

# Last millennium Northern Hemisphere summer temperatures from tree rings: Part II, spatially resolved reconstructions

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## Abstract

Climate field reconstructions from networks of tree-ring proxy data can be used to characterize regional-scale climate changes, reveal spatial anomaly patterns associated with atmospheric circulation changes, radiative forcing, and large-scale modes of ocean-atmosphere variability, and provide spatiotemporal targets for climate model comparison and evaluation. Here we use a multiproxy

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network of tree-ring chronologies to reconstruct spatially resolved warm season (May-August) mean temperatures across the extratropical Northern Hemisphere (40-90°N) using Point-by-Point Regression (PPR). The resulting annual maps of temperature anomalies (750 to 1988 CE) reveal a consistent imprint of volcanism, with 96% of reconstructed grid points experiencing colder conditions following eruptions. Solar influences are detected at the bicentennial (de Vries) frequency, although at other time scales the influence of insolation variability is weak. Approximately 90% of reconstructed grid points show warmer temperatures during the Medieval Climate Anomaly when compared to the Little Ice Age, although the magnitude varies spatially across the hemisphere. Estimates of field reconstruction skill through time and over space can guide future temporal extension and spatial expansion of the proxy network.

*Keywords:* Tree-rings, Northern hemisphere, Last millennium, Common Era, Summer temperatures, Reconstruction, spatial

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## 1. Introduction

Global and hemispheric temperature anomalies reflect the influence of both internal variability in the climate system as well as the consequences of changes in radiative forcing, such as insolation, volcanic eruptions, and greenhouse gas concentrations. Surface temperature is determined by the planetary energy balance and serves as a symptom of perturbations to that balance, but also contains variability due to natural climate system dynamics. Rising global mean surface temperature is a key diagnostic for the influence of increasing greenhouse gases on the Earth's climate system. Yet changes in incoming solar radiation, orbital (Milankovich) changes, albedo and land use alterations, and natural and anthropogenic aerosols also influence surface temperatures. Different radiative forcing mechanisms as well as internal modes of coupled ocean-atmosphere variability may have distinct fingerprints on temperature anomalies across different spatial, temporal, and seasonal scales (Hegerl et al., 1997; Rind et al., 1999; Shindell et al., 2001b; Hegerl et al., 2003; Rind et al., 2004; Shindell et al., 2003, 2004;

16 Hegerl et al., 2006, 2007; Shindell and Faluvegi, 2009; Shindell, 2014; Shindell  
17 et al., 2015). Surface temperature anomalies are therefore controlled by the  
18 superposition of various external radiative and internal dynamical influences on  
19 the climate system. Detection and attribution of the causes of temperature fluc-  
20 tuations, as well as the prediction of future regional-scale changes, thus depend  
21 on accurate quantification and understanding of spatial and temporal variations  
22 in surface temperature (Hegerl et al., 1997; Stott and Tett, 1998; Meehl et al.,  
23 2004; Lean and Rind, 2008; Stott and Jones, 2009; Stott et al., 2010; Solomon  
24 et al., 2011; Hegerl and Stott, 2014).

25 Paleoclimate reconstructions of past temperature extend knowledge of cli-  
26 mate system variability beyond that available from the limited instrumental  
27 observational record. They offer longer timescales over which to observe a more  
28 complete range of variability in solar and volcanic forcing, extended opportuni-  
29 ties to characterize internal climate system fluctuations at decadal and longer  
30 timescales, and the potential to separate forced and unforced responses to better  
31 understand their magnitude and spatiotemporal patterns (Hegerl et al., 2003,  
32 2007). Spatially-explicit reconstructions provide additional opportunities to re-  
33 fine our understanding of fundamental climate system characteristics, diagnose  
34 the influence of different forcings on various aspects of the climate system, and  
35 provide insight into both regional climate changes and the response of large-  
36 scale modes of ocean-atmosphere variability (Seager et al., 2007; Cook et al.,  
37 2010a,b; Hegerl and Russon, 2011; Phipps et al., 2013; PAGES 2k-PMIP3 group,  
38 2015; Goosse, 2016). Comparisons between palaeoclimatic data and models also  
39 provide out-of-sample tests of the general circulation models (GCMs) used for  
40 future climate projections and can indicate where the modeled forced response or  
41 internal variability requires further evaluation and continued refinement. Such  
42 comparisons may help constrain the probable range of model parameters or  
43 identify the forcing configurations most consistent with past climate variabil-  
44 ity (Edwards et al., 2007; Anchukaitis et al., 2010; Schmidt, 2010; Hegerl and  
45 Russon, 2011; Brohan et al., 2012; Schurer et al., 2013; Schmidt et al., 2014;  
46 Harrison et al., 2015; Tingley et al., 2015).

47 Reconstructions of last millennium and Common Era surface temperatures  
48 have focused predominantly on single time-series to represent continental- to  
49 global-scale variations in mean annual or growing season temperatures aggregated  
50 over space (Frank et al., 2010; Masson-Delmotte et al., 2013; PAGES2k,  
51 2013; Stoffel et al., 2015; Smerdon and Pollack, 2016) while fewer have used climate  
52 field reconstruction (CFR) methods (Fritts, 1991; Cook et al., 1994; Evans  
53 et al., 2001; Tingley et al., 2012) to quantify past temperature anomalies simultaneously  
54 through time and across space (c.f. Mann et al., 1998; Tingley and Huybers,  
55 2013; Wang et al., 2015). Spatial field reconstructions offer the benefit  
56 of characterizing regional-scale climate changes, can reveal spatial anomaly patterns  
57 or fingerprints associated with atmospheric circulation, radiative forcing,  
58 and large-scale modes of ocean-atmosphere variability, and provide complete  
59 spatiotemporal targets for GCM evaluation (Evans et al., 2001; Anchukaitis  
60 and McKay, 2014; Kaufman, 2014; Schmidt et al., 2014).

61 Here, we develop and evaluate a climate field reconstruction of extratropical  
62 Northern Hemisphere summer temperatures using an updated network of  
63 temperature-sensitive tree-ring proxy chronologies and existing temperature reconstructions  
64 back to 750 CE (Wilson et al., 2016). We are motivated by two  
65 fundamental challenges to the development of skillful large-scale last millennium  
66 temperature reconstructions revealed over the last two decades (c.f. Frank  
67 et al., 2010; Smerdon and Pollack, 2016): First, biases arising from characteristics  
68 of the proxies themselves; Second, uncertainties arising from the choice of  
69 reconstruction methodologies.

70 Tree-ring proxies provide precise annual dating and are broadly distributed  
71 across extra-tropical land areas, making them one of the most widely used proxies  
72 for climate reconstructions of the Common Era (Hughes, 2002; Jones et al.,  
73 2009; Smerdon and Pollack, 2016). Yet despite these advantages, certain challenges  
74 or limitations exist: they preferentially reflect growing season temperature conditions,  
75 they require some manner of processing to remove non-climatic age or tree geometry  
76 related growth trends, and there exist a wide range of climate responses amongst  
77 the more than two thousand tree-ring chronologies

78 currently archived in public repositories (Briffa, 1995, 2000; Briffa et al., 2002,  
79 2004; St. George, 2014; St George and Ault, 2014). A decade ago, D’Arrigo et al.  
80 (2006) and Wilson et al. (2007) used small high-latitude networks of tree-ring  
81 proxy chronologies to reconstruct mean annual Northern Hemisphere tempera-  
82 tures. These and subsequent efforts have illuminated several extant challenges:  
83 a relatively limited number of unambiguously temperature-sensitive chronolo-  
84 gies, a predominance of ring-width chronologies in comparison to the more  
85 temperature-sensitive wood density measurements (D’Arrigo et al., 1992; Schwe-  
86 ingruber et al., 1993; Briffa et al., 2004; Frank et al., 2007; D’Arrigo et al., 2009;  
87 Esper et al., 2015; Wilson et al., 2016), and the influence of non-stationarity in  
88 climate/tree growth associations (‘divergence’; Briffa et al., 1998b; Wilson et al.,  
89 2007; D’Arrigo et al., 2008), a particular problem for North American treeline  
90 tree-ring width chronologies (Jacoby and D’Arrigo, 1995; Andreu-Hayles et al.,  
91 2011; Anchukaitis et al., 2013) and many previously collected wood density  
92 chronologies (Briffa et al., 2002). Since the publication of D’Arrigo et al. (2006)  
93 and Wilson et al. (2007), dozens of new tree-ring chronologies and local temper-  
94 ature reconstructions have become available, including many new and updated  
95 latewood density (MXD) measurement series that do not appear to exhibit any  
96 divergence (c.f. D’Arrigo et al., 2009; Esper et al., 2010; Anchukaitis et al.,  
97 2013). We draw on these new, published, and updated data here to develop a  
98 spatial reconstruction of past summer temperature stretching back to 750 CE.  
99 This work extends the non-spatial hemisphere mean reconstruction published  
100 by Wilson et al. (2016).

101 Methodologies for last millennium climate reconstructions have been exten-  
102 sively investigated and tested over the last two decades (Mann and Rutherford,  
103 2002; Rutherford et al., 2003; Zorita et al., 2003; von Storch et al., 2004; Esper  
104 et al., 2005; Mann et al., 2005; Smerdon et al., 2008; Lee et al., 2008; Smer-  
105 don et al., 2010; Mann et al., 2007; Li et al., 2010; Smerdon et al., 2011; Wang  
106 et al., 2015; Smerdon and Pollack, 2016). However, neither reduced space, em-  
107 pirical orthogonal regression methods (c.f. Fritts, 1991; Cook et al., 1994; Mann  
108 et al., 1998) nor most variants of regularized expectation maximization (RegEM

109 Schneider, 2001; Rutherford et al., 2003; Mann et al., 2009) explicitly consider  
110 the location of the proxies relative to the reconstruction target, and therefore  
111 reconstructions of individual grid points can be significantly influenced by dis-  
112 tant proxy sites. While there is potential value in taking into account large-scale  
113 teleconnections between the climate in a remote region and the local conditions  
114 controlling proxy formation (Evans et al., 2001, 2002), the potential disadvan-  
115 tage to these approaches is that they rely implicitly on the existence, stability,  
116 and persistence of those teleconnections (Gershunov and Barnett, 1998; Rimbu  
117 et al., 2003; Anchukaitis et al., 2006; Wilson et al., 2010; Lehner et al., 2012;  
118 Gallant et al., 2013; Ortega et al., 2015; Wise, 2015; Lewis and LeGrande, 2015),  
119 assume the time-stability of large-scale covariance patterns, and risk admitting  
120 distal predictors with spurious relationships to local temperatures. However,  
121 several methods exist which do account for the spatial distribution and rela-  
122 tionship of the proxy predictor network to the target field (Cook et al., 1999;  
123 Tingley and Huybers, 2010; Steiger et al., 2014). Here we use Point-by-Point  
124 regression (PPR; Cook et al., 1999) to develop, for the first time, a hemisphere-  
125 scale spatial temperature reconstruction. For an extensive review of the full  
126 range of climate field reconstruction methods, see Tingley et al. (2012).

127 We seek a spatial reconstruction of past temperatures that utilizes a predic-  
128 tor network with a clear biophysical and statistical relationship with temper-  
129 ature. We therefore apply the expert knowledge of the original developers of  
130 the individual chronologies and reconstructions, relying on their experience with  
131 respect to climate signal evaluation and statistical treatment (c.f. Esper et al.,  
132 2016). PPR is a well-established, transparent, and spatially-explicit methodol-  
133 ogy for climate field reconstruction. The result of PPR here is a gridded map of  
134 summer temperature anomalies for each year of the last  $\sim 1200$  years with asso-  
135 ciated validation skill metrics in both space and time. An additional benefit of  
136 PPR is that limiting the grid point reconstructions to proximal predictors and  
137 avoiding assumptions about the global covariance structure ensures that distant  
138 grid points remain independent of one another. The spatial features of the field  
139 can therefore be examined and compared without concern that they arise from

140 use of the same predictors.

141 We compare our results against radiative forcing over the last millennium  
142 (Schmidt et al., 2012), diagnosing the potential role of volcanic eruptions and  
143 insolation variability in shaping the Northern Hemisphere extratropical warm  
144 season temperature response across space and time. Our field reconstruction  
145 can serve as a resource for understanding temperature variability in the past,  
146 for comparison with other proxy records of environmental and climate change,  
147 and to provide context for coupled human-natural systems response to climate  
148 variability (e.g. Buckley et al., 2010; Büntgen et al., 2011; Pederson et al., 2014;  
149 Büntgen et al., 2016). Our reconstruction also provides a spatiotemporal target  
150 appropriate for formal detection and attribution of the influence of different  
151 sources of radiative forcing on the Earth’s climate system.

## 152 **2. Materials and Methods**

### 153 *2.1. Tree-ring series*

154 Tree-ring based reconstructions of large-scale climate variability, whether a  
155 single mean time series or a spatial climate field reconstruction, require selecting  
156 potential proxy predictors from the many thousands of chronologies that have  
157 been developed over the last century of tree-ring research (St. George, 2014).  
158 The majority of these chronologies were not collected or developed with the  
159 intended purpose of temperature reconstruction and do not contain a primary  
160 temperature signal. It is therefore necessary to apply some selection procedure,  
161 lest chronologies lacking temperature information overwhelm the predictor set  
162 with ‘noise’ associated with soil moisture, archaeological selection, ecological  
163 processes, or spurious non-temperature signals. Two broad categories of ap-  
164 proach have been used for initial predictor selection: statistical screening and  
165 expert assessment. In statistical screening (c.f. Mann et al., 2008), proxy se-  
166 ries are assessed for their correlations against local or regional observational  
167 temperature data, with those data passing some significance, sign, or effect  
168 size threshold then admitted to the pool of potential predictors. While this

169 approach clearly has merit as an objective and automated approach, it virtu-  
170 ally guarantees the selection of a proportion of proxies without a realistic bio-  
171 physical or substantial statistical association with the climate variable targeted  
172 for reconstruction. A second approach utilizes expert assessment of individual  
173 chronologies based on an understanding of whether a proxy has the requisite  
174 ecological, biological, geographical, and climatological characteristics to serve  
175 as a reasonable temperature proxy. Although this approach is not completely  
176 isolated from statistical considerations, the advantage is that it strongly reduces  
177 the likelihood that a non-temperature proxy or nonsense predictor will enter into  
178 a temperature reconstruction model. Expert assessment can include not only  
179 climate signal and ecological criteria, but also the methods used to develop the  
180 proxy series (e.g. Esper et al., 2016). A potential disadvantage to this approach  
181 is that it is partially subjective and therefore different investigators could make  
182 different selections for the predictor pool. Hybrid approaches that combine sim-  
183 ple mechanistic or physiological assessment with statistical evaluation have also  
184 been applied (c.f. Tierney et al., 2015).

