new, more highly resolved and robust paleoclimatic evidence is creating new opportunities to explore the environmental component of major changes in human society, both in how such changes may have affected a given society, and how that society responded to such changes with greater or lesser success (73). Archaeologists working on ancient civilizations have increasingly recognized that climate changes can have significant socio-economic consequences (74). Although historians, mainly

158 studying societies in the last millennia, have in general been more hesitant to invoke climate
159 to partly explain broader societal changes, the question is presently under investigation (75).
160 Recent studies have claimed or proven far-reaching effects of Little Ice Age (LIA) type events
161 on societies across the globe, and begun to unveil the mechanisms for such impacts, even as
162 they underscore how differing types and degrees of social resilience buffer or exacerbate the
163 responses of different societies to similar climate challenges (76, 77).

164 Even the most skeptical historians acknowledge that year-to-year weather variations and
165 extreme weather events may have impacted the well being of pre-modern societies, while the
166 arguments of those who reject, e.g. climate conditions connected with the LIA as an important
167 factor in specific historical developments (78), have encountered robust criticism on both
168 scientific and historical grounds (79). Scientifically and historically grounded climate
169 reconstructions exploring potential linkages between climate events and human social
170 responses have begun to multiply, even for the first millennium CE (80-82).

171 It is essential to underscore that, to date, climatic changes, be they temperature,
172 precipitation or some combination of these and other components, have rarely been shown to
173 be the only or primary cause of societal change. In most cases, climate can arguably be
174 considered one factor of greater or lesser relevance, depending on the amplitude of the
175 change, its timing, and the resilience of the society which must respond to that change.
176 Correlating in time and space paleoclimatic reconstructions with different historical
177 phenomena identified from the written or archaeological record is only the first step toward
178 identifying what causal relationships may have existed between climate variability and human
179 history. What types of change – swift, slow, clustered, etc. – and which mechanisms
180 contributed most to, e.g. increasing or decreasing food security and societal ability to respond
181 to change by capitalizing on new opportunities, or overcoming unforeseen challenges? These
182 are questions with which historians, archaeologists, and climate scientists have only begun to
183 grapple.

184 Climatic changes have had – and perhaps still have – their most obvious impacts near the
185 limits, or transition zones, of different climate and vegetation zones. The regions near the
186 thermal or hydrological limit for a certain kind of agricultural or pastoral activity are most
187 sensitive to environmental changes because warmer/colder or dryer/wetter conditions can
188 more easily displace these limits. In regions where it was already difficult to conduct a certain
189 type of agriculture or pastoralism, for example, at certain altitudes, they could, in the worst
190 case, become nearly impossible within a slightly colder or dryer climate. Conversely, climate
191 change may push the agricultural or pastoral limits forward into new regions that under
192 previous climate circumstances were unsuitable.

193 Climatic changes affected the subsistence potential for ancient populations of hunters and
194 gatherers. When changes in the climate altered the ecosystem or biomass productivity this
195 either increased or decreased the availability of game animals or triggered their migration.
196 The populations of hunters and gatherers were then obliged to follow the movements of the
197 game or move to new areas with better hunting grounds. The introduction of agriculture and
198 permanent settlements more or less precluded this flexible form of adaptation to the
199 ecological effects of climate changes. Since most pre-modern societies enjoyed small margins
200 or surplus of food, even rather modest climate-driven increases or decreases in agricultural
201 yields could have had rather large positive or negative impacts. Depending on resilience, in
202 this case, the availability of alternative food sources, food shortages might occur even with a
203 relatively modest reduction of the yield (83, 84).

204 The degree of marginality, in the context of agriculture, can be conceptualised as the risk
205 or frequency of harvest failures. Climate change may either increase or decrease this risk in a
206 given region. Clusters of years with bad harvests or outright harvest failures were particularly
207 problematic for food security in pre-modern societies and could easily cause subsistence
208 crises. Reduced agricultural productivity resulting in famines could have demographic
209 consequences, especially among the poorer strata of the population, by increasing mortality

210 and reducing fertility (85). Poorer nutrition moreover resulted in an increased vulnerability to
211 many diseases, and hence the rise of epidemics, significantly escalating mortality (86).
212 Societal elites who might have been able to avoid directly suffering from food shortages
213 nevertheless may have found their status challenged due to smaller surplus wealth for
214 maintaining elite status and functions. Failure to deliver relief during food crises may have
215 decreased their popularity and authority. Attempts to claim the same revenues from the
216 population despite reduced agricultural productivity and dearth – or after a crisis to increase
217 claims to compensate for population/productivity loss – could result in social unrest with risk
218 even of open rebellion or civil war (87).

219 Reactions to adverse effects of climate change, and which (if any) adaptation strategies
220 might be attempted, have likely varied from society to society throughout history. Societies
221 with a high degree of resilience, for instance those which deploy a wide range of subsistence
222 strategies or feature the most developed and robust economies seem, in general, to have
223 shown the greatest resilience to the effects of climate changes and other environmental
224 impacts. Alternative sources of subsistence gave such societies a “safety net” when some of
225 their ordinary means of subsistence faltered (88).

226 Climate-induced biomass shortfalls have sometimes explicitly resulted in forced
227 migrations or triggered population movements to more favourable locations, for instance the
228 migration of Arab pastoralists recorded on the margins of the Roman Empire in 536 (89).
229 Climate change might also, in some cases, have fostered migrations in order to take advantage
230 of opportunities offered by improved climate conditions in another region, as for instance may
231 have occurred during the Norse colonization of Greenland (90). Migrations involving
232 nomadic or semi-nomadic populations could not infrequently have expanded considerably the
233 scope of such groups’ ordinary movements. Climate-induced or climate reinforced
234 subsistence crises involving farming populations have more often resulted in the
235 abandonment of farmsteads and in the migration to wealthier areas near or far, and even

overseas (91). Societal constraints on humans freely adapting to climatic changes by migrating include the division of land through political borders and property rights (e.g. to agricultural lands), the disrespect of which could easily trigger instability and conflict with those already claiming a territory and its resources (92).

Figures S1 to S15

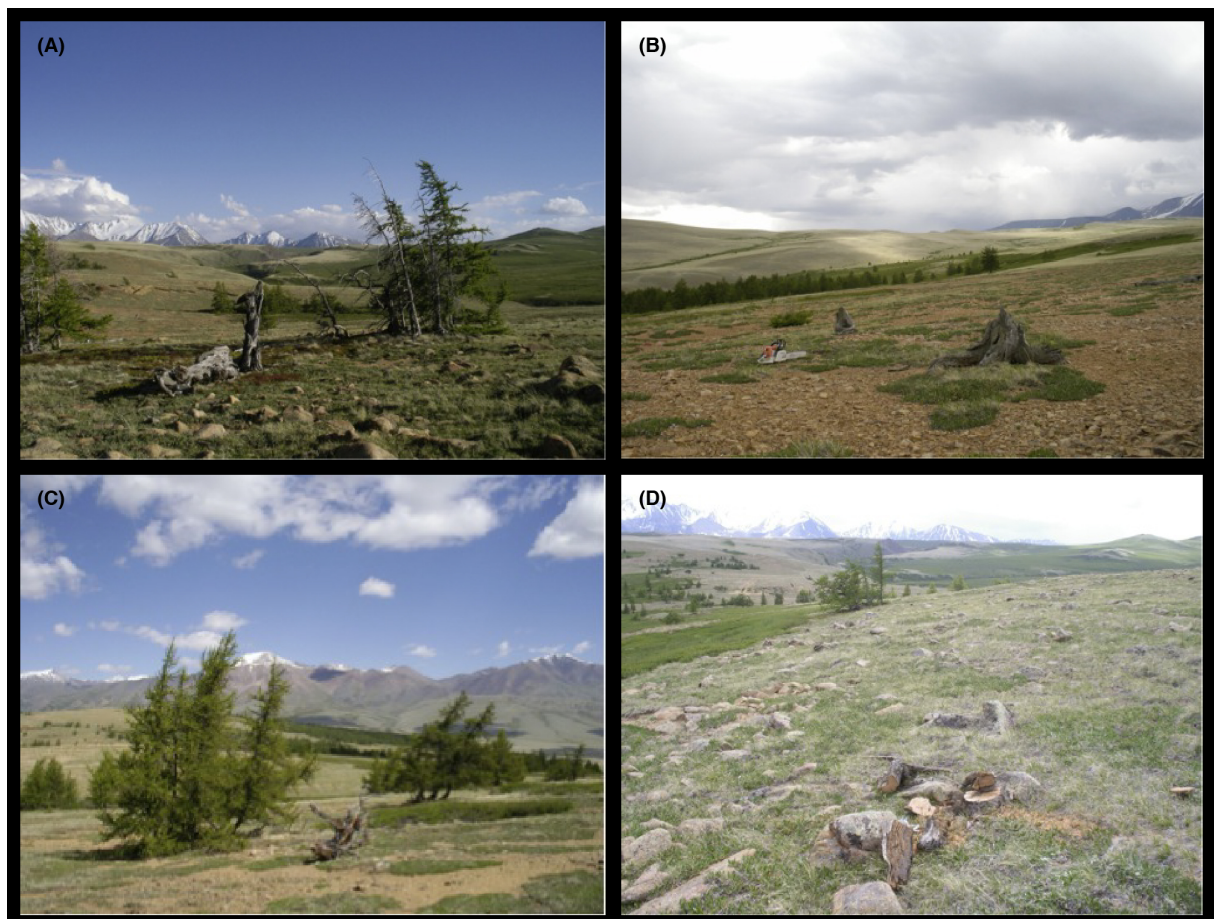


Fig. S1. (A-D) Examples of living and *in situ* relict larch (*Larix sibirica* Ldb.) trees that were sampled within the Mongun (MG) treeline ecotone (between 2100 and 2500 m asl) in the Russian Altai Mountains (see Table S2 for details). This dataset enabled the development of an annually resolved and absolutely dated TRW chronology that continuously covers the Common Era and reflects a distinct summer temperature signal (photos: V. S. Myglan).