185 Here, we follow Wilson et al. (2016) and use as our predictor series only pub-  
186 lished tree-ring chronologies and temperature reconstructions that demonstrate  
187 an established and biophysically reasonable association with local temperatures.  
188 These include both tree-ring chronologies as well as existing temperature recon-  
189 structions from Northern Hemisphere high-latitude and high-altitude locations,  
190 where dendrochronological and ecological principles suggest the most limiting  
191 factor for growth is temperature (Fritts, 1976). We exclude from considera-  
192 tion chronologies south of 40°N to avoid confounding climate signals associated  
193 with moisture-sensitive trees (St. George, 2014) and we also reject chronologies  
194 known to demonstrate evidence of the ‘divergence problem’ (Briffa et al., 1998b;  
195 D’Arrigo et al., 2008, 2009; Wilson et al., 2007), a problem previously observed  
196 to affect North American *Picea glauca* tree-ring width chronologies (D’Arrigo  
197 et al., 1992, 2009; Andreu-Hayles et al., 2011; Esper et al., 2012) and many  
198 MXD chronologies developed in the 1970s and 1980s. We require that our pre-  
199 dictors extend back to at least 1750 CE and completely forward to 1988 CE.



200 The predictors retain the detrending and standardization choices of the original  
201 authors. The resulting dataset is designated N-TREND2015 and is composed  
202 of a mixture of tree-ring width (TRW), MXD (Schweingruber et al., 1978), and  
203 blue intensity (BI; McCarroll et al., 2002) data. N-TREND2015 is archived and  
204 publicly available<sup>1</sup> and is the same dataset used in Part 1 of this study (Wilson  
205 et al., 2016). N-TREND is intended to be a ‘living dataset’ that will grow or  
206 be modified as new proxies become available or are updated. Details of the  
207 predictor time series used here and in Wilson et al. (2016) are available in Table  
208 1. As shown in Figure 1, the NTREND-2015 network reflects a mix of proxy  
209 types, dominated by MXD or BI (43 series, vs. 11 composed of tree-ring width  
210 data only). A total of 54 series are available from 1750 to 1988 CE, the time  
211 period of full (denoted ‘BEST’) coverage of the network . The number of sites  
212 drops precipitously toward the present, down to 34 by 1990 CE and to 25 series  
213 by 2000 CE, with only 3 sites remaining by 2011. There are 23 series at 1000  
214 CE and 4 remain at 750 CE, the limits of our reconstruction. While all the  
215 tree-ring chronologies and reconstructions have significant and substantial cor-  
216 relations with local temperatures in one or more months, locations that include  
217 MXD and BI data overall have higher correlations compared to sites composed  
218 of tree-ring width data alone (Figure 2; see also Wilson et al. (2016)).

## 219 *2.2. Observational data*

220 Our target field (predictand) for our temperature reconstruction is the inter-  
221 polated hybrid (surface and satellite information) version of HadCRUT4 from  
222 Cowtan and Way (2014). The original HadCRUT4 (Morice et al., 2012) con-  
223 sists of monthly temperature anomalies relative to the mean of the 1961 to 1990  
224 CE period on a regular 5° latitude/longitude grid and combines CRUTEM4  
225 (Jones et al., 2012) over land with HADSST3 (Kennedy et al., 2011a,b) for the  
226 oceans. Use of the Cowtan and Way (2014) dataset provides several advantages  
227 here: First, this dataset seeks to compensate for observational coverage bias

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<sup>1</sup>[https://www.ncdc.noaa.gov/cdo/f?p=519:1:0::::P1\\_study\\_id:19743](https://www.ncdc.noaa.gov/cdo/f?p=519:1:0::::P1_study_id:19743)

228 and provides gridded estimates of monthly temperature at high latitudes, in-  
229 cluding portions of our target reconstruction region north of  $40^{\circ}\text{N}$ . Second, it is  
230 spatially and temporally complete, allowing us to use the same calibration and  
231 validation periods for our reconstruction at every location in the field, which  
232 in turn permits straightforward comparisons of reconstruction skill. Following  
233 Wilson et al. (2016), we use May through August (MJJA) mean temperature  
234 anomalies as our target variable, as this season provides a network-wide bal-  
235 ance across the diverse site-local monthly or seasonal climate responses of the  
236 individual predictor series (Wilson et al., 2016).

### 237 *2.3. Reconstruction and statistical methodology*

238 As a prelude to our climate field reconstruction, we assess the spatial char-  
239 acteristics of the temperature signal across our network. We first calculate the  
240 Pearson Product Moment correlation between each series and the local grid-  
241 ded MJJA temperatures from Cowtan and Way (2014) as a measure of the  
242 correspondence between tree growth and local gridded temperatures. We also  
243 calculate, for each site, the field correlation between the individual predictor  
244 series and the entire MJJA and annual mean temperature field, using both the  
245 original data as well as first-differenced series. For assessing the association  
246 between the target field and the predictor network, we follow Schneider et al.  
247 (2015) and compute the median correlation coefficient between the temperature  
248 record at each grid box and all the predictor times series within 2000 km of the  
249 centroid (see below; Briffa and Jones, 1993; Jones et al., 1997; Cook et al., 2013).  
250 We perform this procedure for both the best replicated part of our predictor  
251 network (1750 to 1988 CE) and at 1000 CE in the midst of the Medieval epoch.  
252 Collectively, these statistical assessments provide an evaluation of the climate  
253 signal embedded in the predictor network through time and space.

254 We use Point-by-Point Regression (PPR; Cook et al., 1999) to reconstruct  
255 the MJJA surface temperature anomaly field north of  $40^{\circ}\text{N}$  using our predictor  
256 network of tree-ring proxy chronologies and temperature reconstructions. We  
257 follow the method as developed, tested, and described by Cook et al. (1999,

258 2010a, 2013). PPR incorporates the spatial structure of the predictor network  
259 and predictand field and confines the potential region of influence of the pre-  
260 dictors to a distance estimated from the underlying correlation structure of the  
261 temperature field. PPR proceeds by calculating a nested multivariate regression  
262 model for each grid point in the target field with the predictors restricted to  
263 those within some radius from the grid point centroid. We adopted a dynamic  
264 search radius for each grid point in the target field, first identifying predictor  
265 series within 1000km. If no chronologies were found within 1000km, the radius  
266 was allowed to expand in 500km increments up to a maximum of 2000km to find  
267 predictors. These distances are based on the decorrelation decay as a function of  
268 distance in the target field data (Cowtan and Way, 2014) and are also consistent  
269 with the findings of other studies (Briffa and Jones, 1993; Jones et al., 1997;  
270 Cook et al., 2013). If no predictors were found within 2000km, then no climate  
271 reconstruction was produced for that grid point. In evaluating our methods,  
272 we found that this dynamic search radius provides an optimal balance between  
273 maximizing the number of grid points available for reconstruction while allowing  
274 the local predictor series in data-dense regions (for instance, Fennoscandia and  
275 western Europe) to provide the paleoclimate information for their neighborhood  
276 of grid points.

277 A multivariate regression model was calibrated for each grid point and its  
278 associated predictors over the period 1945 to 1988 CE (the latest date for which  
279 all chronologies had data) and then validated on withheld observational data  
280 over the period 1901 to 1944 CE (Michaelson, 1987). We also checked the sen-  
281 sitivity of our reconstruction to this choice of calibration and validation periods  
282 by swapping them and assessing cross-validation. As the number of predictor  
283 series declines back through time, the model is newly calibrated and validated at  
284 each change in sample depth. In our reconstruction, we use the individual series  
285 themselves as predictors as opposed to the leading principal components (PCs).  
286 In our sensitivity tests of the PPR method, we discovered that using PCs from  
287 a relatively sparse network of chronologies and reconstructions, combined with  
288 the expansive target field, created clearly artificial inhomogeneities or disconti-

289 nities when predictor numbers declined. Using the individual predictor time  
290 series themselves in a stepwise regression model with an adjusted  $R^2$  entry rule  
291 (Meko, 1997) ameliorated such discontinuities. Model skill was assessed using  
292 the calibration  $R_c^2$  (adjusted for the number of predictors), the validation  $R_v^2$ ,  
293 the Reduction of Error ( $RE$ ) and the Coefficient of Efficiency ( $CE$ ) (c.f. Cook  
294 et al., 1999; Wilson et al., 2006). In addition to the annual maps of reconstructed  
295 temperatures, we calculate an extratropical Northern Hemisphere mean MJJA  
296 time series using a latitude weighted average of all the reconstructed grid cells  
297 where reconstructed values are available back to at least 1000CE and  $RE$  is  
298 greater than zero.

299 Following Masson-Delmotte et al. (2013), we calculated the difference be-  
300 tween reconstructed Medieval Climate Anomaly (MCA) and Little Ice Age  
301 (LIA) temperatures by taking the difference between the mean values over the  
302 field for 950 to 1250 CE (MCA) and 1450 to 1850 CE (LIA). As there is no single  
303 accepted definition of these two periods (Hughes and Diaz, 1994; Bradley et al.,  
304 2001; Matthews and Briffa, 2005; Seager et al., 2008; Mann et al., 2009), we  
305 also tested the sensitivity of the calculated MCA-LIA to differences in the time  
306 definition of these periods. We estimated the temperature response to tropical  
307 explosive volcanism (Robock, 2000; Ammann and Naveau, 2003) by calculating  
308 the composite mean anomaly using superposed epoch analysis (e.g. Haurwitz  
309 and Brier, 1981). Event years were extracted from most recent updated esti-  
310 mates of Common Era volcanic forcing from Sigl et al. (2015), here selecting  
311 those years corresponding to an estimated maximum negative event forcing (in  
312  $Wm^{-2}$ ) with a magnitude at least as large as Krakatoa in 1883.

### 313 **3. Results**

#### 314 *3.1. Network climate signal*

315 All our predictor series show significant and typically high correlations with  
316 local summer temperatures over one or several months (Figure 2, 3; Wilson  
317 et al. (2016)). At sites where the highest local summer temperature signal

318 in the series is confined to one or two months – for example, at Yakutia in  
319 Russia – local correlations with the broader MJJA season are lower ( $r = 0.52$ ,  
320  $p < 0.05$  for July vs.  $r = 0.09$ ,  $p > 0.05$  for MJJA at Yakutia). The highest  
321 individual monthly/seasonal site local temperature correlations (Wilson et al.,  
322 2016) range from  $r = 0.39$  to  $r = 0.84$  (mean  $r = 0.63$ ), while correlations with  
323 local MJJA temperatures range from  $r = 0.04$  to  $r = 0.78$  (mean  $r = 0.43$ ).  
324 The highest correlations with local MJJA temperatures are in Fennoscandia  
325 and north central Russia, with strong local temperature signals also evident in  
326 Scotland, the Alps, the Pyrenees, the Altai, and Japan. In North America, MXD  
327 chronologies from western Canada and the northern treeline have the strongest  
328 MJJA signals. Tree-ring width only chronologies in North America have a  
329 generally weaker association with their local MJJA instrumental temperature  
330 than MXD or mixed proxy sites (D’Arrigo et al., 1992; Jacoby and D’Arrigo,  
331 1995; D’Arrigo et al., 2009; Andreu-Hayles et al., 2011; Anchukaitis et al., 2013).

332 Field correlations between the predictor series and the full MJJA tempera-  
333 ture field (Figure 4) suggest that the chronologies and temperature reconstruc-  
334 tions reflect climate variability over many hundred or thousands of kilometers,  
335 with some exceptions at those sites where the local proxy response to MJJA is  
336 already weak (e.g. locations in east Asia, Yakutia, and ring width-only chronolo-  
337 gies from the North American treeline). The spatial extent of the large-scale  
338 correlation structure is partially related to the common positive trends in predic-  
339 tors and temperatures during the 20th century, as temporarily removing these  
340 trends by first differencing both the field and the predictors (Figure 5) reduces  
341 the regions of positive correlations to between 500 and 2000 kilometers. In  
342 some case – e.g. the Idaho (USA) chronology – a significant interannual corre-  
343 lation with the MJJA temperature field is entirely absent, indicating the sum-  
344 mer association there is driven by common trends and a narrow local monthly  
345 temperature response. Annual temperatures are often a target for tempera-  
346 ture reconstruction (e.g. Esper et al., 2002; D’Arrigo et al., 2006; Mann et al.,  
347 2008, 2009); however, correlations with the annual temperature field herein are  
348 uniformly lower and many chronologies show no significant association with an-

349 nual mean temperatures (Figure 6). In general, higher correlations with annual  
350 mean temperatures are observed for those sites with higher local correlations  
351 with MJJA temperatures ( $r_{MJJA,annual} = 0.73$ ,  $p < 0.01$ ). The association  
352 between the annual signal and the best (highest) monthly or seasonal local cli-  
353 mate correlations is weaker ( $r_{best,annual} = 0.44$ ,  $p < 0.01$ ), indicating it is not  
354 a strong climate signal *alone* that corresponds with a useful *annual* proxy, but  
355 rather a strong *and* broad seasonal climate response.

356 Figure 7 shows the statistical relationships between the predictand target  
357 field and predictor network. Grid points in Asia, Fennoscandia, and Northern  
358 Europe have 15 or more proxy sites that can serve as predictors over the best  
359 replicated epoch (1750 to 1988 CE). In northwestern North America as many as  
360 10 predictors are within 2000km of the grid centroids. In contrast, grid points in  
361 southern Europe and eastern North America have many fewer predictors avail-  
362 able, in some cases only a single series. For the best replicated portion of the  
363 reconstruction, 1750 to 1988 CE, the median correlation between observational  
364 MJJA temperatures at each grid point and the predictors within 2000km of  
365 that grid point range from  $r = -0.19$  in northeastern Russia to  $r = 0.66$  in  
366 Fennoscandia. In general, the median grid correlations are highest where they  
367 are co-located with one or more chronologies (e.g. interior British Columbia)  
368 and regions with clusters of strong MJJA temperature proxies (e.g. Fennoscan-  
369 dia, the Alps and Pyrenees, northern North American treeline). By 1000 CE,  
370 when the number of predictors is reduced to 23, with only 3 of these in North  
371 America, the possible reconstruction domain is reduced as substantially fewer  
372 predictors are available for each grid point reconstruction. Nevertheless, grid  
373 median correlations between observed temperatures and the available predictors  
374 remain significant and of similar magnitude to the better replicated recent por-  
375 tion of the reconstruction. This is because predictors with strong temperature  
376 signals remain in the Alps, Icefields in British Columbia, the Gulf of Alaska,  
377 Fennoscandia, northern Russia, and the Altai. The reconstruction at 1000 CE  
378 and earlier therefore relies on a reduced number of predictors, but those that  
379 remain contain a significant and substantial temperature signal.

380 *3.2. Field reconstruction, calibration, and validation*

381 The full NTREND spatial reconstruction consists of 1239 yearly fields of  
382 MJJA temperature anomalies covering up to a potential 792 grid points each  
383 year (40 to 90°N, -180 to 180°E). Figure 8 shows the length of the reconstructed  
384 temperature series at each grid point in our domain. In practice the number  
385 of grid points with a value in each year of the reconstruction is less than that  
386 potential maximum, as the full network can only support reconstruction at 85%  
387 ( $n = 701$ ) of the total grid cells in the domain, declining to 51% ( $n = 401$ )  
388 by 1000 CE and 24% ( $n = 190$ ) by 750 CE. The shortest reconstructions are  
389 for ocean grid points in the northwestern Atlantic and northeastern Pacific,  
390 where in both cases a single and relatively short chronology is the only source  
391 of proxy information. Other relatively short portions of the field reconstruction  
392 include the central northern treeline in Canada, Japan, northeastern Russia,  
393 and the southernmost part of the Eurasian domain in the Mediterranean. By  
394 contrast, along the Eurasian treeline in central northern Russia, throughout the  
395 Nordic region, and in western Europe, it is possible to reconstruct more than a  
396 millennium of past temperatures.

397 Figure 9 shows reconstruction skill (the adjusted  $R_c^2$ ,  $R_v^2$ ,  $RE$ , and  $CE$ )  
398 for the best replicated ( $n = 54$ ) period of the predictor network, 1750 to 1988  
399 CE. Significant skill is observed over the entire domain, but is clearly highest  
400 closest to those predictors with the highest correlations to MJJA temperatures,  
401 especially where MXD or BI data are available. Grid cells more distal from  
402 the predictor network, including cells over the oceans, or where only a single or  
403 relatively weak predictor is available, show lower explained variance and in some  
404 cases lack positive  $RE$  and  $CE$  scores. Validation  $R_v^2$  scores are lower but largely  
405 mirror the calibration  $R_c^2$ , with the exception of the eastern Mediterranean  
406 and Black Sea region, east Asia, and western parts of Central Asia. A similar  
407 phenomenon of lower  $R_v^2$  was observed by Cook et al. (2013) for Asia, due at  
408 least in part to the lack of instrumental climate data from these regions during  
409 the reconstruction model validation period (Jones et al., 2012; Harris et al.,  
410 2013; Cook et al., 2013; Cowtan and Way, 2014). Lack of instrumental data

411 likely confounds out-of-sample validation in the eastern Mediterranean prior to  
412 the 1930s (c.f. Touchan et al., 2014). Skillful regions with  $RE$  and  $CE$  scores  
413 greater than 0 are more spatially confined but likewise show skill with respect  
414 to these metrics in regions where chronologies are present or abundant, with the  
415 exception once again of the regions mentioned previously.  $R_c^2$  values range from  
416  $\sim 0.0$  to 0.78,  $R_v^2$  values range from  $\sim 0.0$  to 0.76,  $RE$  up to 0.72, and  $CE$  up  
417 to 0.70. A cross-validation (not shown) interchanging the time periods used for  
418 calibration and validation reveals that the reconstruction’s skill characteristics  
419 are largely insensitive to the choice of these periods.

420 By 1000 CE, the reduction in the number of predictors and a contraction in  
421 their spatial distribution influences both the number of grid points reconstructed  
422 and the spatial patterns of skill (Figure 10). The loss of the northern treeline  
423 MXD chronologies in North America reduces the reconstructed regions of the  
424 continent in the west to coastal Alaska, the Pacific Northwest, Interior British  
425 Columbia and parts of Alberta, and in the east to Quebec, Newfoundland, and  
426 Labrador. Likewise, the loss of the Scottish and Pyrenees chronologies no longer  
427 allow for reconstruction of temperatures over the British Isles and the Iberian  
428 peninsula. The lack of Japanese, coastal Russia, and east Asia series at 1000 CE  
429 leads to a contraction of the reconstructed spatial domain in the east. However,  
430 skillful temperature reconstructions persist over parts of northwestern North  
431 America, Fennoscandia, northern Russia, the Alps, and the Altai.  $R_c^2$  values  
432 range from  $\sim 0.0$  to 0.73,  $R_v^2$  values range from  $\sim 0.0$  to 0.70,  $RE$  up to 0.65,  
433 and  $CE$  up to 0.61. Interchanging the calibration and validation periods has a  
434 minor influence on field skill at 1000 CE, with  $n = 229$  grid points with  $RE > 0$   
435 for a late calibration (1945 to 1988 CE), and  $n = 213$  for an early calibration  
436 (1901 to 1944 CE).

437 Figure 11 shows domain-wide reconstruction and skill metrics over time.  
438 For the time span of the full predictor network (1750 to 1988 CE), we can  
439 reconstruct a temperature anomaly value for 88.5% of the grid points ( $n = 701$ )  
440 in our extratropical Northern Hemisphere domain. Of those 701 grid points,  
441 63% have  $RE > 0$ , and 37% have  $CE > 0$ . These skill percentages remain



442 remarkably stable ( $RE$ , 56% to 65%;  $CE$ , 34% to 37%) even as the number  
443 of reconstructed grid points with a reconstructed value declines back through  
444 time, a consequence of the shrinking predictor network. Only when the number  
445 of reconstructed grid points declines precipitously in the earliest 10th century,  
446 falling to 38% of the target domain in 905 CE, do the percent of grid points with  
447  $RE$  and  $CE$  greater than zero *increase* substantially. These patterns indicate  
448 that the reduction in the number of reconstructed grid cells comes at the cost  
449 of locations with already marginal skill scores, while the core reconstruction  
450 regions associated with sensitive predictor series persist through much of the  
451 length of the reconstruction back to 750 CE.

### 452 3.3. Large-scale mean temperature anomalies and climate forcing

453 We computed a mean summer temperature anomaly series from our domain  
454 by calculating a latitude-weighted mean of all gridded values where a reconstruc-  
455 tion is available back to 1000CE and  $RE$  is greater than zero (Figure 12). This  
456 time series is highly and significantly correlated with the comparable observa-  
457 tional temperature anomalies averaged over those same grid points (1850–1988,  
458  $n = 139$ ,  $r = 0.78$ ,  $p \ll 0.001$ ). The mean series indicates a broad warm  
459 period from at least 750 CE to the early 1400s, with maximum values centered  
460 around the late 900s, the late 1000s and 1100s, and the individual warmest years  
461 of the Medieval epoch in 790, 990, 995, 1014, 1016, and 1168 CE. 1168 CE is  
462 the warmest year in our reconstruction (750 to 1988 CE), although matched by  
463 values in the middle of the 20th century and then exceeded during the compara-  
464 ble filtered instrumental record in the early 21st century. Temperatures decline  
465 in the late 13th century, coincident with a series of tropical volcanic eruptions  
466 (Crowley, 2000; Gao et al., 2008; Schmidt et al., 2012; Sigl et al., 2015) and the  
467 Wolf solar irradiance minimum, before warming again during the 14th century.  
468 Temperatures then decline sharply in the early 1400s, slightly before the signif-  
469 icant volcanic eruptions in the 1450s (Gao et al., 2006; Sigl et al., 2015) and the  
470 Spörer solar irradiance minimum (1460 to 1550 CE). The mid-1400s through the  
471 mid-1800s show cooler conditions during the LIA (Matthews and Briffa, 2005;

472 Masson-Delmotte et al., 2013), with local minima in the 1450s, the late 1500s  
473 and earliest 1600s, the late 1600s, and the early 1800s. Many of the field-mean  
474 coldest years in the reconstruction, including 1259, 1453, 1601, 1643, 1783, 1810,  
475 1817, and 1836 CE are associated with or follow closely after major tropical or  
476 Northern Hemisphere volcanic eruptions (Briffa et al., 1998a; Sigl et al., 2015).  
477 In several instances extremely cold years in our reconstruction occur or persist  
478 at least one or two years after eruptions (e.g. 1643 and 1644, 1816 and 1817,  
479 1836 and 1837 CE), consistent with the findings of Esper et al. (2013b). The  
480 coldest year in our reconstruction is 1601 CE, which also agrees with a prior  
481 temperature reconstruction based solely on MXD data by Briffa et al. (1998a),  
482 and which follows the eruption of Huaynaputina in Peru in 1600 CE (Verosub  
483 and Lippman, 2008). Our ten coldest years include at least 2 (1699 and 1867  
484 CE) that do not appear to be associated with a known volcanic eruption (Briffa  
485 et al., 1998a).

486 Our filtered time series is highly and significantly correlated (750–1988 CE,  
487  $n = 1239$ ,  $r = 0.71$ ,  $p \ll 0.01$ ) with the index reconstruction from Wilson  
488 et al. (2016) and perhaps unsurprisingly has many of the same features - a  
489 broad warm period during the Medieval until the early 1400s, the LIA from  
490 the 15th through early 19th century, warming out of the LIA, a warm mid-  
491 20th century, cool 1970s, and recent warming (Figure 13a). There are epochs  
492 where our mean time series and that of Wilson et al. (2016) are less strongly  
493 correlated (Figure 13b); interestingly, the rapid returns to higher correlation  
494 values appear to be associated with the timing of major individual or clusters of  
495 volcanic events, suggesting that strong radiative forcing due to volcanism may  
496 impose a common spatial forcing, causing the Wilson et al. (2016) mean index  
497 reconstruction and the spatial mean from our climate field reconstruction to  
498 converge.

499 Our Northern Hemisphere extratropical time series shows associations with  
500 large-scale radiative forcing changes during the last millennium (Figure 14;  
501 Schmidt et al., 2012). Colder periods in the late 13th and early 14th, mid-15th,  
502 and 19th century occur at the same time as large explosive volcanic eruptions

503 and solar minima (Mann et al., 1998; Crowley, 2000; Shindell et al., 2001a, 2003;  
504 Wagner and Zorita, 2005; Ammann et al., 2007; Breitenmoser et al., 2012; An-  
505 chukaitis et al., 2013; PAGES2k, 2013). Warming after the middle of the 19th  
506 century is consistent with a reduced number of volcanic eruptions, increasing  
507 insolation, and the rapid rise in greenhouse gases (Andronova and Schlesinger,  
508 2000; Zwiers and Weaver, 2000; Gillett et al., 2012; Jones et al., 2013; Estrada  
509 et al., 2013). Although the timing of several epochs of colder temperatures ap-  
510 pear to align with solar minima – for instance, the late 13th/early 14th century  
511 and the Wolf Minimum or the mid-15th century Spörer Minimum – the corre-  
512 lations between our hemisphere mean reconstruction and estimates of past solar  
513 variability (Schmidt et al., 2012) are low (range,  $r = 0.10$  to  $r = 0.29$ ,  $p < 0.01$ ).  
514 At centennial timescales, however, there is evidence that solar variability may  
515 play a more substantial role. Wavelet coherence (Figure 15; Grinsted et al.,  
516 2004) between our hemisphere mean reconstruction and last millennium total  
517 solar irradiance estimates assembled by Schmidt et al. (2012) shows high and  
518 stable coherence and consistent phasing at bi-centennial time scales (194 to 222  
519 year periods), which bracket and include the  $\sim 206$  year ‘de Vries’ (or Suess)  
520 solar cycle (Stuiver and Braziunas, 1993; Wagner et al., 2001) and are likely  
521 related to the reconstructed temperature response to the major solar minima.  
522 The spectral signal of the bi-centennial de Vries cycle has been recognized in nu-  
523 merous tree-ring chronologies and temperature reconstructions (c.f. Raspopov  
524 et al., 2008; Breitenmoser et al., 2012; Ogurtsov et al., 2016), and Emile-Geay  
525 et al. (2013) also identified bi-centennial periodicity in a reconstruction of east-  
526 ern tropical Pacific sea surface temperatures. Phase relationships between our  
527 hemisphere mean temperature and the total solar irradiance time series suggest  
528 a decadal-scale lag of  $\sim 11$  years (range, 5 to 20 years), with solar changes lead-  
529 ing temperature anomalies, consistent with both climate modeling and prior  
530 analysis of tree-ring chronologies and solar variability (Rind et al., 1999; Waple  
531 et al., 2002; Breitenmoser et al., 2012). Both the hemisphere mean as well as  
532 the spatial grid point sensitivity to solar variability ( $C/Wm^{-2}$ ) are extremely  
533 uncertain in this analysis, however, as this quantity is highly sensitive to the

534 choice of solar reconstruction (Schmidt et al., 2012).

535 Medieval Climate Anomaly (MCA; 950 to 1250 CE) temperatures compared  
536 against those during the Little Ice Age (LIA; 1450 to 1850 CE) show warmer  
537 temperature during the MCA at  $\sim 90\%$  of the grid points with minimally skillful  
538 ( $RE > 0$ ) reconstructed values available back to 950 CE (Figure 16). Colder  
539 Medieval temperatures are reconstructed over parts of the Altai and Central  
540 Asia, and are associated with tree-ring width chronologies from Mongolia that  
541 show reduced growth in the 900s and 1100s, despite warmer conditions in the  
542 11th century (Davi et al., 2015) and an MXD chronology from the Altai dis-  
543 playing a cold Medieval epoch and warm LIA (Schneider et al., 2015). Other  
544 grid cells that show a colder Medieval period tend to be distal from the pre-  
545 dictor network – for instance in central Greenland – and must be treated with  
546 caution. Defining a different MCA or LIA in this case has relatively little effect  
547 on the percentage of grids showing warmer vs. colder conditions, as the cooler  
548 Central Asia grid cells and marginal cells in Greenland, central Canada, the  
549 southern Caspian Sea, and the northeastern Pacific remain irrespective of the  
550 specific date range applied. The precise boundaries for both MCA and LIA  
551 are, in any case, both arbitrary and uncertain (Hughes and Diaz, 1994; Bradley  
552 et al., 2001; Matthews and Briffa, 2005; Seager et al., 2008). The cause of ap-  
553 parently extremely high ( $\sim 3^\circ\text{C}$ ) Medieval temperatures in several grid points  
554 in northeastern North America is discussed below.