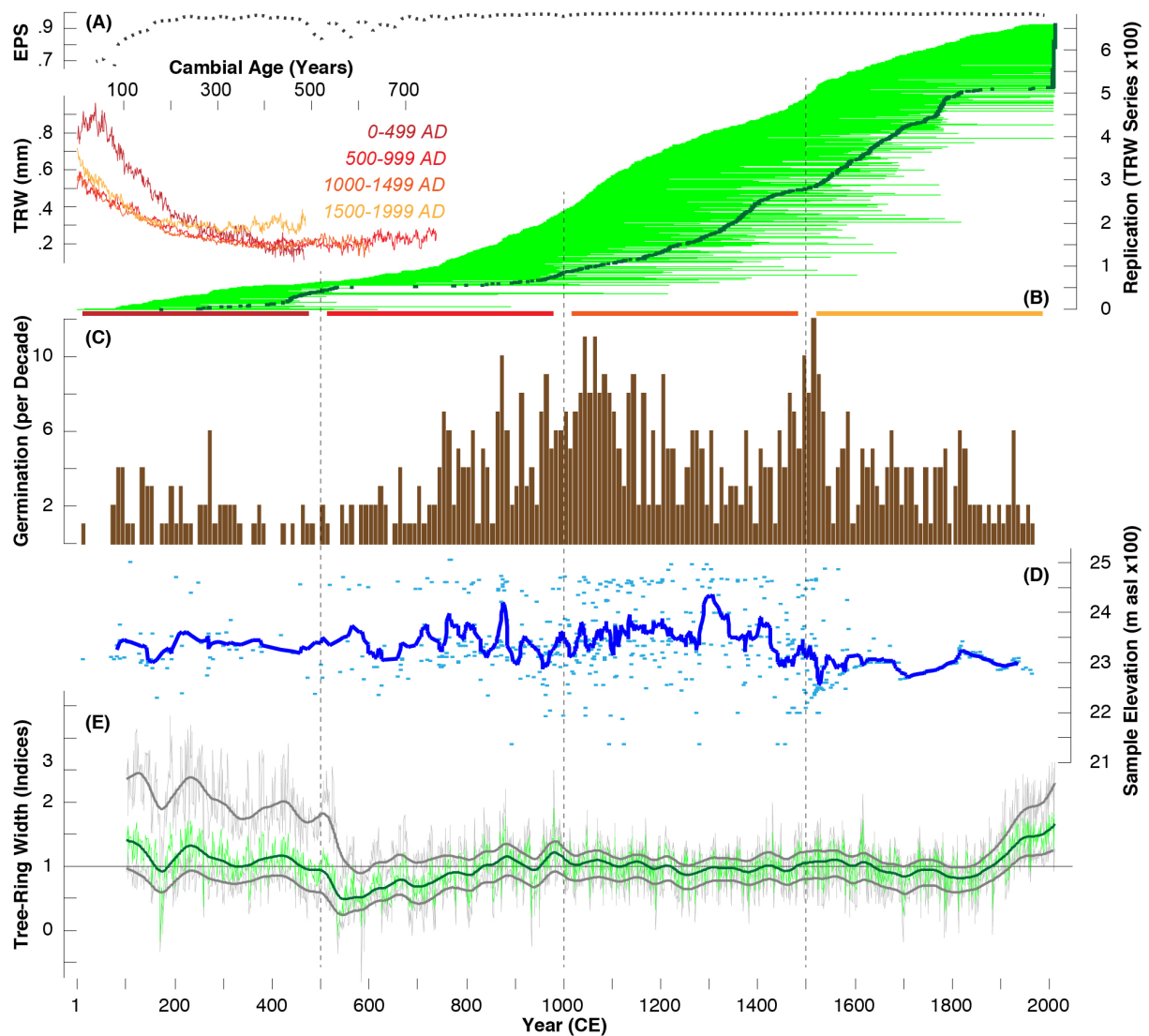


Fig. S2. (A) Growth coherency in 30-year moving and 15-year lagged values of the Expresses Population Signal (EPS) (93) using (B) the 152 living and 508 relict larch (*Larix sibirica* Ldb.) TRW samples (Fig. S1; Table S2). This composite dataset continuously covers the period 104-2011 CE with a replication of >10 samples per year (green bars represent the individual measurement series sorted by the year of their first ring, whereas the dark dots refer to sample distribution when sorted by end dates). Left side inset shows four Regional Curves (RCs) (44), which were independently computed over four different periods (dark red to light orange bars at the bottom). (C) Decadal-resolved sums of estimated larch germination dates (vertical brown bars), (D) sample elevation of each larch specimen (in m asl) together

with the corresponding 11-year running mean values (blue line), and (E) the final larch TRW chronology (using weighted RCS chronologies after power-transformation (PT) (42) and horizontal data splitting in two subsets from 0-499 and 500-1999 CE), together with its 80-year low-pass filter (dark green). The light (annual) and dark grey (filtered) lines refer to the accumulated chronology uncertainty (94), which includes sampling, detrending and bootstrapping error (95, 96).

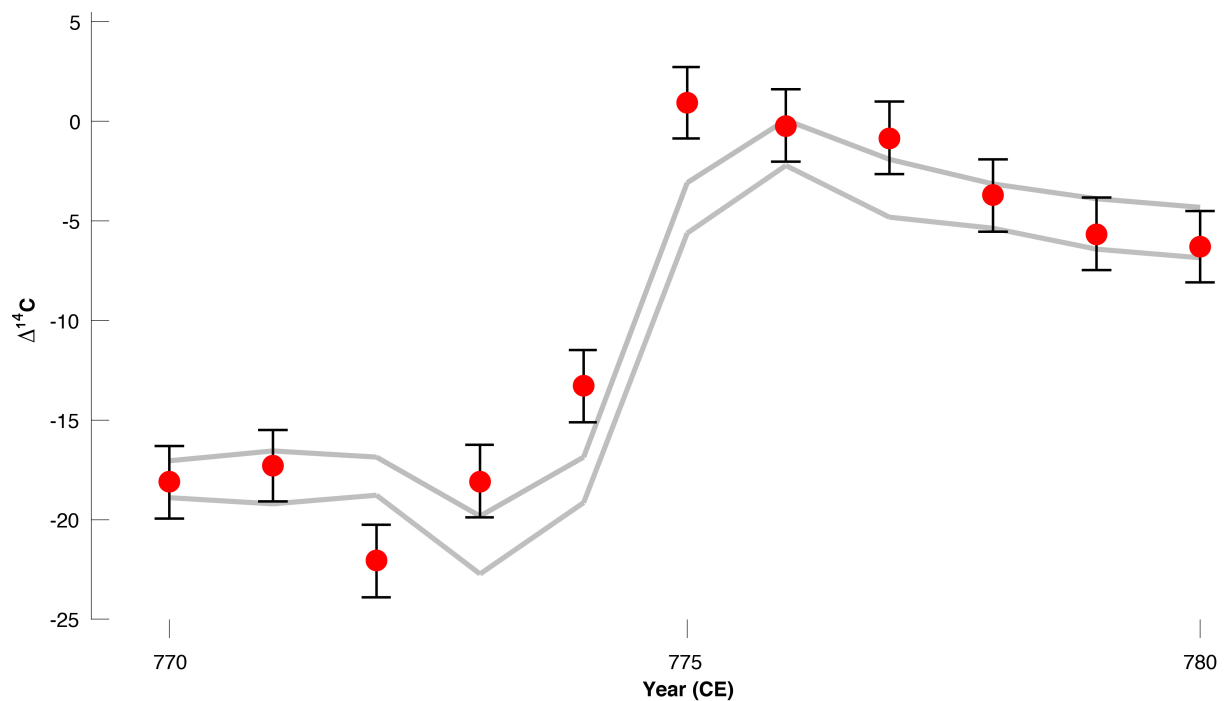


Fig. S3. Annual-resolved radiocarbon values (red) derived from one relict larch sample (#C150525DG1) from the new Altai TRW dataset, superimposed on the standard error from a set of independent, high-resolution radiocarbon measurements from different sites across the globe IntCal12 (grey; ± 1 standard deviation). The ^{14}C concentration is expressed as $\Delta^{14}\text{C}$, which is the deviation (in %) from the $^{14}\text{C}/^{12}\text{C}$ ratio of the sample with respect to modern carbon (standard sample), after correcting for the age and isotopic fractionation (24). The specific $\Delta^{14}\text{C}$ values between 770 and 780 CE are range from -22.1 to 0.9 (-18.1, -17.3, -22.1, -18.1, -13.3, 0.9, -0.2, -0.8, -3.7, -5.7, and -6.3 associated with a standard error of ± 1.8).