555 Composite mean MJJA temperature anomaly fields following major volcanic  
556 eruptions show coherent, broad-scale cooling associated with large tropical erup-  
557 tions (Figure 17). 96% of grid points show composite mean colder temperatures  
558 compared to the three years prior to the 20 large eruptions considered here.  
559 Similar to the MCA-LIA difference discussed above, regions that apparently  
560 have an overall composite warming response to volcanic eruptions are largely  
561 on the margins of the reconstruction domain, away from the predictor grid, and  
562 over the ocean, central Greenland, and the southern Caspian Sea. The cold-  
563 est grid point composite mean is  $-1.61^\circ\text{C}$ , and the mean composite response  
564 across all grid points and all eruptions is  $-0.44^\circ\text{C}$ . Closer examination of indi-

565 vidual eruption events (not shown) finds that for some regions, post-volcanic  
566 cooling may persist for several years and maximum cold anomalies may be 1  
567 or 2 years after the year of the eruption itself, consistent with observations of  
568 other regional temperature reconstructions (D'Arrigo et al., 2013; Cook et al.,  
569 2013; Esper et al., 2013b; Davi et al., 2015; Linderholm et al., 2015; Schneider  
570 et al., 2015; Wilson et al., 2016). If we consider large Northern Hemisphere  
571 high-latitude eruptions only (Figure 18), the large-scale response is likewise to-  
572 ward cold anomalies overall: 89% of grid points have a composite anomaly less  
573 than zero. Over the entire field the mean composite response is  $-0.39^{\circ}\text{C}$  and the  
574 maximum cold composite anomaly is  $-2.31^{\circ}\text{C}$ . There is also spatial structure to  
575 the temperature anomalies, with the coldest composite conditions over Alaska  
576 and the Bering Strait, northeastern North America, parts of western Europe,  
577 and central northern Russia, suggestive of a dynamical, in addition to direct  
578 radiative, influence of large-magnitude high latitude eruptions (Robock, 2000;  
579 Oman et al., 2005; Stenchikov et al., 2006; Schneider et al., 2009; Zanchettin  
580 et al., 2012; Pausata et al., 2015). However, the number of radiatively signifi-  
581 cant high latitude eruptions considered here is smaller ( $n = 5$ ), and therefore  
582 this structure may appear due to the limited sample size.

## 583 4. Discussion

### 584 4.1. Proxy data and predictor network

585 Our results here demonstrate that a relatively small ( $n = 54$ ) network of  
586 proxy sites (Table 1, Figure 1) with well-established physically and ecologically  
587 reasonable climate signals (Figure 3, 7) can be used to reconstruct the large-scale  
588 summer temperature history of the extratropical Northern Hemisphere (Figure  
589 8, 9, 10, 12). While restricting the reconstruction to the higher latitudes of one  
590 hemisphere and to only the growing season does not provide a global annual es-  
591 timate of temperature, it nonetheless accurately reflects the geographic and bi-  
592 ological signal that dominates the predictors. Moreover, we have demonstrated  
593 here that the reconstruction preserves the signature and influence of external

594 forcing on the global energy balance. Skill in our reconstruction is, perhaps not  
595 surprisingly, greatest in those locations where high quality temperature-sensitive  
596 proxies are available (Figure 9, 10), and declines at increasing distances from  
597 the predictors themselves. It is clear we could realize substantial benefits in  
598 terms of increased reconstruction skill and spatial extent by developing MXD  
599 and BI chronologies from currently undersampled regions as well as extending  
600 the length of existing MXD and BI chronologies. Although large and useful  
601 multiproxy datasets have resulted from community efforts by the paleoclimate  
602 community (PAGES2k, 2013), our analysis here demonstrates that a relatively  
603 small well-distributed network of highly sensitive millennium-length tree ring  
604 chronologies provide skillful reconstructions over a large extratropical region.  
605 Encouragingly, this means that rapid and important gains could be made from  
606 the addition of a relatively small number of new sites and the temporal exten-  
607 sion and recollection of current sites known to contain a strong climate signal.  
608 In particular, a greater number of long MXD and BI chronologies are critically  
609 needed from North America (Figures 1, 10). Fulfilling this need will require a  
610 collaborative and concerted effort to locate subfossil materials and to measure  
611 density proxies, but the potential gain for Northern Hemisphere temperature  
612 reconstructions of the Common Era would be substantial. MXD contains a  
613 stronger temperature signal than TRW alone and is better able to accurately  
614 resolve rapid temperature changes associated with volcanic eruptions (Figure  
615 2; Frank et al., 2007; D’Arrigo et al., 2013; Esper et al., 2015). Continuing  
616 advances in blue intensity (BI) measurements suggest some of the benefits of  
617 wood density analysis can be realized without the expense and difficulty of an-  
618 alyzing MXD itself (Campbell et al., 2007; Wilson et al., 2014; Rydval et al.,  
619 2014; Björklund et al., 2014; Björklund et al., 2015) although the low-frequency  
620 characteristics of BI still requires additional exploration.

621 Detrending and standardization issues for long chronologies remain an ongo-  
622 ing challenge and persistent source of uncertainty (e.g. Cook et al., 1995; Briffa  
623 and Melvin, 2011; Melvin and Briffa, 2008; Esper et al., 2012; Matskovsky and  
624 Helama, 2014; Esper et al., 2016; Matskovsky and Helama, 2016). One surpris-

625 ing feature of the epochal comparison between the MCA and LIA is the large  
626 ( $> 3^{\circ}\text{C}$ ) difference calculated for northeastern North America. This feature is  
627 due to a single predictor, a black spruce (*Picea mariana*) tree-ring width RCS  
628 chronology developed by Gennaretti et al. (2014). Other, non-tree ring proxies  
629 from the region suggest a lower amplitude of cooling between the MCA and the  
630 LIA, although issues related to time uncertainty, transfer function calibration,  
631 and different seasonal climate signals complicate exact comparisons. A com-  
632 pilation of Holocene paleoenvironmental data for the Arctic (Sundqvist et al.,  
633 2014) suggests a range of values for MCA to LIA cooling in northeastern North  
634 America and Greenland of 0 to  $1.5^{\circ}\text{C}$ , which is at least a degree less than the  
635 magnitude inferred from the Quebec MXD record. Alkenone SST reconstruc-  
636 tions near Nova Scotia (Keigwin et al., 2003) suggest a cooling of approximately  
637  $0.7^{\circ}\text{C}$ . The multiproxy PAGES Arctic2k reconstruction (McKay and Kaufman,  
638 2014) has a whole-Arctic reconstructed MCA-LIA difference of  $0.64^{\circ}\text{C}$  for the  
639 same time periods used here. Finally, and perhaps most importantly, a tree-  
640 ring oxygen isotope temperature reconstruction from the same site in Quebec  
641 (Naulier et al., 2015) shows a substantially smaller estimated MCA-LIA differ-  
642 ence of  $0.4^{\circ}\text{C}$ . It seems likely that, despite the care taken in applying regional  
643 curve standardization to the Quebec black spruce samples (Autin et al., 2015)  
644 as well as its suitability with respect to other chronology metrics (Esper et al.,  
645 2016), artifacts remain in this tree-ring width chronology that unintentionally  
646 but artificially amplify the difference between MCA and LIA temperatures in  
647 this region.

648 More generally, detrending, the removal of non-climatic trends, and there-  
649 fore the retention of low frequency variability, remains an important source  
650 of uncertainty in the amplitude of past temperatures reconstructed from tree  
651 rings (Cook, 1987; Briffa et al., 1996), even when conservative detrending tech-  
652 niques have been applied. While regional curve standardization and signal free  
653 methods have been shown to be able to retain the full spectrum of low- and  
654 medium-frequency variability, they are also subject to their own uncertainties  
655 and assumptions (Melvin, 2004; Melvin and Briffa, 2008; Briffa and Melvin,

2011; Anchukaitis et al., 2013; Briffa et al., 2013). It is in most cases not possible to know from calibration and validation statistics which detrending method yields the true or most accurate low frequency signal (Cook, 1987; Cook and Kairiūkštis, 1990; Wilson et al., 2007). Possible approaches to this problem include both ensemble and simulation-based methods (e.g. Esper et al., 2007; D’Arrigo et al., 2011; Anchukaitis et al., 2013), although these have not yet been applied to large and heterogeneous tree-ring proxy networks.

#### 4.2. Radiative forcing and temperature history

Our reconstruction demonstrates coherent responses to radiative forcing in time and space (Figures 14, 15, 16, 17, 18). Temporal features of the reconstruction are associated with changes in solar irradiance, large and/or clustered tropical volcanic eruptions, and the anthropogenic rise in well-mixed greenhouse gases. Temperatures decline across the field during the Spörer and Maunder Minima, in particular, likely compounded in both cases by a series of volcanic eruptions. Temperatures remained cold during the early 1600s, at least in part due to the eruption of Huaynaputina in Peru in 1600 CE (Verosub and Lippman, 2008). 1601 CE is the coldest year of our entire reconstruction, as it was in the 600 year temperature reconstruction by Briffa et al. (1998a) and the hemisphere mean reconstruction by Wilson et al. (2016). 1601 CE was also one of the coldest years in the Bayesian field reconstruction by Tingley and Huybers (2013), as was 1453 CE, which is the 4th coldest year in our study, associated with the eruption of of Kuwae, Vanuatu (but see Plummer et al., 2012; Cole-Dai et al., 2013). Interestingly, Tingley and Huybers (2013) find 1642 CE amongst their coldest years, whereas in our reconstruction it is 1643 CE that is exceptionally cold (5th coldest in our reconstruction). In Wilson et al. (2016), 1641, 1642, and 1643 CE are all amongst the coldest 15 years of their reconstruction. Tingley and Huybers (2013) also find that 1695 CE was anomalously cold, whereas here it is indeed cold but unremarkable ( $-0.60^{\circ}\text{C}$ , 284th coldest). These differences highlight extant uncertainties likely related to different reconstruction methods, spatial skill and averaging, and the use of different proxies (D’Arrigo et al.,



686 2013; Esper et al., 2015), but also demonstrate that there is no evidence for a  
687 one-to-one correspondence between inferred volcanic forcing from ice cores and  
688 the magnitude of hemisphere-scale cooling. For instance, the eruption of Huay-  
689 naputina in 1600 is believed to have caused a smaller negative radiative forcing  
690 anomaly than eruptions in 1458, 1641, 1809, and 1815 CE, let alone the large  
691 Medieval eruption of Samalas (1257 CE) (Verosub and Lippman, 2008; Lavi-  
692 gne et al., 2013; Sigl et al., 2015). Our finding here of a large-scale, coherent  
693 cooling in response to explosive volcanism is yet further evidence (Anchukaitis  
694 et al., 2012; Brohan et al., 2012; D’Arrigo et al., 2013; Esper et al., 2013b,a;  
695 St. George et al., 2013; Büntgen et al., 2014; Jull et al., 2014; Esper et al., 2015;  
696 Sigl et al., 2015; Stoffel et al., 2015; Wilson et al., 2016) against the hypothesis  
697 that tree-ring proxies are missing the volcanic cooling signal due to undetected  
698 absent rings (Mann et al., 2012, 2013).

699 Low frequency coherence between solar variability and our reconstruction  
700 appears to be a stable characteristic through time (Figure 14, 15), likely linked  
701 to reconstructed cold anomalies during solar Grand Minima. Because varia-  
702 tions in total solar irradiance are relatively small, on the order of a few tenths  
703 of a  $Wm^{-2}$ , the mechanism that could result in a detectable cooling remains  
704 uncertain. The most likely connection is via changes in large-scale Northern  
705 Hemisphere circulation, which favor colder temperature over continents (e.g.  
706 Shindell et al., 2001a, 2003; Swingedouw et al., 2010) and thus would be cap-  
707 tured in our reconstruction. Nevertheless, while variability in solar forcing may  
708 be important on bicentennial and perhaps at continental scales, fingerprinting  
709 suggests that the solar effect in the hemisphere-scale anomalies is otherwise rel-  
710 atively small and that volcanic forcing is more important overall in determining  
711 pre-industrial temperature trajectories (Schurer et al., 2014; McGregor et al.,  
712 2015). There is no sign in our reconstruction of a discernible temperature re-  
713 sponse to the shorter 11 and 22 year sunspot cycle (Schwabe/Hale), which is  
714 consistent with other investigations of the insolation signal in tree rings (e.g.  
715 Briffa, 1994). There are a number of possible reasons for the absence of this sig-  
716 nal: Internal climate system variability is substantially stronger at interannual

717 and decadal time scales, which may prevent statistical detection of solar influ-  
718 ences with similar frequencies, but still allow it at the centennial scale when the  
719 magnitude of internal variability is smaller than the forced signal. Short-term  
720 climate anomalies caused by explosive volcanism could also disrupt detection of  
721 a decadal solar signal. The temperature response to solar variability at lower  
722 frequencies may also reflect slow temperature feedbacks that enhance its direct  
723 effect.

724 Over those grid points available back to 950 CE with minimum level of re-  
725 construction skill ( $RE > 0$ ),  $\sim 90\%$  show warmer conditions during the MCA  
726 than during the LIA, with a field median difference of  $0.32^\circ\text{C}$ . Removing likely  
727 individual grid point outliers (Greenland and northeastern North America, see  
728 above) results in a slightly smaller epochal field median difference ( $0.30^\circ\text{C}$ ) and  
729 a range of grid point values of  $-0.64$  to  $+1.05^\circ\text{C}$ . Mann et al. (2009) calculated  
730 a  $0.24^\circ\text{C}$  global summer mean difference between MCA and LIA, but the differ-  
731 ence in season, spatial domain and geographic extent, and the ‘fragility’ (Wang  
732 et al., 2015) of reconstructing a cold Medieval tropical Pacific make any direct  
733 comparison difficult. Calculating the MCA-LIA epochal difference using the  
734 spatial mean time series (Figure 12) gives a value of  $0.36^\circ\text{C}$ , approximately in  
735 the middle of the distribution for other large-scale Northern Hemisphere recon-  
736 structions, and within the higher end of the range of values from climate model  
737 simulations (Fernández-Donado et al., 2013; Wilson et al., 2016).