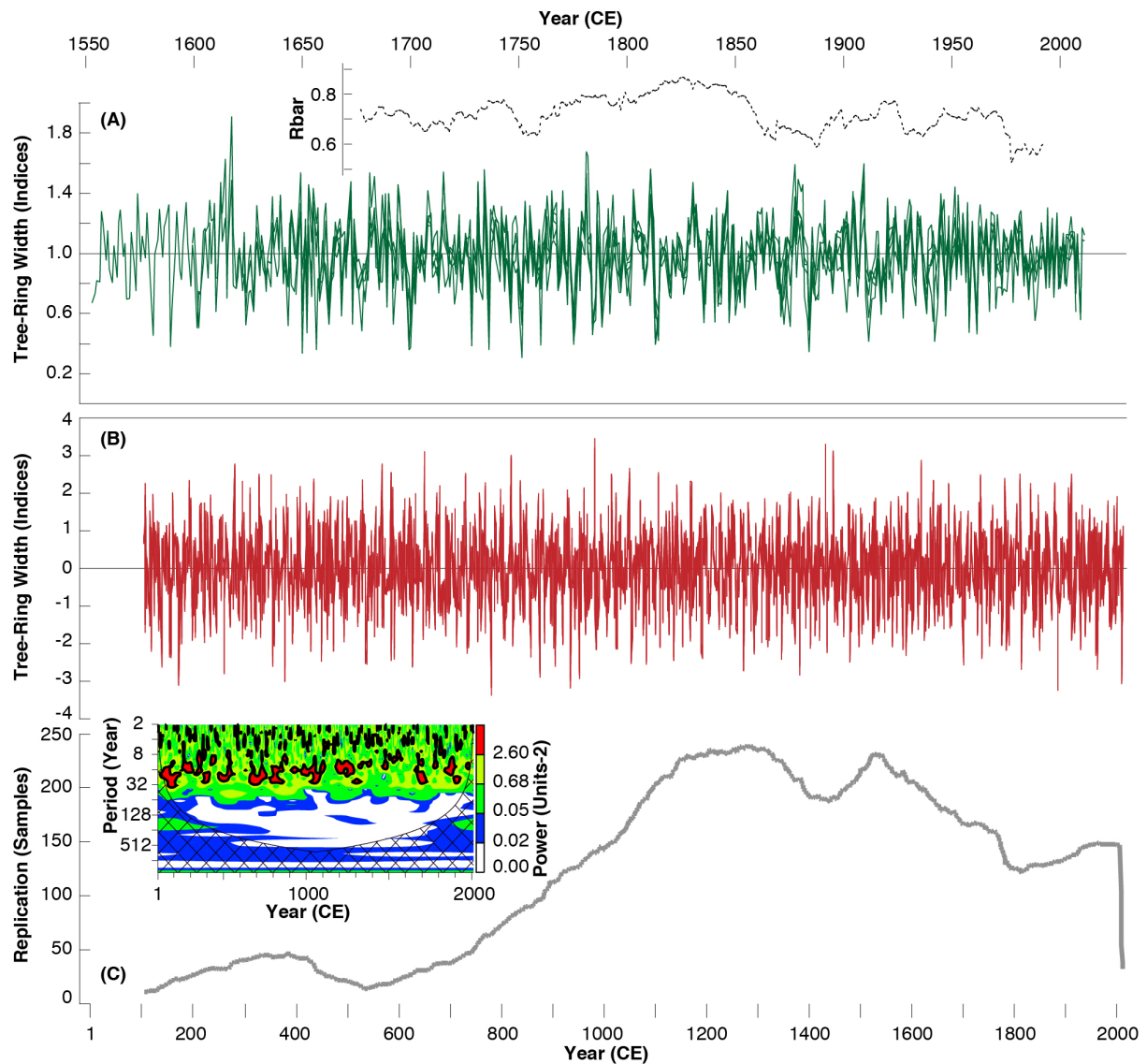


Fig. S4. (A) Comparison between the five larch TRW site chronologies from the Russian Altai Mountains (Table S2), which were calculated using power transformation and individual cubic smoothing splines, with 50% frequency cutoff at 32 years (41), together with their corresponding inter-series correlation coefficients (R_{bar}), using 30-year windows lagged by 15 years. (B) The high frequency (32-year spline PT) composite chronology based on all TRW measurement series and after using additional variance stabilization (97). Inset refers to wavelet power spectrum of the high-frequency chronology, with contour levels chosen so that 75%, 50%, 25%, and 5% of the wavelet power is above each level, respectively. The crosshatched region is the cone of influence, where zero padding has reduced the variance. Black

contour is the 10% significance level, using a white-noise background spectrum (98).
 (C) Temporal changes in sample replication range from 11 in 104 CE to 240 series in 2011 CE.

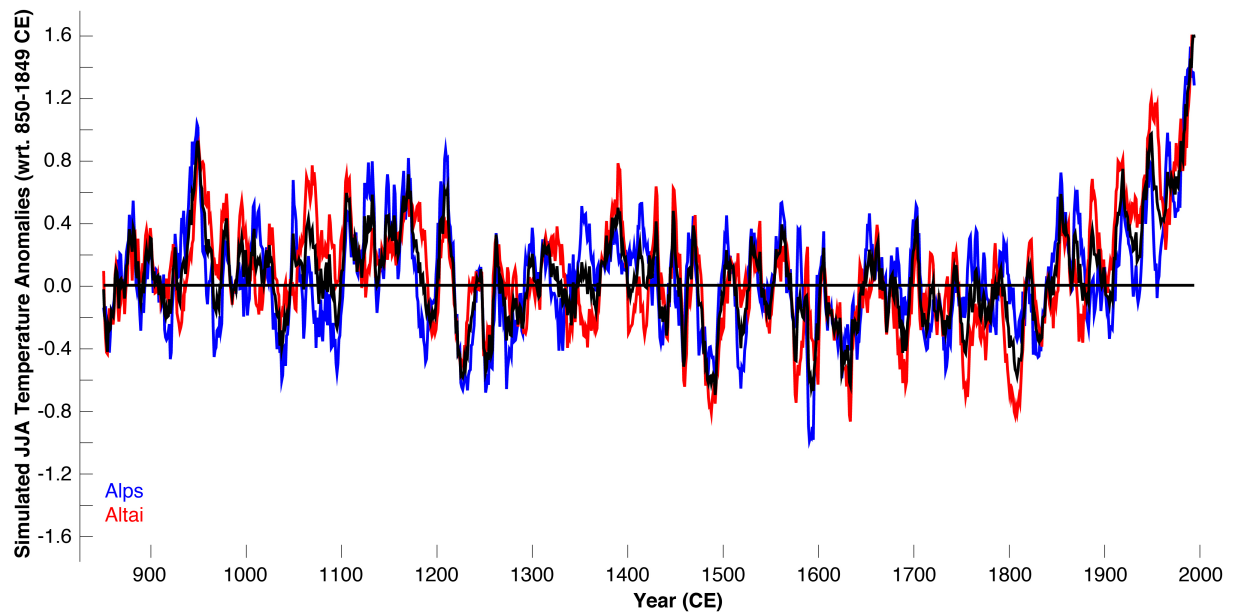


Fig. S5. Simulated temperature anomalies for central Europe (blue) and central Asia (red) with respect to the pre-industrial mean (850-1849 CE). The black line is the combined average of the two records that was also used for spatial field correlations (Fig. S5). All time-series were smoothed with an 11-year running mean. Coherency between the regional temperature evolutions over the last millennium was derived from climate model simulations, i.e. the Max Planck Institute Earth System Model (MPI-ESM-P) “past1000-r2” (27). Data for central Europe and central Asia spatially represent the Alps and the Altai, respectively, and were averaged over 10-15°E and 45-50°N, as well as 85-90°E, 45-50°N (48).

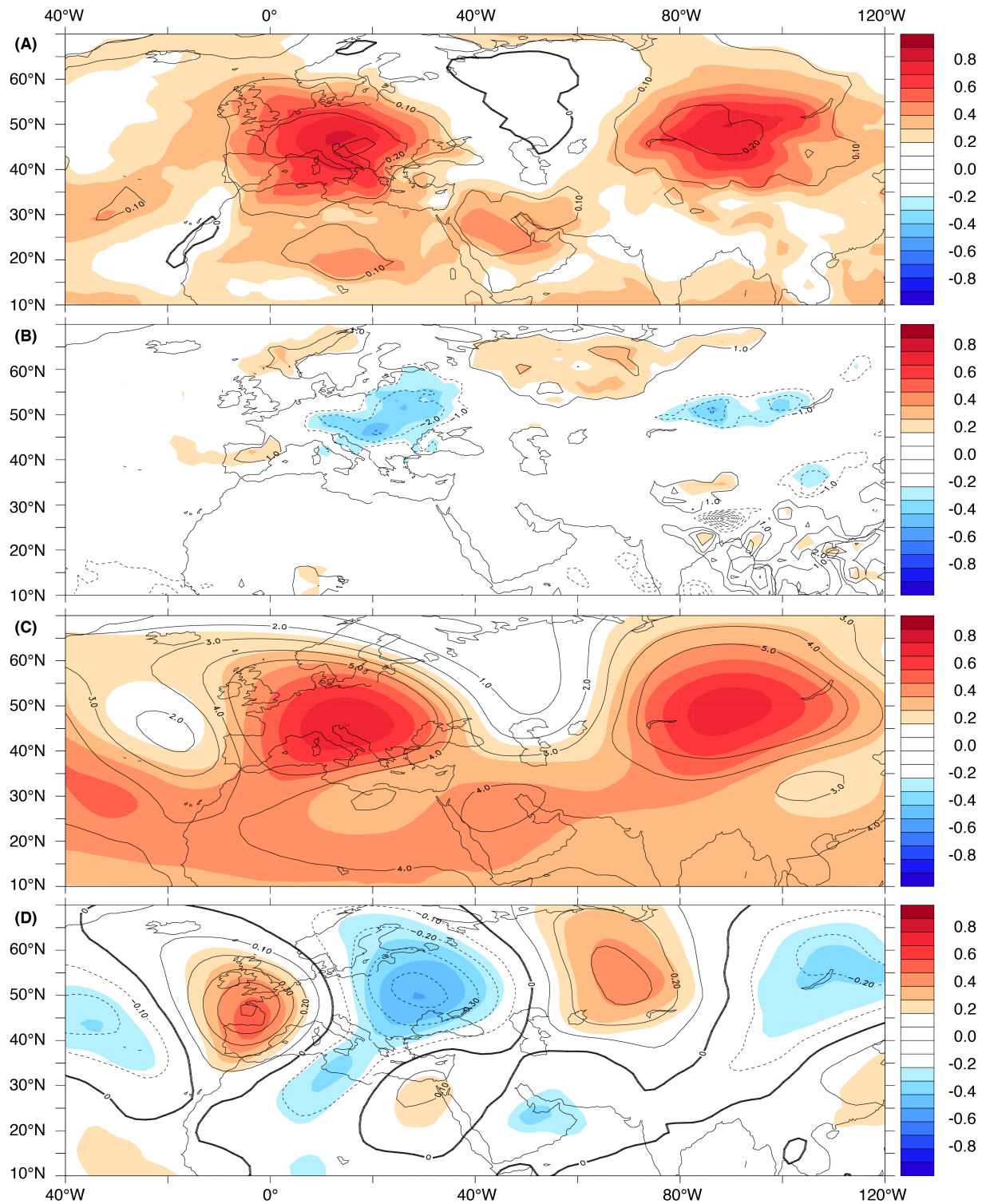


Fig. S6. Spatial field correlation (color shading) of the combined Alpine/Altai summer temperatures from last millennium simulations (Fig. S4) against: **(A)** summer surface air temperature, **(B)** total summer precipitation, **(C)** summer 200 hPa geopotential-height, and **(D)** summer 200 hPa meridional velocity. Contours depict the respective regression coefficients in **(A)** K, **(B)** mm/month, **(C)** m, and **(D)** m/s per standard

deviation of the combined temperature time series. Correlations between the combined Alps and Altai time series and selected spatial fields indicate that the covarying temperature field is most likely modulated by a wave train-like pattern in the upper troposphere (26, 99). The particular pattern in 200hPa geopotential height or meridional velocity that is responsible for the surface temperature coherence resembles the circumglobal wave train, which has been identified to modulate European and Eurasian summer climate. Warm summers are associated with regional high-pressure situations and (slightly) lower-than-normal rainfall.

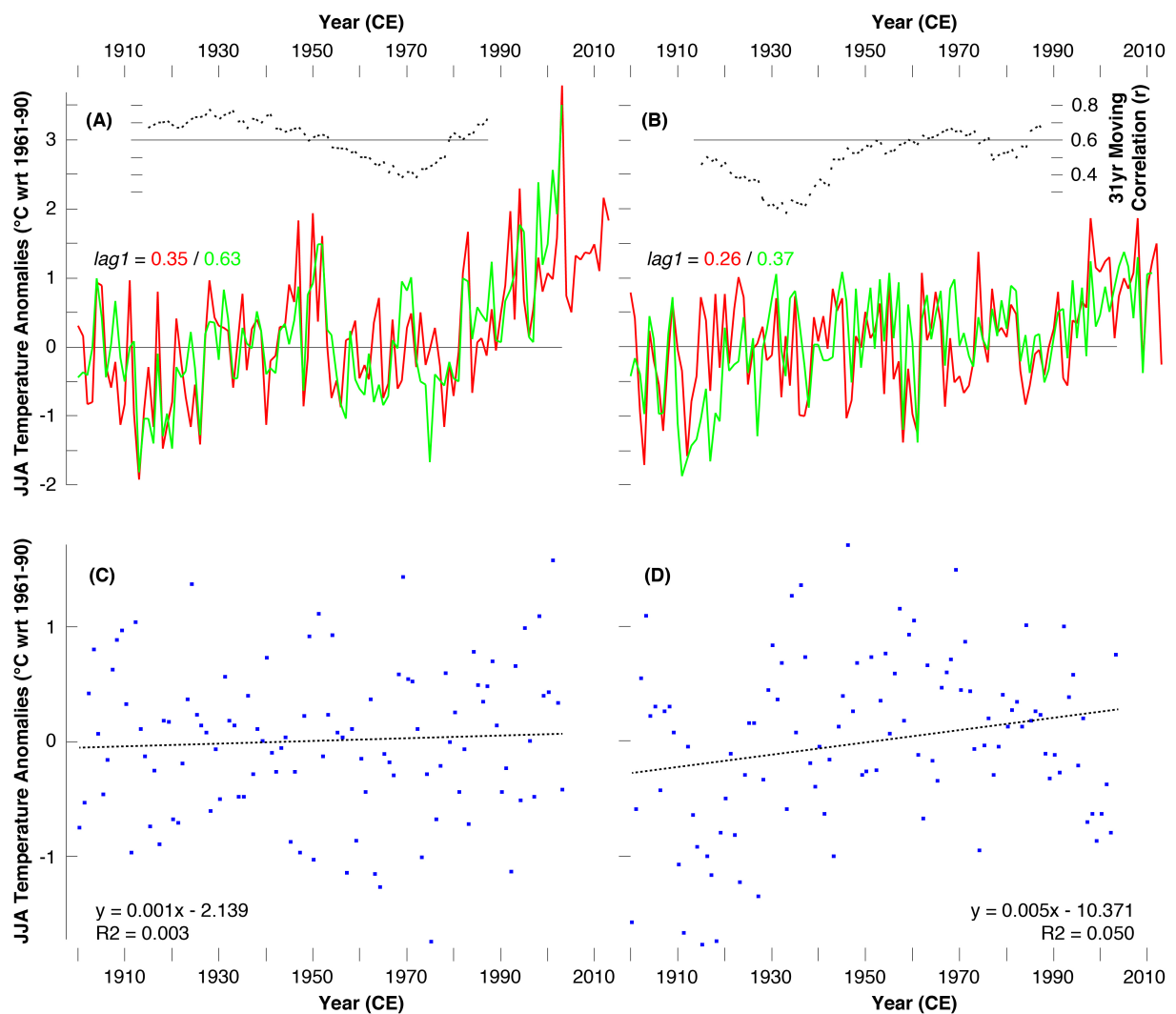


Fig. S7. Measured (red) and reconstructed (green) JJA temperature anomalies from (A) the European Alps (5) and (B) the Russian Altai Mountains, together with 31-year

moving correlations (black dots). Proxy data were scaled against gridded summer temperatures over the 1900-2003 common period of overlap (Berkeley update from 2015; 48), i.e. variance and mean of the RCS chronologies were fitted to the regional JJA temperature anomalies (47). (C-D) Annual residuals between the scaled RCS TRW chronologies from (A) the Alps and (B) the Altai, calculated against their corresponding gridded targets in central Europe and central Asia, with the dashed grey lines indicating linear trends between 1900 and 2003.

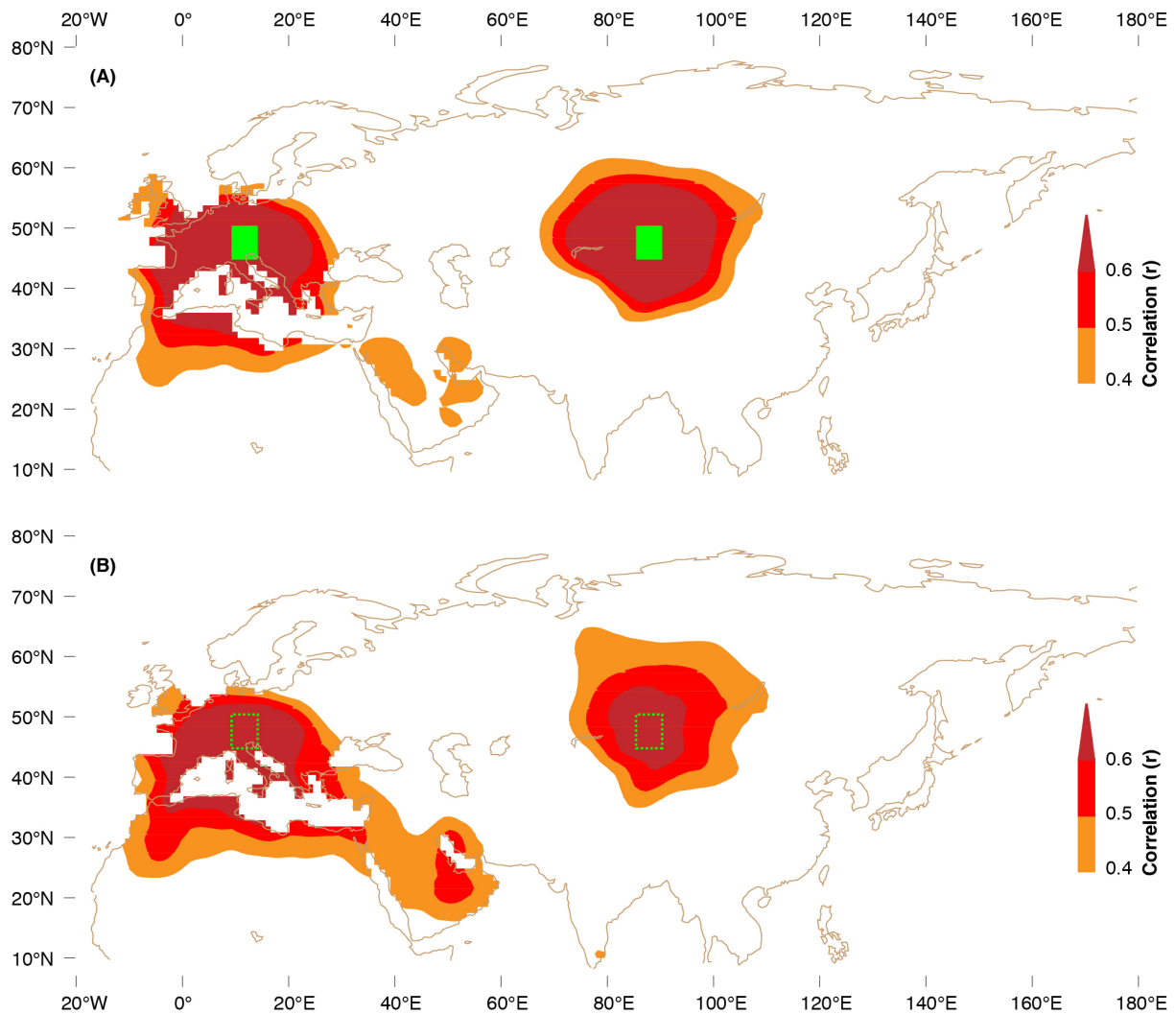


Fig. S8. (A) Spatial field correlations (1900-2003) of high-resolution (1°×1°) gridded JJA temperatures (Berkeley update from 2015; 48) averaged over the greater Alpine (45-50°N and 10-15°E) and Altai (45-50°N and 85-90°E) regions against the same

dataset over Eurasia (10-80°N and 20°W to 180°E), i.e. target versus target (see black inset map in Fig. 1). **(B)** The same as described above but using a combined mean record of the Alpine and Altai RCS chronologies for spatial field correlations at the Eurasian-scale.