## 738 5. Conclusions and future work

739 We have reconstructed the extratropical Northern Hemisphere MJJA tem-  
740 perature anomaly field back to 750 CE using a network of temperature-sensitive  
741 predictors. The reconstruction shows significant field skill associated with prox-  
742 imity to the predictors, particularly where proxy density data are available. In  
743 other words, we observe the most reconstruction skill and smallest errors where  
744 we have the most sensitive tree-ring proxies, whereas higher errors and lower skill  
745 are associated with grid points distal from the predictor network or where only

746 tree-ring width data are available, particularly in North America. These obser-  
747 vations will be used to guide future sampling and proxy development priorities,  
748 including the development of new sites, efforts to increase the number of MXD  
749 and BI series, and the extension in time of existing high quality chronologies.

750 Our field reconstruction reveals coherent responses to changes in radiative  
751 forcing over the last 1200 years, including the influence of solar and volcanic  
752 forcing. Future research with our field reconstruction will use fingerprint detec-  
753 tion (Hegerl et al., 2007; Schurer et al., 2013, 2014) to quantitatively assess the  
754 role of forcing and internal variability, including identification of spatial patterns  
755 linked to large-scale modes of variability and specific forcing agents. Formal,  
756 quantitative comparison between our reconstruction and paleoclimate model  
757 simulations (Schmidt et al., 2012; Kageyama et al., 2016) will be used to assess  
758 climate model performance and to investigate the dynamical context for re-  
759 constructed spatial temperature anomalies. Using proxy system models (Evans  
760 et al., 2013), the NTREND network could also be applied within an offline data  
761 assimilation framework (Steiger et al., 2014; Hakim et al., 2016). Finally, our  
762 spatially-explicit reconstructions can be used to explore and understanding the  
763 possible role of past temperature variability – especially volcanic eruptions – in  
764 contributing to historical societal dynamics, resilience, and change (McCormick  
765 et al., 2007; Ludlow et al., 2013; Sigl et al., 2015; Büntgen et al., 2016)

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783 project website, along with the archived TR chronologies and temperature re-  
784 constructions can be found at [https://www.ncdc.noaa.gov/cdo/f?p=519:1:](https://www.ncdc.noaa.gov/cdo/f?p=519:1:0:::P1_study_id:19743)  
785 [0:::P1\\_study\\_id:19743](https://www.ncdc.noaa.gov/cdo/f?p=519:1:0:::P1_study_id:19743) and additional information is available at [https:](https://ntrenddendro.wordpress.com/)  
786 [//ntrenddendro.wordpress.com/](https://ntrenddendro.wordpress.com/). Lamont-Doherty Earth Observa-  
787 tion contribution #####.

Table 1: Tree-ring chronology and temperature reconstruction predictor series used in the reconstruction. Additional information is available in Wilson et al. (2016). Latitude is given in degrees North, and longitude is in degrees East. A range of latitude/longitude indicates the data or reconstruction cover a larger region. Detrending: STD: traditional detrending standardization; RCS: regional curve standardization; SF: signal free extension to STD or RCS standardization. † indicates series used in D'Arrigo et al. (2006) and ‡ those used in Wilson et al. (2007)

Site Name	Code	Latitude	Longitude	Time Span	Proxy Type	Detrending	Citation
<b>NORTH AMERICA</b>							
Seward	NTR	65.11 to 65.22	-162.18 to -162.27	1710-2001	MXD	STD	D'Arrigo et al., 2004
Coastal Alaska	GOA	60.01 to 60.45	-149.31 to -141.42	800-2010	TRW	RCS	Wiles et al., 2014
Wrangells	WRAX	60-65	-145.00 to -140.00	1593-1992	MXD	STD	Davi et al., 2003 ‡
Firth	FIRT	68.39	-141.38	1073 - 2002	MXD	RCS-SF	Anchukaitis et al. 2013
Southern Yukon	YUS	59 to 62	-140 to -133	1684-2000	TRW	STD	Youngblut and Luckman, 2008 ‡
Northern Yukon	YUN	65 to 70	-125 to -135	1638-1988	TRW	STD	Szeicz and MacDonald, 1995 †
Int. British Columbia	IBC	49.02 to 50.59	-121.43 to -117.03	1600-1995	TRW/MXD/BI	STD/SF	Wilson et al., 2014
Icefields	ICE	52.16	-117.19	918 - 1994	RW/MXD	RCS	Luckman and Wilson, 2005 †
Idaho	IDA	40 to 45	-110 to -120	1135-1992	TRW	STD	Biondi et al., 1999 ‡
Coppermine	COP	67.14	-115.55	1551 - 2003	MXD	STD	D'Arrigo et al. 2009, Anchukaitis et al. 2013
Thelon	THE	64.02	-103.52	1492 - 2004	MXD	STD	D'Arrigo et al. 2009, Anchukaitis et al. 2013
Quebec	QUEX	57.30	-76.00	1373-1988	MXD	RCS	Schneider et al. 2015
Quebec	QUEW	57.30	-74.00	910-2011	TRW	RCS	Gennaretti et al., 2014
Northern Quebec	NQU	55 to 60	-70 to -65	1642-2002	TRW	STD	Payette, 2007 ‡
Labrador	LABrec	56.33 to 57.58	-62.25 to -61.56	1710-1998	TRW/MXD	STD/RCS	D'Arrigo et al., 2003, 2013
<b>EURASIA</b>							
Scotland	SCOT	57.08	-3.44	1200-2010	TRW/BI	STD/RCS	Rydval et al. in review
Pyrenees	PYR	42 to 43	0 to 1	1260-2005	MXD	RCS	Dorado-Liánán et al., 2012
W Alps - Lhotschental	ALPS	46.5	9	755-2004	MXD	RCS	Büntgen et al., 2006
E Alps - Tyrol	TYR	47.30	12.30	1053-2003	MXD	RCS	Schneider et al., 2015
Jaemland	JAEM	63.30	13.25	783-2011	MXD	RCS	Zhang et al., 2016
Tjeggelvas, Arjeplog, & Ammaråns composite	TAA	65.54 to 66.36	16.06 to 18.12	1200-2010	MXD	RCS	Linderholm et al., 2015
North Fennoscandia	EFmean	66 to 69	19 to 32	750-2010	MXD	RCS	Esper et al., 2014, Matskovsky and Helama, 2014
Forfjordalen	FORF	68.47	15.43	978-2005	MXD	RCS	McCarroll et al., 2013
Tatra	TAT	48 to 49	19 to 20	1040-2010	TRW	RCS	Büntgen et al., 2013
Mt Olympus, Greece	MOG	40.09	22.37	1521-2010	MXD	RCS	Klesse et al., 2015
South Finland	SFIN	62.19	28.19	760-2000	MXD	RCS	Helama et al., 2014
Khibiny (Kola)	KOL	67.38 to 67.50	33.13 to 34.15	821-2005	RW/BI	RCS/STD	McCarroll et al., 2013
Polar Urals	POLx	66.51	65.40	891-2006	MXD	RCS	Schneider et al. 2015
Yamal	YAM	67.32	69.54	750-2005	TRW	RCS-SF	Briffa et al., 2013
Asia Grid 1	Grid1	40.15 to 46.15	60.15 to 68.15	817-1989	mix	RCS/STD	Cook et al., 2013
Asia Grid 2	Grid2	40.15 to 46.15	70.15 to 78.15	827-1989	mix	RCS/STD	Cook et al., 2013
Asia Grid 10	Grid10	48.15 to 54.15	60.15 to 68.15	937-1989	mix	RCS/STD	Cook et al., 2013
Asia Grid 11	Grid11	48.15 to 54.15	70.15 to 78.15	937-1989	mix	RCS/STD	Cook et al., 2013
Kyrgyzstan	KYR	41.36 to 42.11	75.09 to 78.11	1689-1995	TRW/MXD	STD	Wilson et al., 2007 ‡
Mangazjea	MAN	66.42	82.18	1328-1990	MXD	RCS	Schneider et al. 2015
Asia Grid 3	Grid3	40.15 to 46.15	80.15 to 88.15	800-1989	mix	RCS/STD	Cook et al., 2013
Asia Grid 12	Grid12	48.15 to 54.15	80.15 to 88.15	800-1989	mix	RCS/STD	Cook et al., 2013
Altai MXD	ALT	50.00	88.00	750-2007	MXD	RCS	Schneider et al. 2015
Asia Grid 4	Grid4	40.15 to 46.15	90.15 to 98.15	800-1989	mix	RCS/STD	Cook et al., 2013
Asia Grid 13	Grid13	48.15 to 54.15	90.15 to 98.15	1024-1989	mix	RCS/STD	Cook et al., 2013
Mongolia	OZN	51.15	99.04	931-2005	TRW	RCS	Davi et al., 2015
Taymir	TAY	72.01	102.00	755-1997	TRW	RCS	Jacoby et al., 2000 †
Asia Grid 5	Grid5	40.15 to 46.15	100.15 to 108.15	800-1989	mix	RCS/STD	Cook et al., 2013
Asia Grid 14	Grid14	48.15 to 54.15	100.15 to 108.15	1396-1989	mix	RCS/STD	Cook et al., 2013
Asia Grid 6	Grid6	40.15 to 46.15	110.15 to 118.15	800-1989	mix	RCS/STD	Cook et al., 2013
Asia Grid 15	Grid15	48.15 to 54.15	110.15 to 118.15	1396-1989	mix	RCS/STD	Cook et al., 2013
Asia Grid 7	Grid7	40.15 to 46.15	120.15 to 128.15	1024-1989	mix	RCS/STD	Cook et al., 2013
Asia Grid 16	Grid16	48.15 to 54.15	120.15 to 128.15	1510-1989	mix	RCS/STD	Cook et al., 2013
Asia Grid 8	Grid8	40.15 to 46.15	130.15 to 138.15	1510-1989	mix	RCS/STD	Cook et al., 2013
Asia Grid 17	Grid17	48.15 to 54.15	130.15 to 138.15	1510-1989	mix	RCS/STD	Cook et al., 2013
Asia Grid 9	Grid9	40.15 to 46.15	140.15 to 148.15	1510-1989	mix	RCS/STD	Cook et al., 2013
Asia Grid 18	Grid18	48.15 to 54.15	140.15 to 148.15	1510-1989	mix	RCS/STD	Cook et al., 2013
North Japan	NJAP	43 to 51	142 to 145	1640-1993	TRW/MXD	STD	D'Arrigo et al., 2015
Yakutia	YAK	67.27 to 70.33	142.37 to 150.17	1342-1994	TRW	RCS	Hughes et al., 1999 †

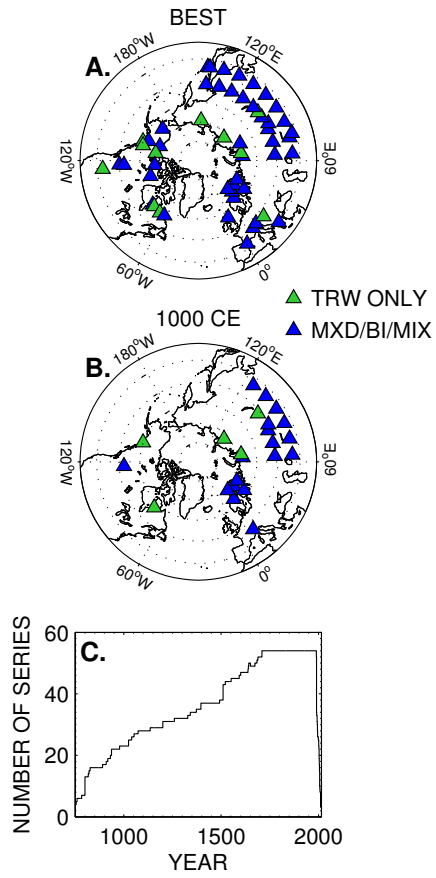


Figure 1: Spatial and temporal distribution of sites by proxy type. Top panel (A) shows all the records available during the best replicated (most recent, 1750 to 1988 CE) nest, while the bottom panel (B) shows the distribution of available proxy sites at 1000 CE. Sites with tree-ring width (TRW) data only are shown in green, while sites that have MXD, BI, or a mix of proxy types are shown in blue. (C) Shows the total number of series through time.

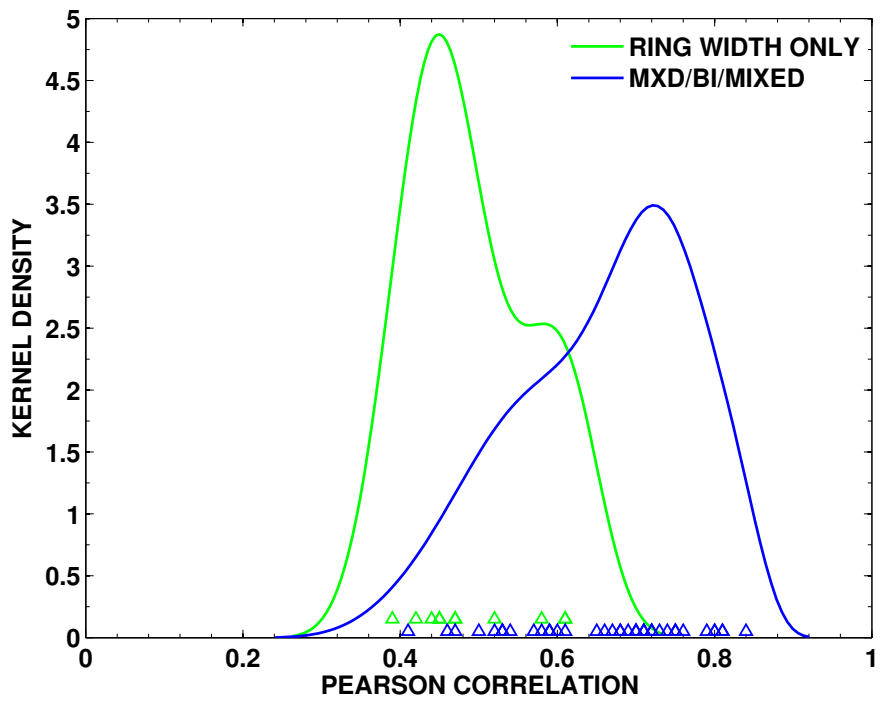


Figure 2: Kernel probability density estimate (Bowman and Azzalini, 1997) for the highest local seasonal or monthly coefficient correlations as a function of proxy type. Data are from Wilson et al. (2016)). The density estimate is calculated for a support of  $[-1,1]$  and the values contributing to the distribution are indicated by symbols along the x-axis.