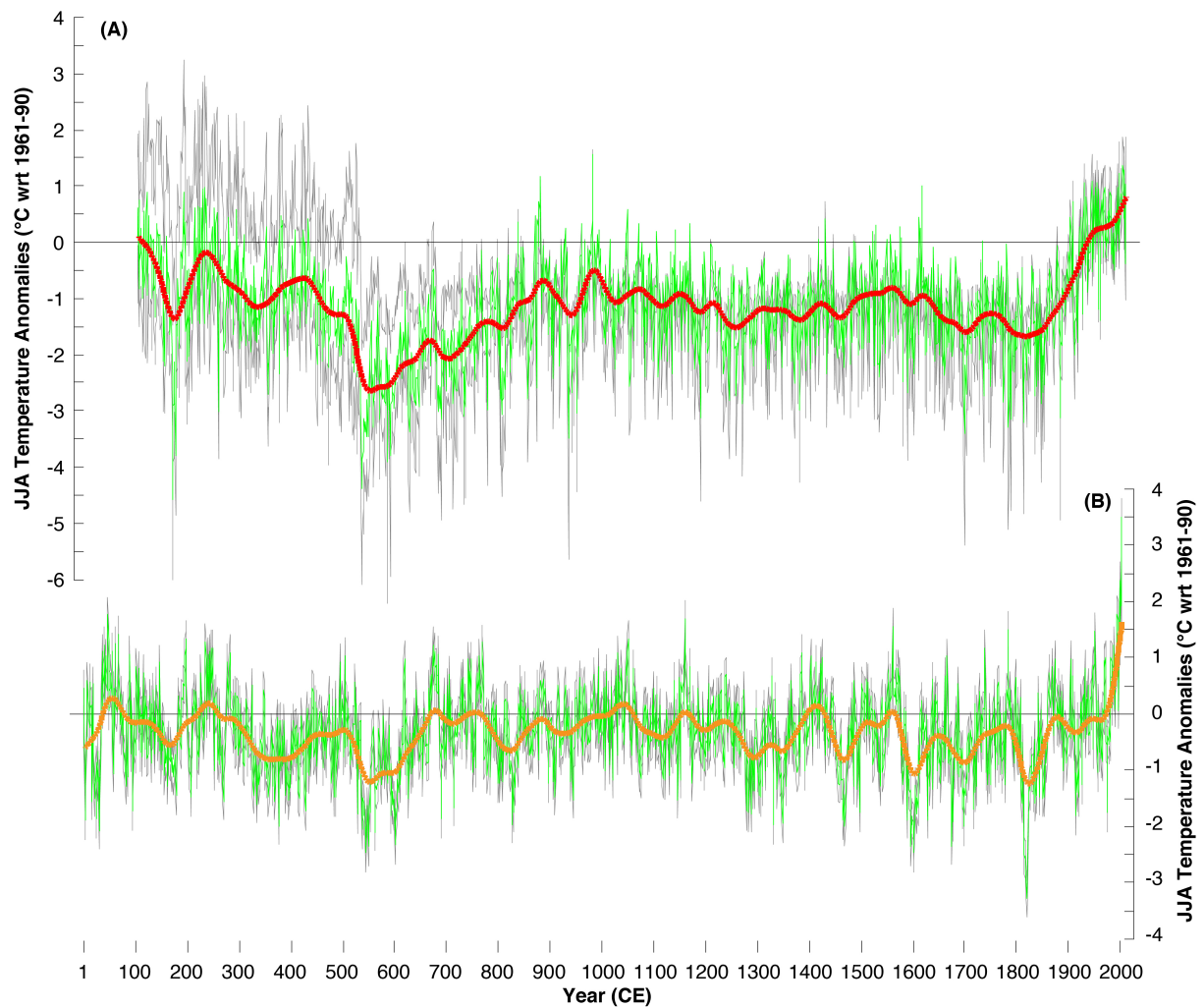


Fig. S9. (A) Annual resolved uncertainty ranges (grey) of the **(A)** Altai (94-96) and **(B)** Alpine summer temperature reconstructions (light green) (5), with the red and orange curves indicating 80-year low-pass filters of the main reconstructions.

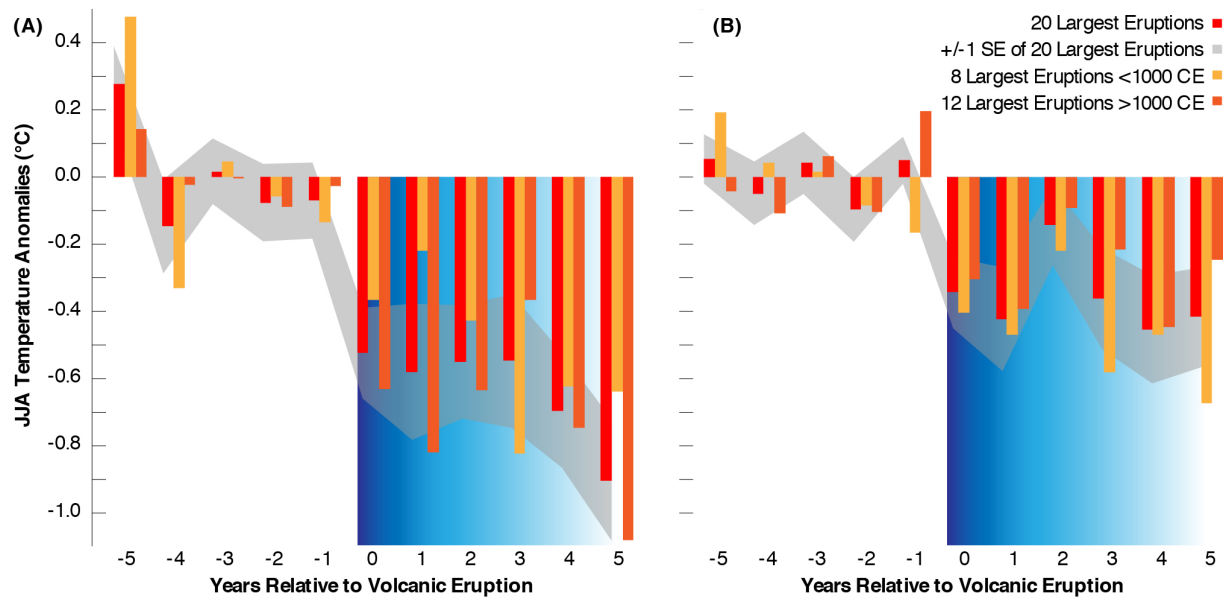


Fig. S10. Superposed epoch analysis (SEA) for the 20 largest volcanic eruptions that occurred during the Common Era (see Table S4) (14), using the summer temperature reconstructions from (A) the Russian Altai (this study) and (B) the European Alps (5). All time-series were normalized over five years before the eruptions and the grey background shadings are two standard errors.

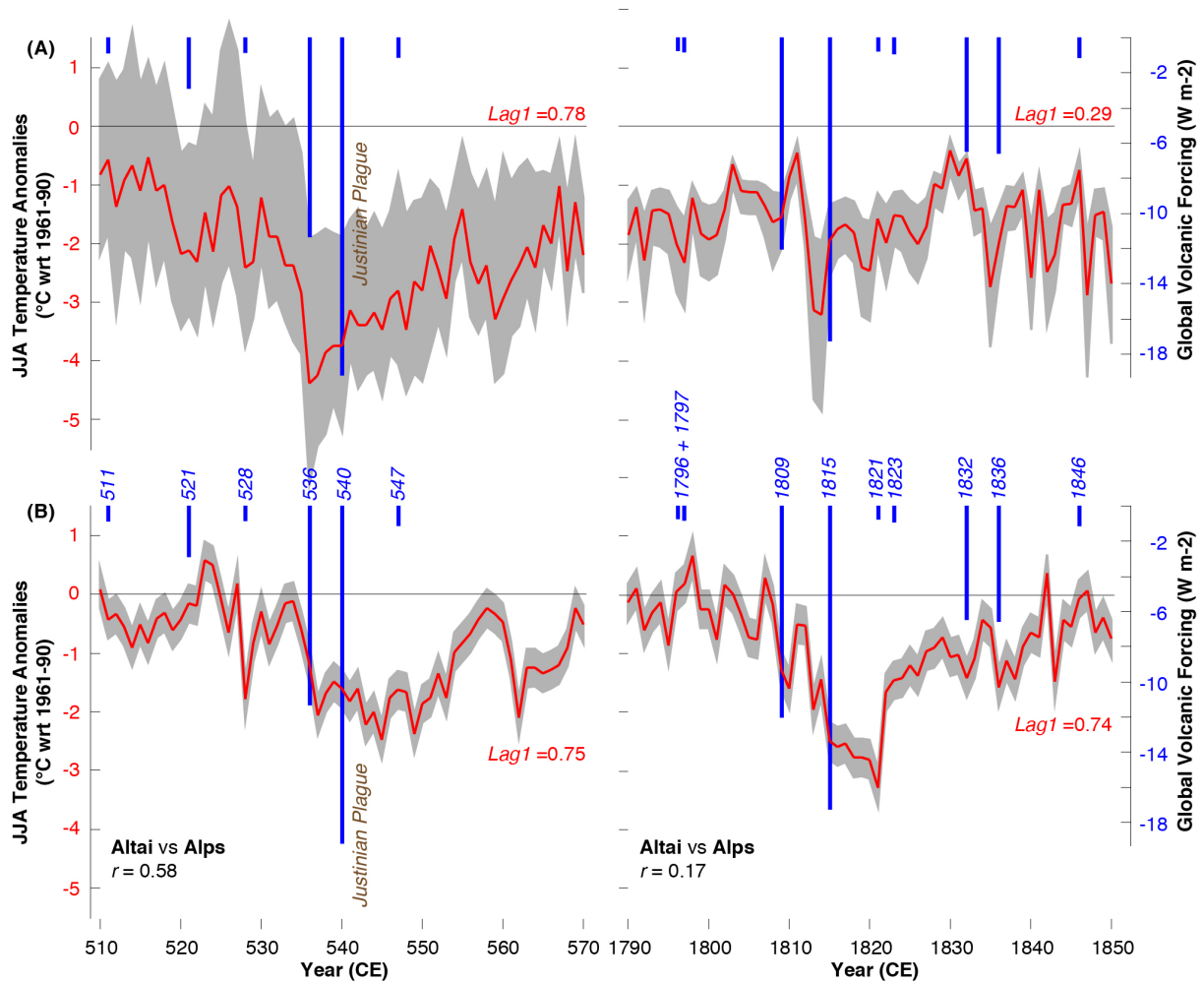


Fig. S11. Comparison between the volcanic-induced mid-6th century and the early-19th century summer cooling in (A) the new Altai and (B) the existing Alpine temperature reconstructions (red), together with their corresponding uncertainty range (grey shadings). Blue vertical lines refer to the individual eruptions (14).