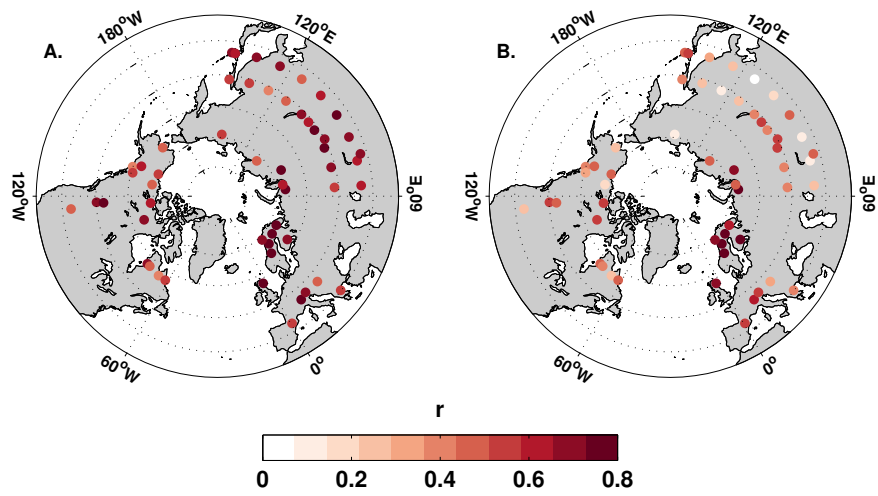


Figure 3: Local correlations between tree-ring proxy chronologies and local temperature data. (A) Correlations reported by the original authors for local correlations with the optimal seasonal or monthly window (see Wilson et al., 2016, their Table 1). (B) Correlations between each proxy series and the local May through August (MJJA) temperature data from Cowtan and Way (2014) used here.



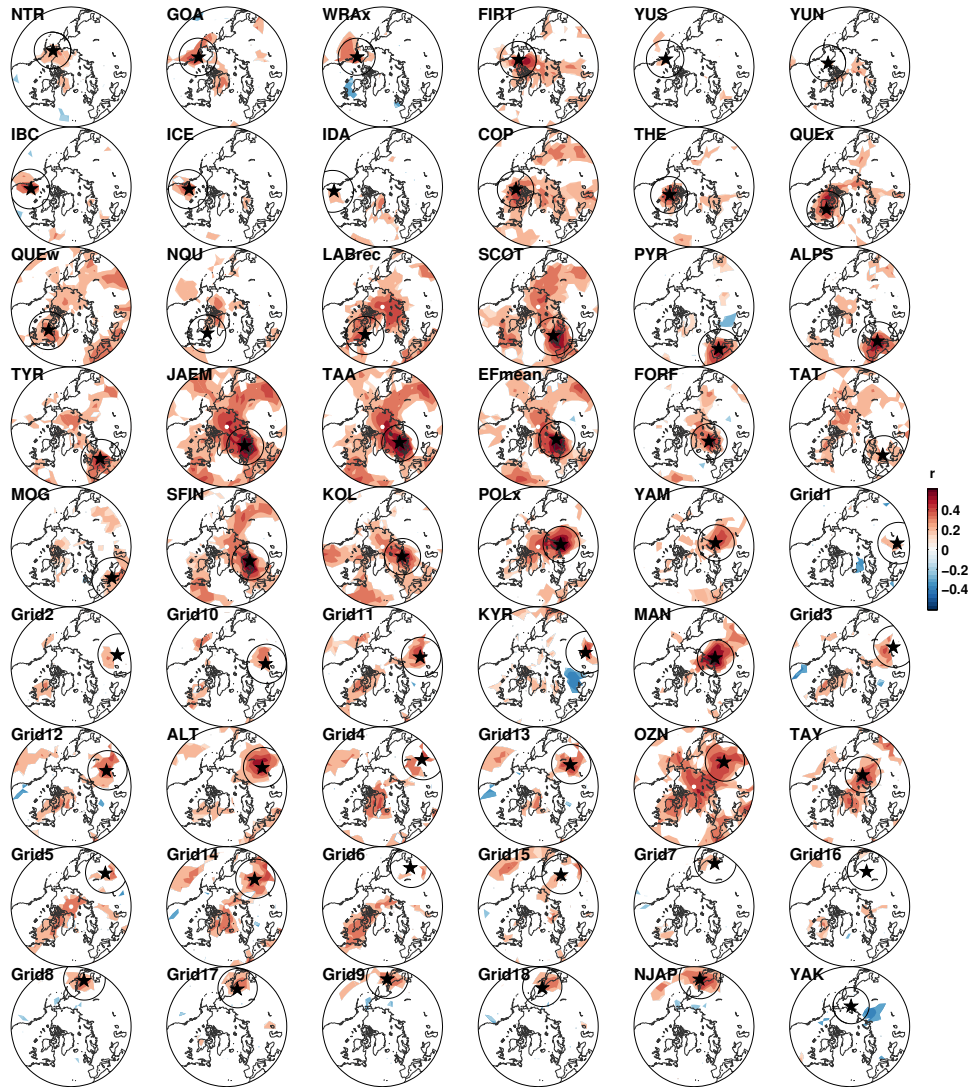


Figure 4: Field correlations between each tree-ring site (indicated by the black stars) and the MJJA mean temperature field from Cowtan and Way (2014). Labels correspond with the site codes from Table 1. Around each site the black range ring indicates a radii of 2000 km. Only Pearson Product Moment correlation coefficients ( $r$ ) significant at  $p < 0.05$  as adjusted for autocorrelation (Trenberth, 1984) are plotted.

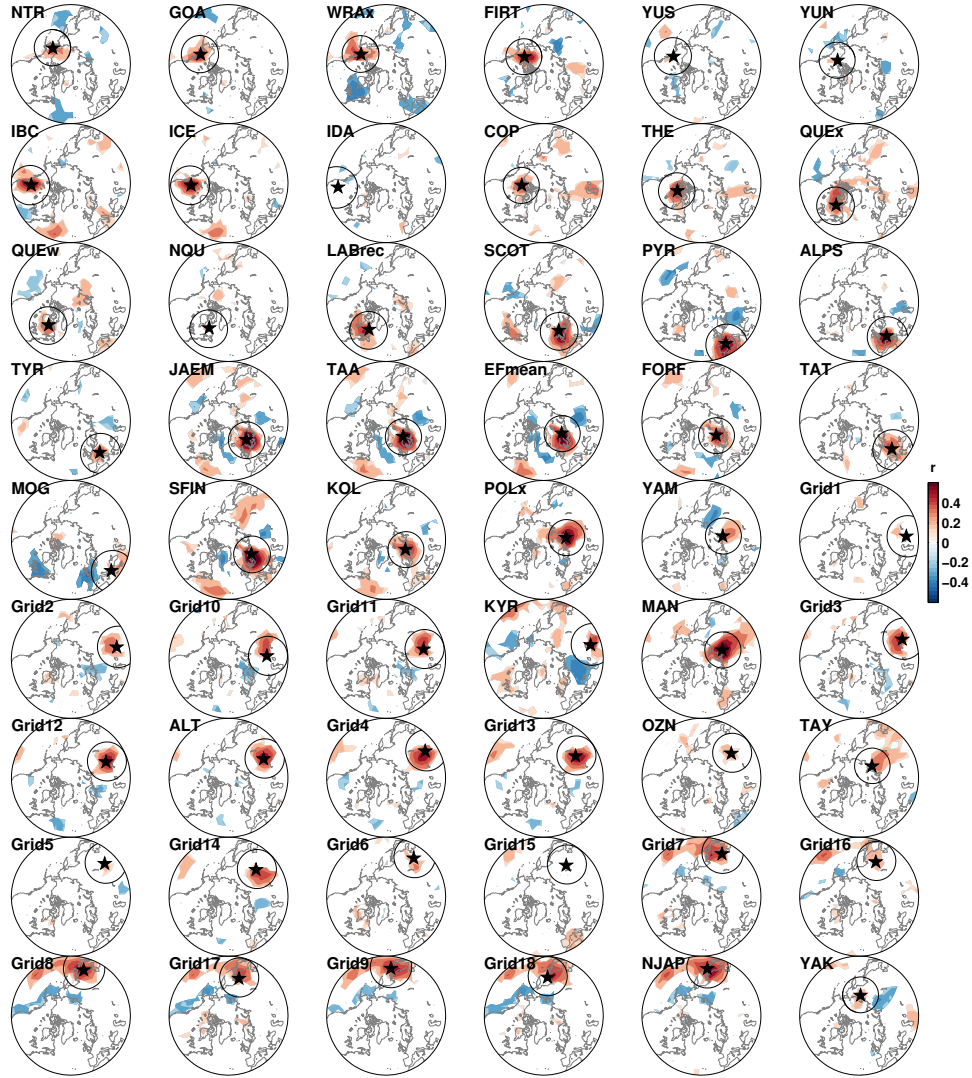


Figure 5: Field correlations between each tree-ring site (indicated by the black stars) and the MJJA mean temperature field from Cowtan and Way (2014) after first-differencing each variable to reduce the influence of common trends. Labels correspond with the site codes from Table 1. Around each site the black range ring indicates a radii of 2000 km. Only Pearson Product Moment correlation coefficients ( $r$ ) significant at  $p < 0.05$  as adjusted for autocorrelation (Trenberth, 1984) are plotted.

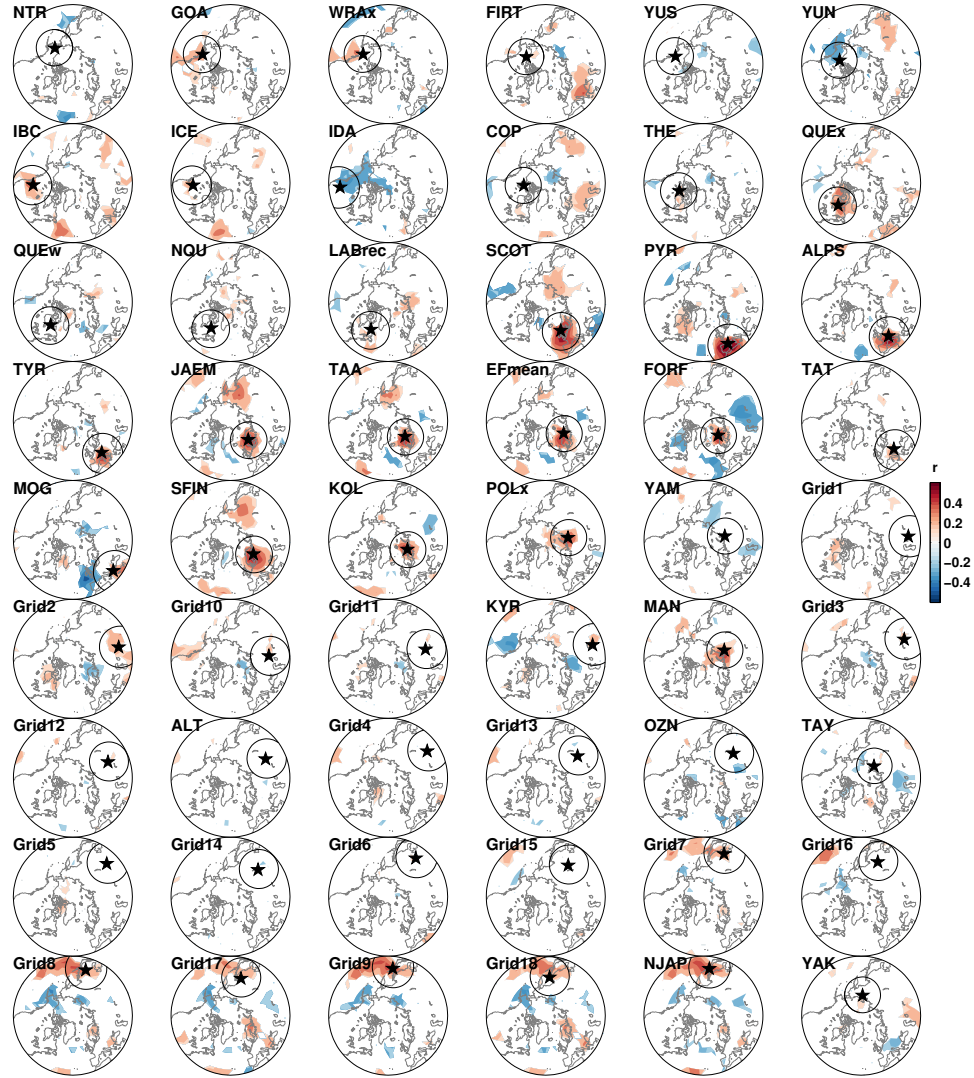


Figure 6: Field correlations between each tree-ring site (indicated by the black stars) and the annual mean temperature field from Cowtan and Way (2014) after first-differencing each variable to reduce the influence of common trends. Labels correspond with the site codes from Table 1. Around each site the black range ring indicates a radii of 2000 km. Only Pearson Product Moment correlation coefficients ( $r$ ) significant at  $p < 0.05$  as adjusted for autocorrelation (Trenberth, 1984) are plotted.