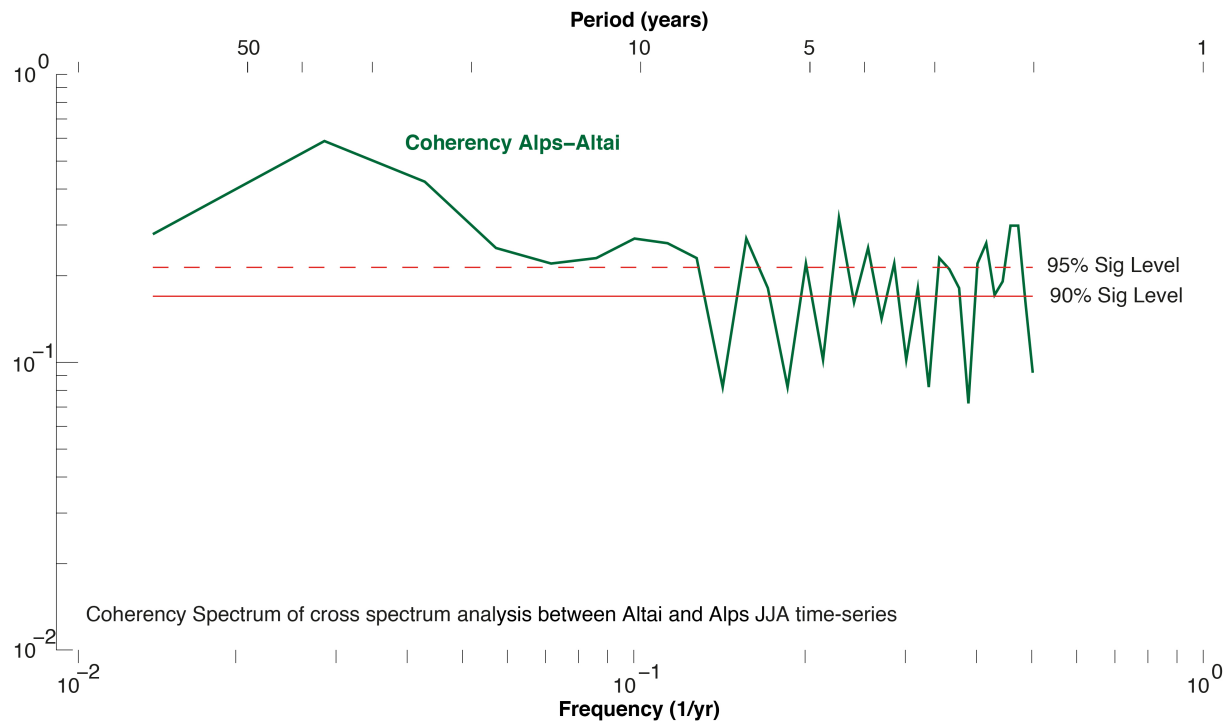


Fig. S12. Cross-spectrum between the reconstructed summer temperatures from the Alps and the Altai shows that the two records are most coherent on lower frequency domains, i.e. when the green line constantly ranges above the corresponding significance level(s). Substantial agreement between summer temperatures in the Alps and the Altai most likely results from a wave train-like pattern in the upper troposphere (Figs. S4-S5), which mainly operates on longer time-scales (26, 99).

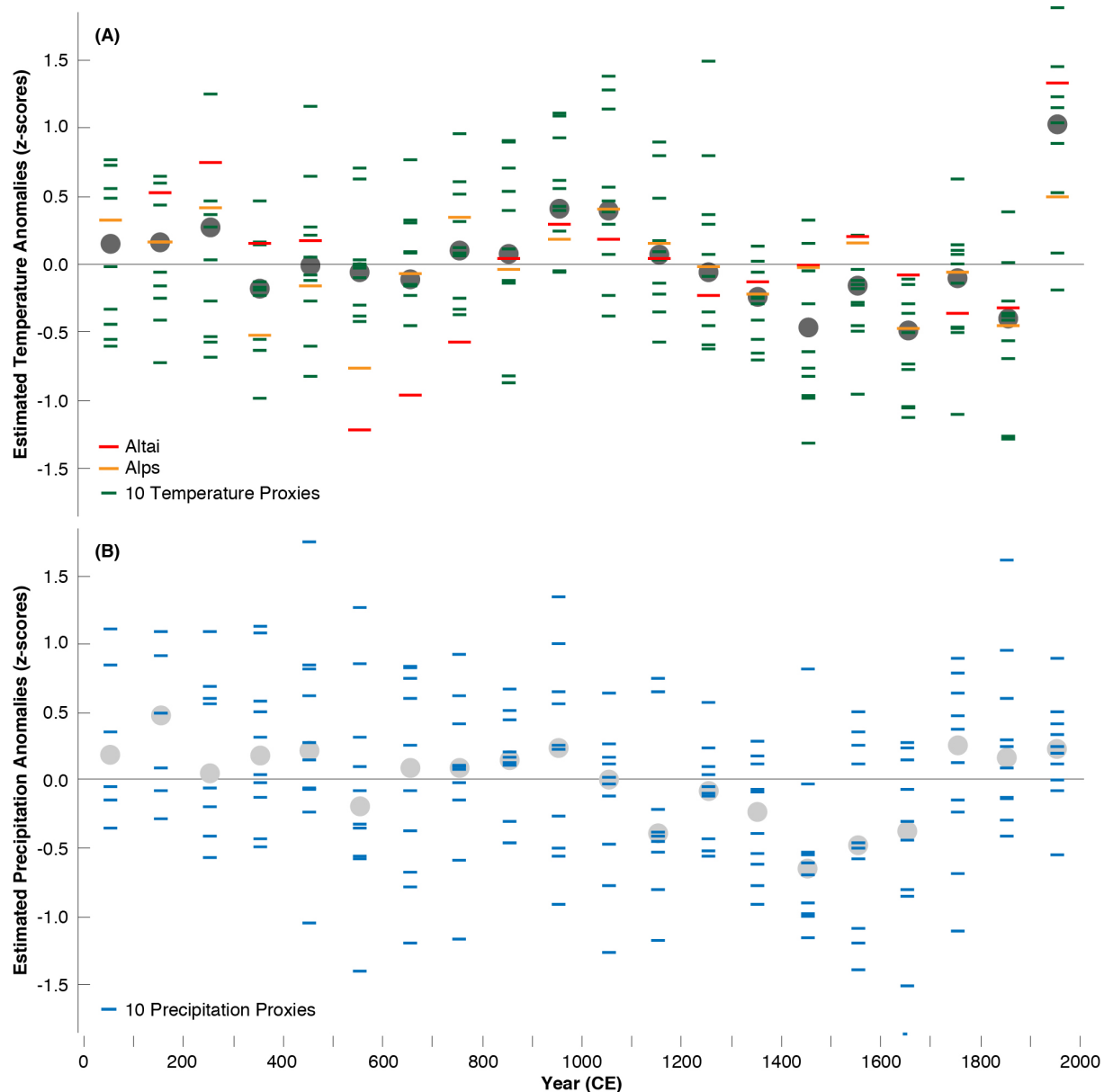


Fig. S13. Comparison of the Altai and Alpine summer temperature reconstructions with ten annual to decadal resolved warm season (A) temperature and (B) hydroclimatic proxy records from different parts of Eurasia (see Fig. 1 and Table S5 for details). Centennial means were calculated for each normalized time-series (colour lines), with grey dots indicating the median of all proxies.

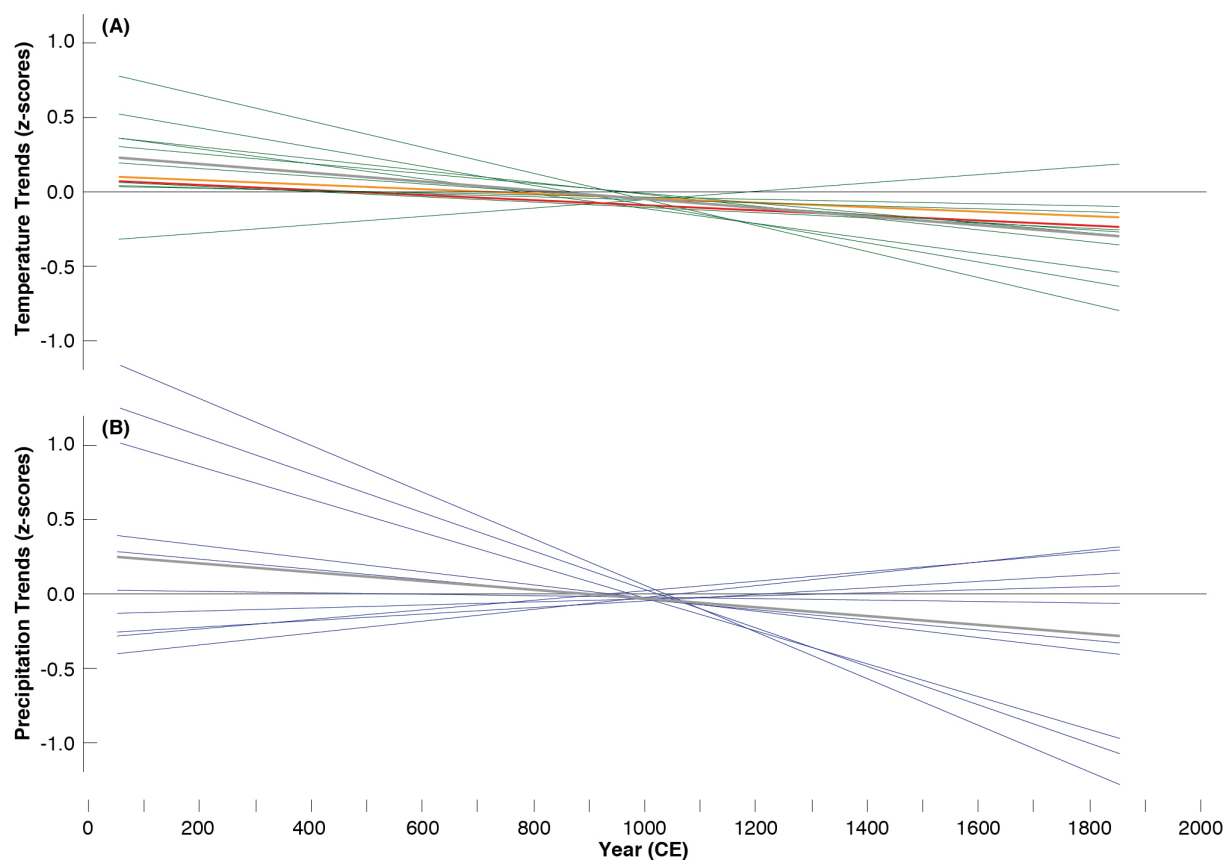


Fig. S14. Pre-industrial long-term trends of all Eurasian (A) temperature and (B) hydroclimatic proxy records (Fig. S12 1 and Table S5). Linear functions were calculated between the 1st and 19th century means.

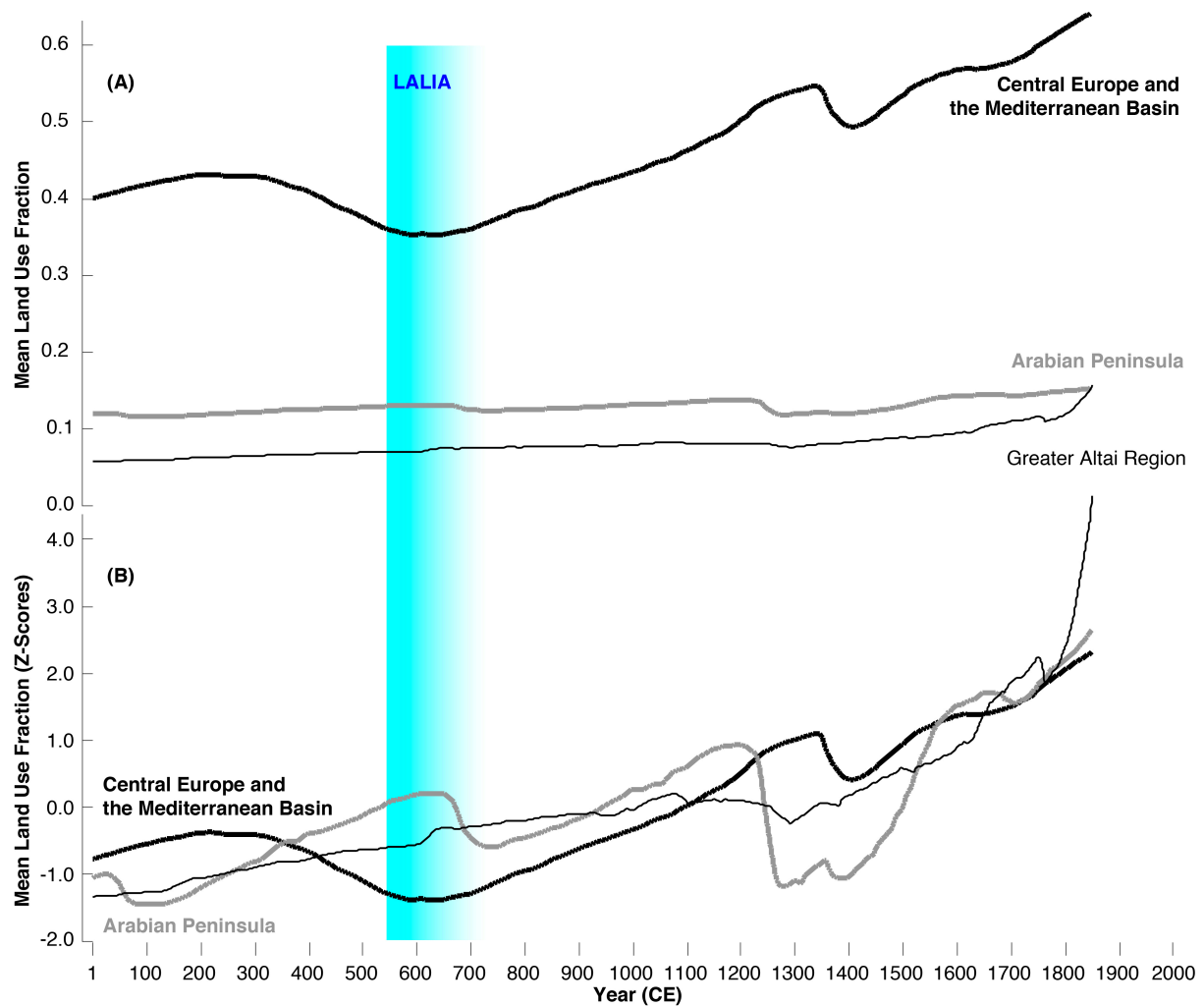


Fig. S15. (A) The mean land use fraction and **(B)** the normalized rate of anthropogenic land cover change are plotted from the KK10 scenario (100). While reconstructed land use in central Europe and the Mediterranean basin started to decline long before the onset of the LALIA, it started to increase towards the end of the LALIA. During the pre-industrial Holocene, Arabia and the Altai were characterized by extensive, low-impact forms of land use including nomadic pastoralism and therefore appear to be less sensitive to the demographic changes over the last two millennia. Nevertheless, according to this model (100), there was an increase in land use during the LALIA on the Arabian Peninsula and slightly delayed and much smaller also across the Altai. Anthropogenic land cover change in the KK10 scenario is driven by paleodemographic estimates that were generated independently of the paleoclimate reconstructions presented in this article. Future

efforts to reconstruct first millennium population dynamics should take the LALIA into account when evaluating changes in carrying capacity and the effects of environmental change on population.

Tables S1 to S5

Table S1. Inventory of all available warm-season temperature-sensitive chronologies (5, 29, 101-111) that continuously cover the last two millennia and either use tree-ring width (TRW) or maximum latewood density (MXD) (i.e. N-Scan).

Code	Record	Continent	~ Long	~ Lat	Species	Season	Reference
1	Tornetraesk	Europe	19.5 E	68.2 N	<i>P. sylvestris</i>	May-Aug	Melvin et al. 2013 (101)
2	Jämtland	Europe	12.5 E	63.2 N	<i>P. sylvestris</i>	Jun-Aug	Linderholm & Gunnarson 2005 (102)
3	Finland	Europe	24.0 E	68.0 N	<i>P. sylvestris</i>	Jun-Aug	Helama et al. 2010 (103)
4	Yamal	Asia	70.0 E	67.5 N	<i>L. sibirica</i>	Jun-Jul	Briffa et al. 2013 (104)
5	Taimyr	Asia	100.0 E	71.0 N	<i>L. gmelinii</i>	Jun-Jul	Briffa et al. 2008 (105)
6	Indigirka	Asia	148.0 E	70.0 N	<i>L. kajanderi</i>	Jun-Jul	Sidorova et al. 2006 (106)
7	Dulan	Asia	98.5 E	36.0 N	<i>S. przewalskii</i>	pJan-pDec	Liu et al. 2009 (107)
8	S-Colorado	N-America	111.7 W	35.3 N	<i>P. aristata</i>	Annual	Salzer & Kipfmüller 2005 (108)
9	Great Basin	N-America	116.0 W	38.5 N	<i>P. longaeva</i>	Jul-Sep	Salzer et al. 2013 (109)
10	Tasmania	Australia	145.5 E	41.6 S	<i>L. franklinii</i>	Nov-Apr	Cook et al. 2000 (110)
11	S-Chile	S-America	72.6 W	41.5 S	<i>F. cupressoides</i>	pDec-pMar	Lara & Villalba 1993 (111)
12	N-Scan	Europe	24.0 E	68.0 N	<i>P. sylvestris</i>	Jun-Aug	Esper et al. 2012 (29)
13	EU-Alps	Europe	12.5 E	47.5 N	<i>L. decidua</i> / <i>P. cembra</i>	Jun-Aug	Büntgen et al. 2011 (5)
14	Altai	Asia	87.5 E	47.5 N	<i>L. sibirica</i>	Jun-Aug	This Study

Table S2. Summary statistics of the new Altai TRW dataset, which compiles information from 152 living and 508 relict larch (*Larix sibirica* Ldb.) trees from five high-elevation, near treeline sites in the Russian Altai-Sayan Mountains (AK = Ak-Kalaha, JL = Jelo, KO = Koksui, MG = Mongun and TA = Tara). This unique TRW composite continuously covers 2371 years between 359 BCE and 2011 CE. Information on the mean segment length (MSL), the average growth rate (AGR), and

the first order autocorrelation (Lag1) of each subset was calculated from the raw TRW measurements.

Site	Series	Start (>10)	End (>10)	Period (>10)	MSL	AGR	Lag1
AK Living	26	1511 (1622)	2010 (2009)	500 (388)	308	0.58	0.739
JL Living	27	1483 (1662)	2011 (2011)	529 (350)	314	0.41	0.682
KO Living	27	1273 (1599)	2011 (2011)	739 (413)	351	0.493	0.671
MG Living	43	1447 (1553)	2007 (2007)	561 (455)	297	0.448	0.694
TA Living	29	1470 (1647)	2011 (2008)	542 (362)	292	0.694	0.7
AK Relict	84	164	1880	1717	367	0.375	0.764
JL Relict	103	112	1812	1701	413	0.329	0.744
KO Relict	22	588	1814	1227	348	0.363	0.746
MG Relict	259	-359	1960	2320	264	0.48	0.768
TA Relict	40	795	1966	1172	296	0.415	0.759
ALL Living	152	1273	2011	739	310	0.519	0.697
ALL Relict	508	-359	1966	2326	368	0.422	0.761
ALL 0-499	62	14	1320	1307	293	0.682	0.755
ALL 500-999	165	501	1818	1318	420	0.395	0.776
ALL 1000-1499	267	1000	2011	1012	377	0.373	0.758
ALL 1500-1999	164	1500	2011	512	279	0.521	0.692
ALL 500-1999	597	501	2011	1511	362	0.42	0.745
ALL	660	-359	2011	2371	355	0.444	0.746

Table S3. Metadata of 65 temperature-sensitive proxy records from seven different archive types distributed over much of the extra-tropical Northern Hemisphere (>20°N). Each dataset visually indicates below average values between 536 and ~660 CE, or cooling during this interval has been mentioned in the corresponding publications (herein sorted by archive first and afterwards decreasingly by latitude and longitude).

Name/site	Latitude	Longitude	Archive	Season	Reference
Eastern China	35.00	114.00	Historical Document	Winter	Ge et al. 2003 (112)
GISP2	72.60	-38.50	Ice Core	Annual	Kobashi et al. 2011 (113)
Kurupa Lake	68.36	-154.65	Lake sediments	Summer	Boldt et al. 2015 (114)
Lake Braya Sø	67.00	-50.42	Lake sediments	Summer	von Gunten et al. 2012 (115)
Screaming Lynx Lake	66.07	-145.40	Lake sediments	Summer	Clegg et al. 2011 (116)