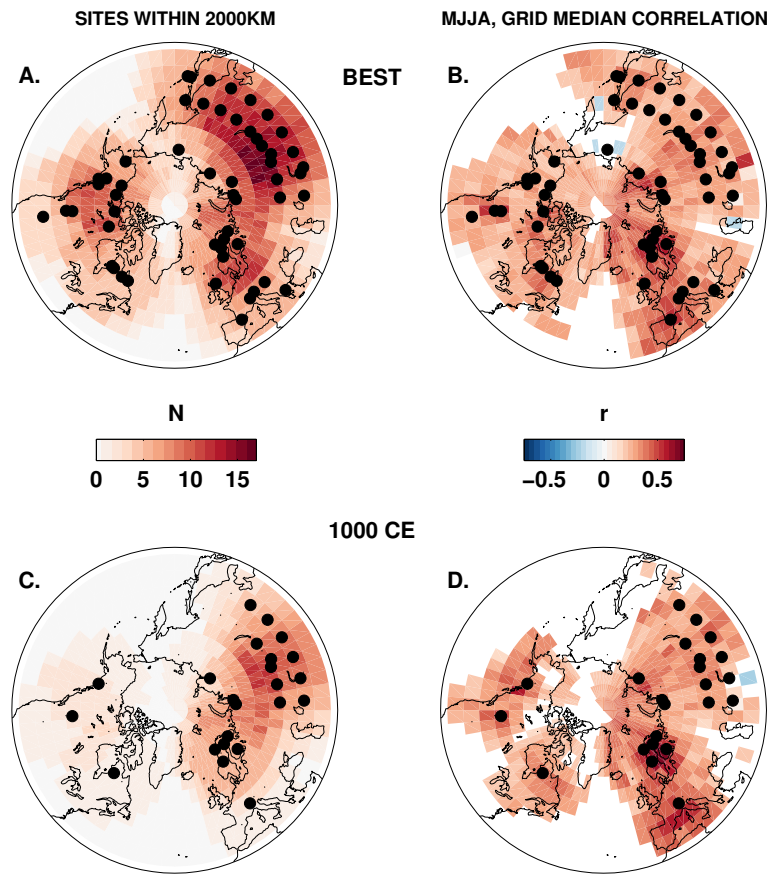


Figure 7: Proxy and target grid characteristics. Panels (A) and (C) show the number of tree-ring sites within 2000km for each grid point in the target field (Cowtan and Way, 2014), during the best replicated (modern) nest (A) and at 1000 CE (C), respectively. Black circles indicate the location of the available tree-ring sites during in each time period. Panels (B) and (D) show the median value at each target field grid point for the Pearson Product Moment correlation between the MJJA temperatures at that grid point and all the tree-ring chronologies within 2000km of that grid, during the best replicated (modern) nest (B) and at 1000 CE (D), respectively (see Schneider et al. (2015)).

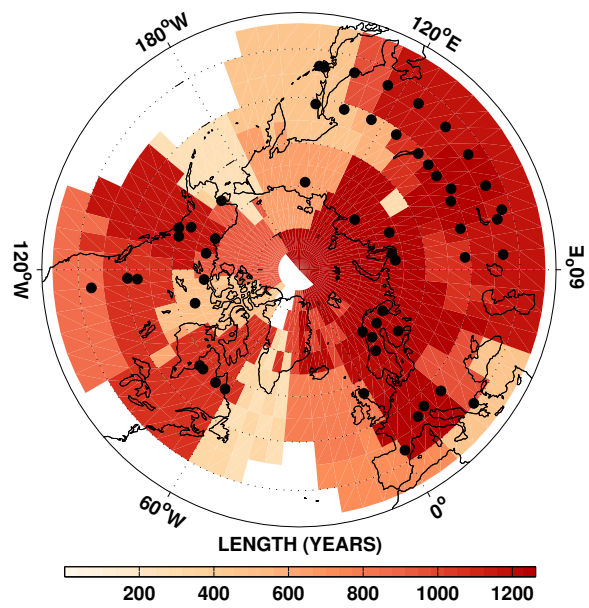


Figure 8: Reconstruction length. The length (in years) of the reconstruction at each target grid point.

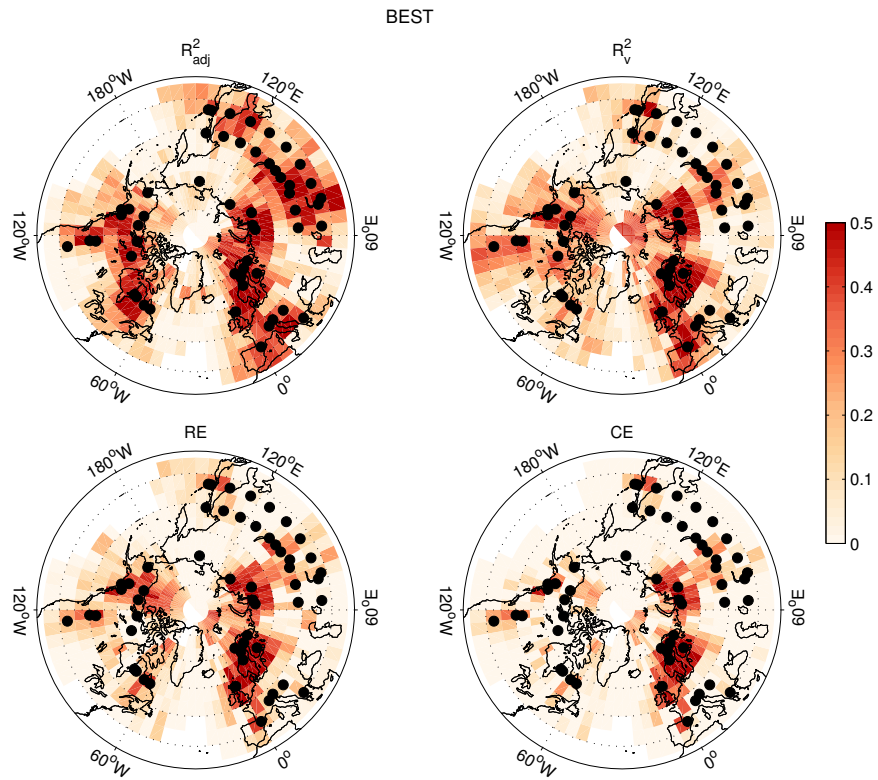


Figure 9: Reconstruction skill for the best replicated (modern) nest. Panels show spatial patterns of skill metrics for the best replicated nest (1750 to 1988 CE), as evaluated for the adjusted calibration  $R^2$ , the validation  $R^2$ , the reduction of error (RE), and the coefficient of efficiency (CE). Available tree-ring sites during this nest are indicated by black circles.

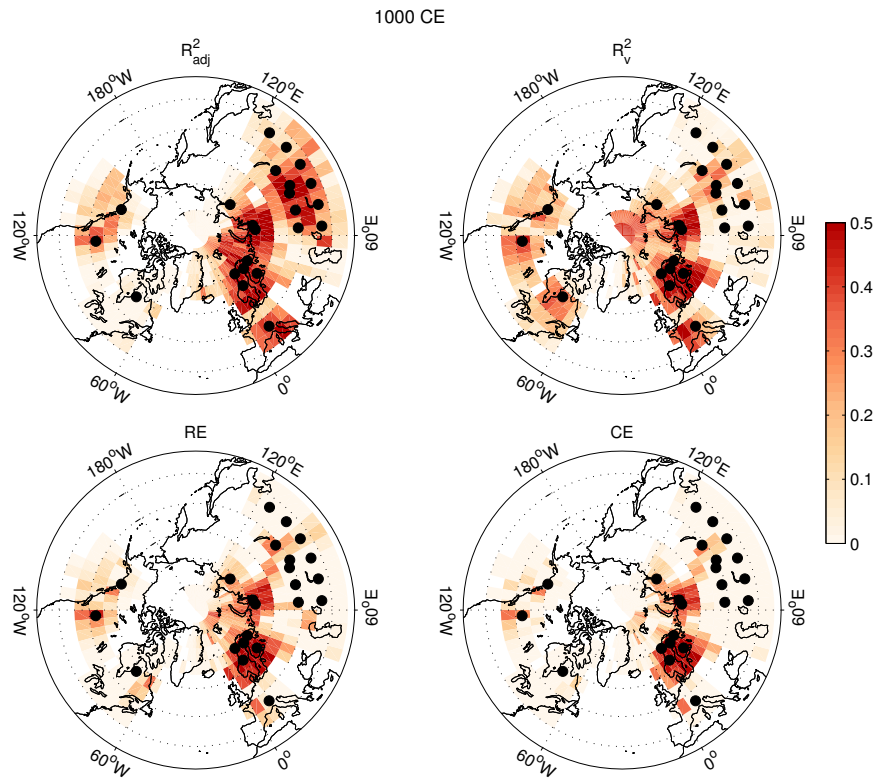


Figure 10: Reconstruction skill at 1000 CE. Panels show spatial patterns of skill metrics for the reconstructed field at 1000 CE in the midst of the Medieval epoch, as evaluated for the adjusted calibration  $R^2$ , the validation  $R^2$ , the reduction of error (RE), and the coefficient of efficiency (CE). Available tree-ring sites at 1000 CE are indicated by black circles.

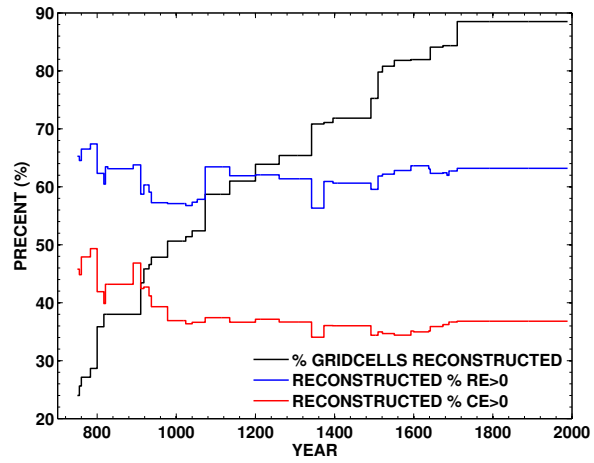


Figure 11: Aggregate temporal reconstruction skill. The percent of all target grid points that are able to be reconstructed for a given year is shown by the black line. For those grid points with a reconstructed value in a given year, the red and blue lines show the percent of those grid points with RE and CE greater than zero.



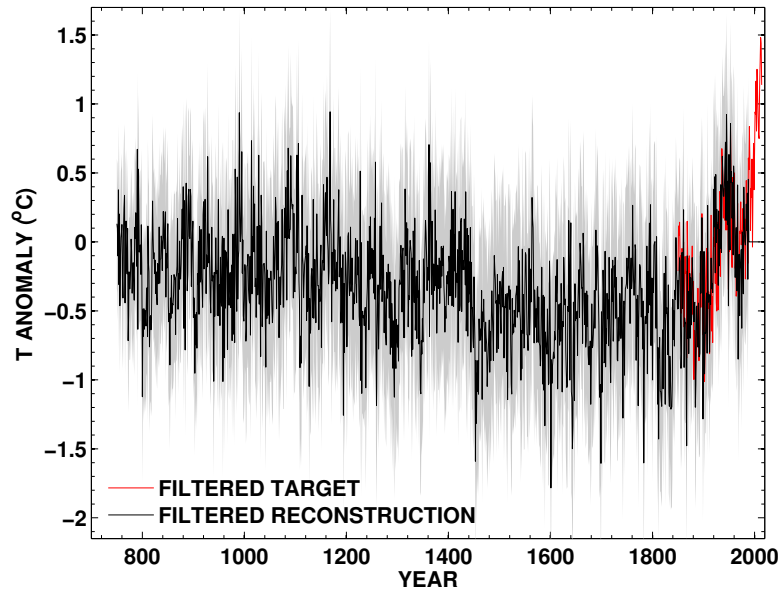


Figure 12: Filtered latitude-weighted mean hemisphere MJJA temperature anomaly reconstruction and target MJJA observational time series. Spatial mean values for both the reconstruction (black) and target (red) MJJA fields are calculated from the set of all grid points that have reconstructed values back to at least 1000 CE and which have an RE score greater than zero at 1000 CE ( $n = 229$ ) and are weighted by latitude. Uncertainty in the reconstruction is indicated by the gray shading, and is calculated as the mean latitude-weighted local mean squared error of validation. The reconstruction and target MJJA temperature series are significantly correlated over their common interval (1850–1988,  $n = 139$ ,  $r = 0.78$ ,  $p \ll 0.001$ ).

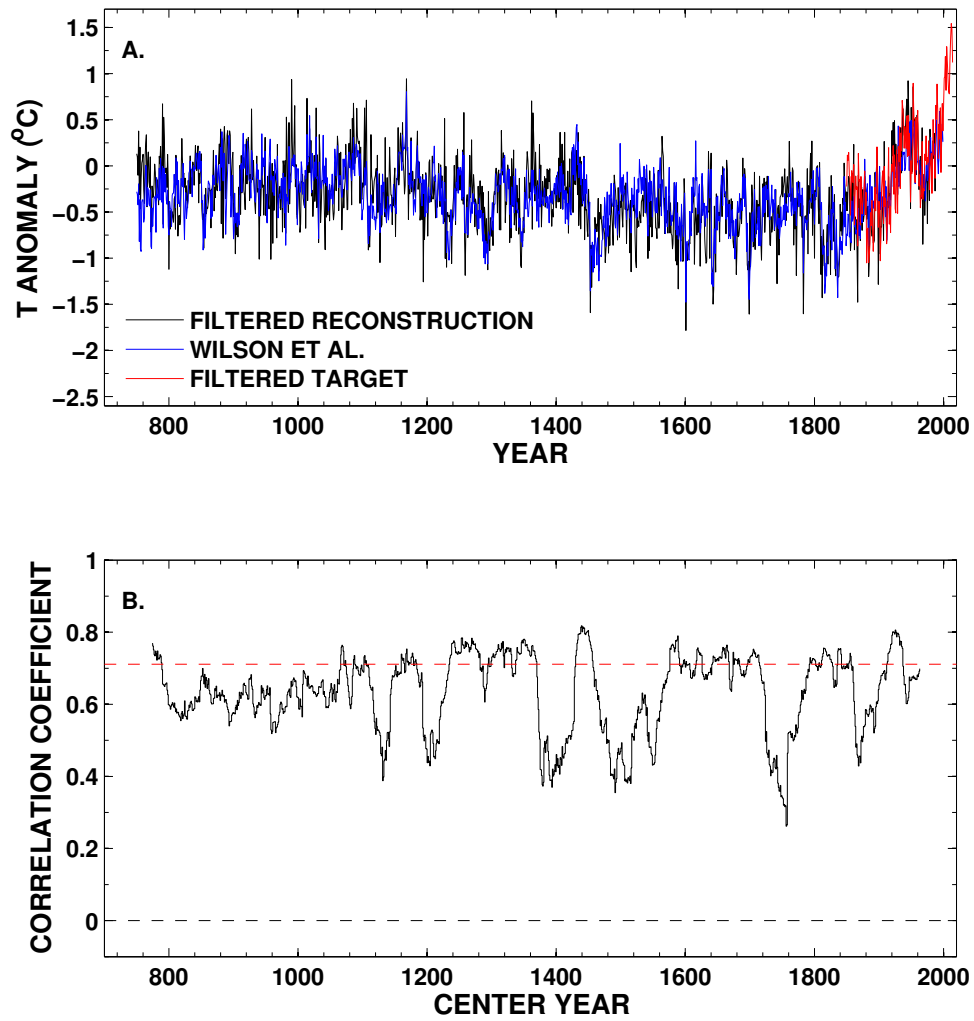


Figure 13: Comparison between time series reconstruction from (blue; Wilson et al., 2016) and the filtered weighted global mean MJJA temperature reconstructed here (black). Time series are shown in (A), and a running correlation (50 year window, 1 year increment) is plotted in (B). The full overall correlation between the two series (750 to 1988 CE,  $r = 0.71$ ) is indicated by the dashed red line in (B).