Farewell Lake	62.55	-153.63	Lake sediments	Summer	Hu et al. 2001 (117)
Hallet Lake	61.50	-146.20	Lake sediments	Summer	McKay et al. 2008 (118)
Iceberg Lake	60.78	-142.95	Lake sediments	Summer	Loso 2009 (119)
NW of Russia	59.50	30.50	Lake sediments	Annual	Arslanov et al. 1999 (120)
Nigula	58.00	24.67	Lake sediments	Annual	Andreev & Klimanov 2000 (121)
Polystovo	56.80	31.10	Lake sediments	Annual	Chernavskaya 1995 (122)
Panfilovo	55.70	40.50	Lake sediments	Annual	Chernavskaya 1995 (122)
Teletskoe Lake	51.76	87.61	Lake sediments	Annual	Kalugin et al. 2009 (123)
Dürres Maar	50.12	6.87	Lake sediments	Summer	Moschen et al. 2011 (124)
Nádas Lake	47.99	19.70	Lake sediments	Annual	Sümegei et al. 2009 (125)
Lake Anterne	45.59	6.47	Lake sediments	Summer	Millet et al. 2009 (126)
Lake Sihailongwan	42.28	126.60	Lake sediments	Summer	Chu et al. 2011 (127)
Polovetsko-Kupanskoye	38.70	56.94	Lake sediments	Annual	Klimanov et al. 1995 (128)
Lake Qinghai	37.00	100.00	Lake sediments	Annual	Liu et al. 2006 (129)
Feni Drift	55.65	-13.99	Marine Sediments	Annual	Richter et al. 2009 (130)
Emerald Basin	45.00	-62.00	Marine Sediments	Annual	Keigwin et al. 2003 (131)
Northern East China Sea	31.68	125.81	Marine Sediments	Annual	Li et al. 2009 (132)
Northeastern Arabian Sea	24.83	65.92	Marine Sediments	Winter	Munz et al. 2015 (133)
Southern Okinawa Trough	24.80	122.50	Marine Sediments	Annual	Wu et al. 2012 (134)
Søylegrotta	66.33	13.55	Speleothems	Annual	Lauritzen & Lundberg 1999 (135)
ShiHua Cave	39.47	115.56	Speleothems	Summer	Tan et al. 2003 (136)
Avam-Taimyr	70.00	93.00	Tree Rings	Summer	(105)
Northeastern Yakutia	69.00	148.00	Tree Rings	Summer	(55, 106)
Finnish Lapland	69.00	25.00	Tree Rings	Summer	(103)
Torneträsk	68.31	19.80	Tree Rings	Summer	(101)
Northern Scandinavia	68.00	25.00	Tree Rings	Summer	(29)
Yamalia	66.92	69.17	Tree Rings	Summer	(104)
Jämtland	63.10	13.30	Tree Rings	Summer	(102)
Solongotyn Davaa	48.30	98.93	Tree Rings	Summer	D'Arrigo et al. 2001 (137)
Great Basin, USA	39.00	-114.00	Tree Rings	Summer	(109)
Qilian Mountains	38.30	99.50	Tree Rings	Annual	Zhang et al. 2014 (138)
Mid-Eastern Tibetan Plateau	36.50	98.50	Tree Rings	Summer	(107)
Cirque Peak	36.45	-118.22	Tree Rings	Summer	Scuderi 1993 (139)
Upper Wright Lakes	36.37	-118.22	Tree Rings	Summer	Lloyd & Graumlich 1997 (140)
Yakushima Island	30.20	130.30	Tree Rings	Annual	Kitagawa & Matsumoto 1995 (141)
Franz Josef Land Northbrook Island	80.00	51.00	Glacier Dynamics	Mixed	Lubinski et al. 1999 (142)
Linnévatnet, Norway	78.00	13.00	Glacier Dynamics	Mixed	Svendsen & Mangerud 1997 (143)
Western Norway	66.00	13.00	Glacier Dynamics	Mixed	Nesje 2009 (144)
Cumberland Peninsula, Baffin Island	66.00	-65.00	Glacier Dynamics	Mixed	Margreth et al. 2014 (145)
Northern Iceland	65.00	-18.00	Glacier Dynamics	Mixed	Stotter et al. 1999 (146)
Lake Hvítárvatn Langjökull ice cap	64.00	-19.00	Glacier Dynamics	Mixed	Larsen et al. 2011; 2012 (147, 148)
Northeastern Alaska Range, Alaska	63.00	-144.00	Glacier Dynamics	Mixed	Young et al. 2009 (149)
Southwest Yukon	60.00	-136.00	Glacier Dynamics	Mixed	Farnell et al. 2004 (150)
Canadian Cordilera	60.00	-140.00	Glacier Dynamics	Mixed	Reyes et al. 2006 (151)
Icy Bay, Southern Alaska	60.00	-141.00	Glacier Dynamics	Mixed	Barclay et al. 2006; 2013 (152, 153)
Southern Alaska	58.00	-137.00	Glacier Dynamics	Mixed	Wiles et al. 2011 (154)
British Columbia, Canada	56.00	-129.00	Glacier Dynamics	Mixed	Hoffman & Smith 2013 (155)
British Columbia, Canada	53.00	-120.00	Glacier Dynamics	Mixed	Maurer et al. 2012 (156)
Russian Altay	50.00	87.00	Glacier Dynamics	Mixed	(46)
Pacific Northwest of North America	48.00	-121.00	Glacier Dynamics	Mixed	Osborn et al. 2012 (157)
Swiss Alps	46.00	8.00	Glacier Dynamics	Mixed	Holzhauser et al. 2005 (158)

Western European Alps	46.00	7.00	Glacier Dynamics	Mixed	Le Roy et al. 2015 (159)
Tien Shan, Kirgiz Republic	42.00	80.00	Glacier Dynamics	Mixed	Solomina et al. 2015 (160)
Mutzag Ata- Konga Shan	38.00	75.00	Glacier Dynamics	Mixed	Seong et al. 2009 (161)
Palisade Glacier, Big Pine drainage	36.00	-118.00	Glacier Dynamics	Mixed	Bowerman & Clark 2011 (162)
Eastern Tibetan Plateau	33.00	101.00	Glacier Dynamics	Mixed	Zhang & Mischke 2009 (163)
Central-eastern Himalaya and Tibet	30.00	92.00	Glacier Dynamics	Mixed	Yang et al. 2003; 2008 (164, 165)
Rongbuk, Everest	28.00	86.00	Glacier Dynamics	Mixed	Owen et al. 2009 (166)
Greater Himalaya	28.00	82.00	Glacier Dynamics	Mixed	Murari et al. 2014 (167)
Himalaya/Karakorum	27.00	85.00	Glacier Dynamics	Mixed	Roethlisberger & Geyh 1985 (168)

Table S4. The 20 largest volcanic eruptions ranked by their global forcing (W m^{-2}) (14), used in the superposed epoch analysis (SEA) (Fig. S10).

Years	Rank	Volcanic eruption	Greenland Sulfate	Antarctica Sulfate	Global Forcing	NH Forcing	SH Forcing
			[kg km^{-2}]	[kg km^{-2}]	[W m^{-2}]	[W m^{-2}]	[W m^{-2}]
169	14	?	39.1	18.4	-11.5	-7.8	-3.7
266	9	?	61.0	11.3	-14.5	-12.2	-2.3
433	20	?	20.6	27.2	-9.6	-4.1	-5.5
536	15	?	99.0	0.0	-11.3	-11.3	0.0
540	3	Ilopango (?), El Salvador	61.2	34.4	-19.1	-12.2	-6.9
574	8	Rabaul (?), New Britain	38.3	34.1	-14.5	-7.7	-6.8
682	7	Pago (?), New Britain	38.4	38.7	-15.4	-7.7	-7.8
939	18	Eldgja, Iceland	88.7	0.0	-10.1	-10.1	0.0
1108	11	?	48.3	11.6	-12.0	-9.7	-2.3
1171	16	?	37.0	19.5	-11.3	-7.4	-3.9
1230	5	?	56.4	23.1	-15.9	-11.3	-4.6
1258	1	Samalas, Indonesia	90.4	73.4	-32.8	-18.1	-14.7
1286	19	Quilotoa (?), Ecuador	27.6	20.8	-9.7	-5.5	-4.2
1458	2	Kuwa, Vanuata	39.0	63.6	-20.6	-7.8	-12.7
1601	13	Huaynaputina, Peru	39.2	18.7	-11.6	-7.9	-3.7
1641	12	Parker, Philippines	44.2	14.9	-11.8	-8.9	-3.0
1695	17	?	28.6	22.5	-10.2	-5.7	-4.5
1783	6	Laki, Iceland	135.8	0.0	-15.5	-15.5	0.0
1809	10	?	34.6	25.4	-12.0	-6.9	-5.1
1815	4	Tambora, Indonesia	39.7	45.8	-17.2	-8.0	-9.2

Table S5. Overview of the ten temperature (1-10 T) and ten hydroclimatic (1-10 P/D) proxy records (Fig. 1B), which have been compiled and normalized for the century-

resolved comparison with the new Altai and Alpine summer temperature histories
(Fig. S13).

Code	Record	~ Long	~ Lat	Archive	Resolution	Season	Climate	Reference
1	N Spain	-3.50	42.90	Speleothems	>Decadal	Annual	T	Martin-Chivelet et al. 2011 (169)
2	Spannagel Cave	11.40	47.05	Speleothems	>Decadal	Annual	T	Mangini et al. 2005 (170)
3	N-Scan	25.00	68.00	MXD	Annual	Summer	T	Esper et al. 2014 (171)
4	Russian Plains	40.00	55.00	Multi-proxy	Decadal	Annual	T	Sleptsov & Klimenko 2003 (172)
5	Yamalia	69.00	69.00	TRW	Annual	Summer	T	(104)
6	Teletskoe Lake	87.61	51.76	L-sediments	Annual	Annual	T	(123)
7	Avam-Taimyr	93.00	70.00	TRW	Annual	Summer	T	(105)
8	Tibetan Plateau	95.00	35.00	Multi-proxy	Decadal	Annual	T	Guangliang et al. 2013 (173)
9	Mongolia	98.93	48.30	TRW	Annual	Summer	T	D'Arrigo et al. 2001 (174)
10	China composite	110.00	35.00	Multi-proxy	Decadal	Annual	T	Ge et al. 2013 (175)
1	Bunker Cave	7.66	51.37	Speleothems	Decadal	Winter	P/D	Fohlmeister et al. 2012 (176)
2	C Europe	10.00	50.00	TRW	Annual	Spring	P/D	(5)
3	Klapferloch Cave	10.55	47.95	Speleothems	Decadal	Annual	P/D	Boch & Spötl 2011 (177)
4	Nar Gölü	38.82	34.82	L-sediments	Decadal	Annual	P/D	Jones et al. 2006 (178)
5	Delingha	97.38	37.38	TRW	Annual	Annual	P/D	Yang et al. 2014b (179)
6	NE Tibet	98.00	38.00	TRW	Annual	Annual	P/D	Yang et al. 2014 (180)
7	Wanxiang Cave	105	33.19	Speleothems	Annual	Annual	P/D	Zhang et al. 2008 (181)
8	Dongge Cave	108.08	25.28	Speleothems	Decadal	Annual	P/D	Wang et al. 2005 (182)
9	Heshang Cave	109.36	19.41	Speleothems	Annual	Annual	P/D	Hu et al. 2008 (183)
10	NC China	112.00	37.00	Speleothems	Decadal	Annual	P/D	Tan et al. 2011 (184)

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