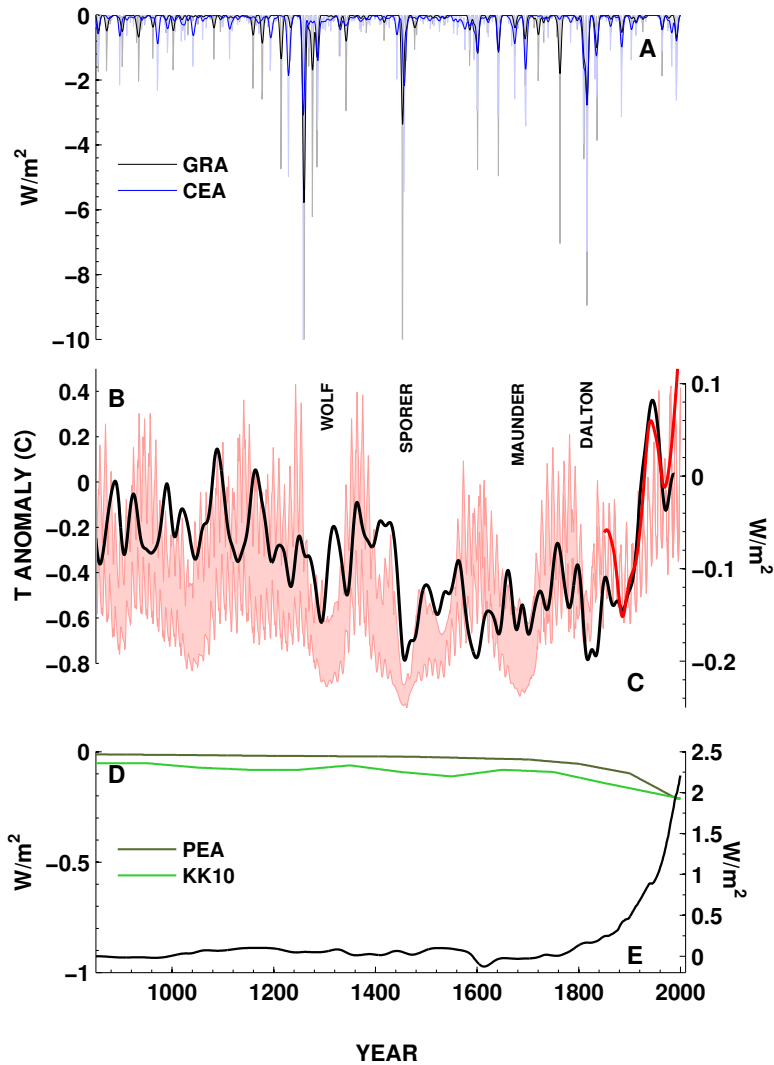


Figure 14: Radiative forcing and reconstructed Northern Hemisphere warm-season temperatures from this study during the last millennium. All forcing series are those compiled by Schmidt et al. (2012) for PMIP3 simulations of the last millennium (version 1.1). (A) Volcanic forcing following Gao et al. (2008) (black/grey, GRA) and Crowley et al. (2008) (blue/light blue, CEA), with individual years as lighter lines and 30-year Gaussian smoothed values in heavy lines. Note that the magnitude of some individual events exceeds the y-axis limits. (B) Northern Hemisphere mean MJJA temperature anomaly time series as described in the text (black line) and corresponding observed temperatures for the same grid points (red line). Here both reconstructed and observed values have been smoothed with a 30 year Gaussian filter. (C) Solar forcing relative to the period 1976 to 2006 CE, with the pink shaded region showing the range of the forcing reconstructions compiled by Schmidt et al. (2012) including Delaygue and Bard (2011), Muscheler et al. (2007), Steinhilber et al. (2009) and Vieira and Solanki (2010). Major solar minima are labeled. (D) Forcing due to land use change from Kaplan et al. (2011) (KK10) and Pongratz et al. (2008) (PEA). (E) Well-mixed greenhouse gas forcing.

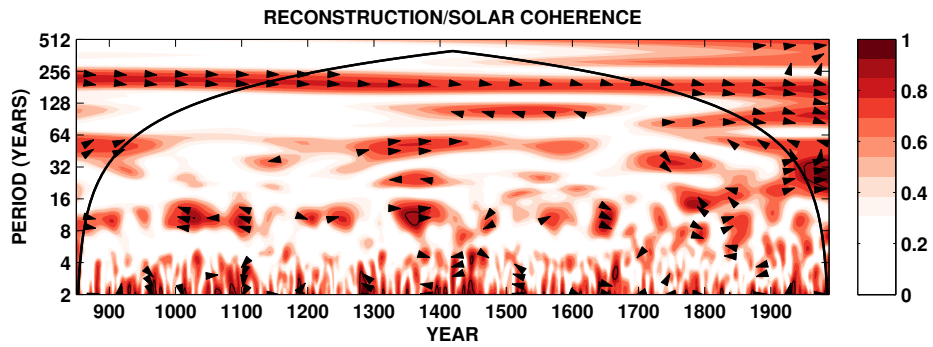


Figure 15: Wavelet coherence (Torrence and Compo, 1998; Grinsted et al., 2004) between our Northern Hemisphere mean MJJA temperature anomaly time series and solar forcing variability from Vieira and Solanki (2010). Arrows indicate the phase of the relationship and for clarity are plotted only where coherence exceeds 0.65. In-phase signals point directly to the right of the plot. Values above the cone of influence (COI; black curve) are potentially influenced by edge effects at that time period and scale.

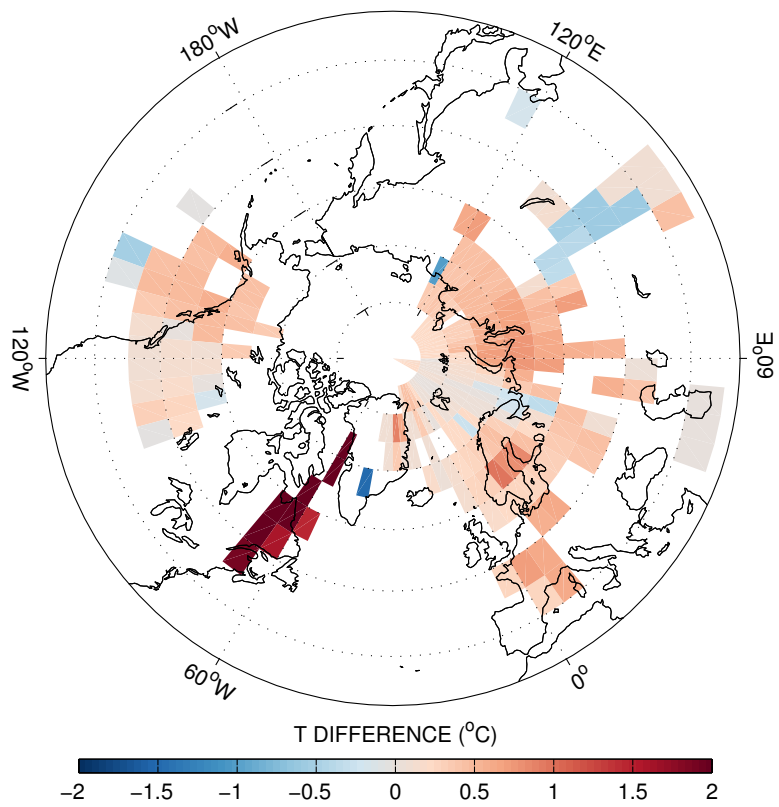


Figure 16: Medieval Climate Anomaly (MCA; 950-1250 CE) vs Little Ice Age (LIA; 1450-1850 CE) mean temperature anomaly fields (MCA-LIA). Only grid points with values reconstructed at  $RE > 0$  at 1000 CE are shown.

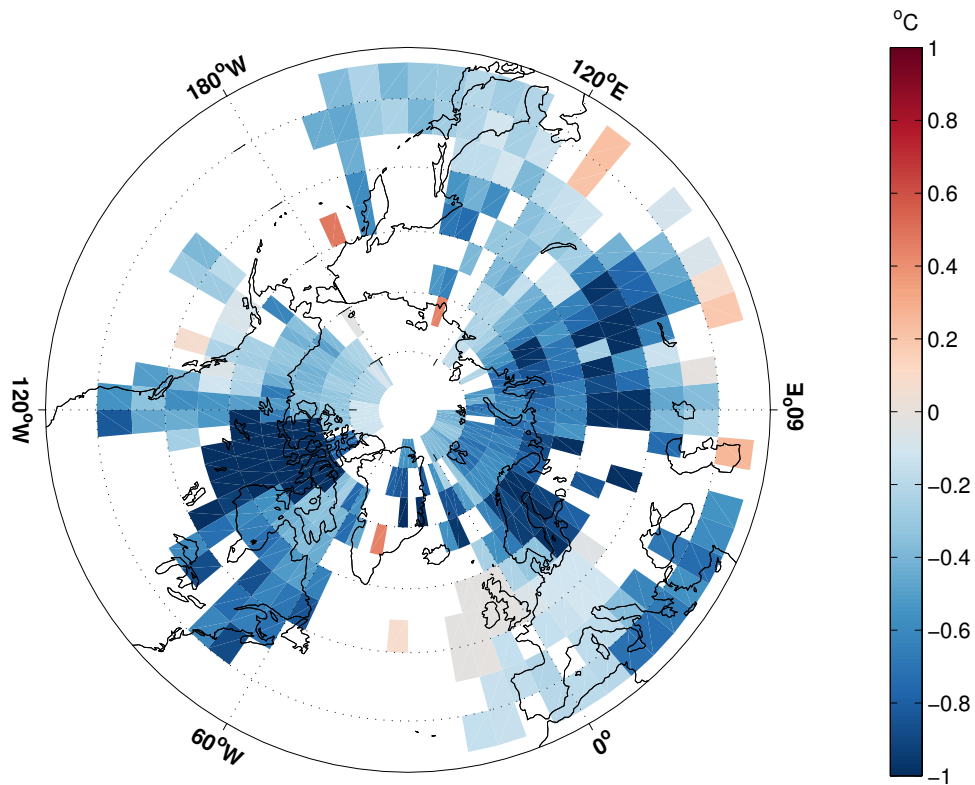


Figure 17: Composite mean reconstructed temperatures following major tropical volcanic eruptions (from Sigl et al. (2015)). Eruption years in the composite ( $n = 20$ ) are those with a global forcing magnitude equal to or larger than that associated with Krakatoa (1884), and include 916, 1108, 1171, 1191, 1230, 1258, 1276, 1286, 1345, 1453, 1458, 1595, 1601, 1641, 1695, 1809, 1815, 1832, 1836, and 1884 CE. Event anomalies are calculated by first subtracting the global field mean over the 3 years prior to the eruption. Only grid points with  $RE > 0$  in an event year are averaged to form the composite and only those grid points with values for at least 6 eruptions are plotted.

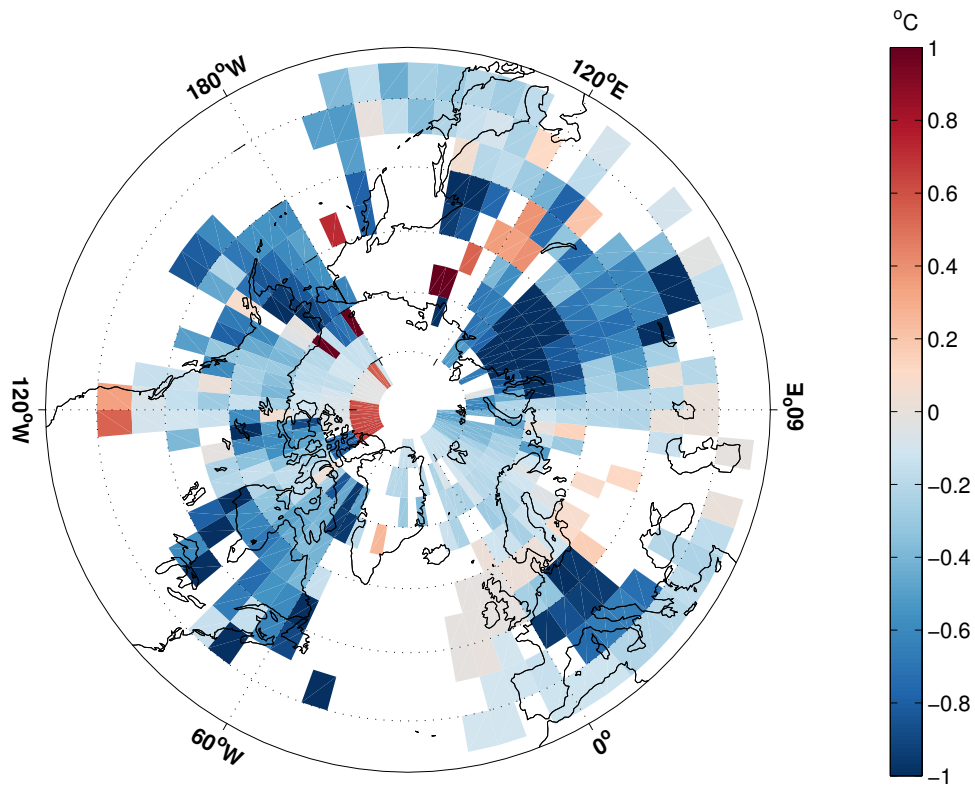


Figure 18: Composite mean reconstructed temperatures following major Northern Hemisphere high latitude volcanic eruptions (from Sigl et al. (2015)). Eruption years in the composite ( $n = 5$ ) are those Northern Hemisphere eruption with a global forcing equal to or larger than the magnitude associated with Katmai (1912), and include 939, 1182, 1210, 1783, and 1912 CE. Event anomalies are calculated by first subtracting the global field mean over the 3 years prior to the eruption. Only grid points with  $RE > 0$  in an event year are averaged to form the composite and only those grid points with values for at least 2 eruptions are plotted.

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