1   # A millennial summer temperature reconstruction for northeastern

2   # Canada using oxygen isotopes in subfossil trees

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10   ## Abstract

11   Climatic reconstructions for north-eastern Canada are scarce such that this area is under-
12   represented in global temperature reconstructions. To fill this lack of knowledge and identify
13   the most important processes influencing climate variability, this study presents the first
14   summer temperature reconstruction for eastern Canada based on a millennial oxygen isotopic
15   series ($\delta^{18}O$) from tree rings. For this purpose, we selected 230 well-preserved subfossil
16   stems from the bottom of a boreal lake and five living trees on the lakeshore. The sampling
17   method permitted an annually resolved $\delta^{18}O$ series with a replication of five trees per year.
18   The June to August maximal temperature of the last millennium has been reconstructed using
19   the statistical relation between Climatic Research Unit (CRU TS3.1) and $\delta^{18}O$ data. The
20   resulting millennial series is marked by the well-defined Medieval Climate Anomaly (MCA; AD
21   1000-1250), the Little Ice Age (AD 1450-1880) and the modern period (AD 1950-2010), and
22   an overall average cooling trend of -0.6°C/millennium. These climatic periods and climatic low
23   frequency trends are in agreement with the only reconstruction available for northeastern
24   Canada and others from nearby regions (Arctic, Baffin Bay) as well as some remote regions
25   like the Canadian Rockies or Fennoscandia. Our temperature reconstruction indicates that
26   the Medieval Climate Anomaly was characterized by a temperature range similar to the one of
27   the modern period in the study region. However, the temperature increase during the last
28   three decades is one of the fastest warming observed over the last millennium (+1.9°C
29   between 1970-2000). An additional key finding of this research is that the coldest episodes
30   mainly coincide with low solar activities and the extremely cold period of the early 19th century
31   has occurred when a solar minimum was in phase with successive intense volcanic eruptions.
32   Our study provides a new perspective unraveling key mechanisms that controlled the past
33   climate shifts in northeastern Canada.

1. **Introduction**

The recently published work of the Intergovernmental Panel on Climate Change (IPCC AR5, 2013; PAGES 2K consortium, 2013) has shown that north-eastern Canada is poorly represented among existing millennial temperature reconstructions in the northern hemisphere. For this reason, a better knowledge of regional past climate variations registered in natural archives is needed. The use of natural archives such as trees, sediment or pollen has permitted the reconstruction of temperature variability at regional, hemispheric and global scales for the past millennium (Hegerl *et al.,* 2007; Mann *et al.,* 2009; Moberg *et al.,* 2005; PAGES 2k consortium, 2013). Although some studies have documented past climatic conditions in northern Canada (Moore *et al.,* 2001; Thomas and Briner, 2009; Luckman and Wilson, 2005; Edwards *et al.,* 2008; Viau and Gajewski, 2009; Gajewski and Atkinson, 2003), only one annually-resolved millennial temperature reconstruction based on tree-ring widths exists for eastern Canada (Gennaretti *et al.,* 2014; summer temperature reconstruction for Eastern Canada, STREC), but none has been based on the isotopic approach. Consequently, obtaining millennial-long, high-resolution temperature reconstructions from additional proxies in north-eastern Canada is important to increase our knowledge of the past climate, and better understand the mechanisms of climate change.

Tree-ring isotope series present the advantage that they generally do not need to be detrended; they retain climatic low frequency variations, and require fewer trees compared to classical dendrological methods (Loader *et al.,* 2013; Robertson *et al.,* 1997; Young *et al.,* 2010). Moreover, oxygen ($\delta^{18}O$) and carbon ($\delta^{13}C$) series have proven their suitability for reconstructing past summer temperatures (Anchukaitis *et al., 2012;* Barber *et al.,* 2004; Daux *et al.,* 2011; Luckman and Wilson, 2005; Porter *et al.,* 2013). Whereas $\delta^{13}C$ series have often been used for long climatic reconstructions, only a few studies have used long $\delta^{18}O$ series (Edwards *et al.,* 2008; Richter *et al.,* 2008; Treydte *et al.,* 2006; Wang *et al.,* 2013). A previous study has already proven that $\delta^{18}O$ is the most suitable isotopic proxy for summer temperature reconstruction in our study region (Naulier *et al.,* 2014; Naulier *et al.,* in press).

In northern Canada, most tree species rarely live more than 300 years (Arseneault *et al.,* 2013). In such regions where old trees are missing, isotopic chronologies can be extended by combining living specimens with subfossil trees preserved in lakes (Boettger *et al.,* 2003;

65 Gagen *et al.,* 2012; Mayr *et al.,* 2003; Savard *et al.,* 2012), and cross-dating stems to

66 determine subfossil tree ages (Arseneault *et al.,* 2013). For the purpose of paleoclimate

67 studies, subfossil stems can be easily extracted and collected from large stocks of drowned

68 subfossil logs in lakes and can be associated with specific edaphic contexts as most

69 specimens are not redistributed in lakes (Gennaretti *et al.,* 2014a).

70 After cross-dating, the development of a robust millennial, isotopic chronology from the

71 combination of living and subfossil stems involves replicating specimens in order to retain the

72 climate variability of the study site (Haupt *et al.,* 2014; Loader *et al.,* 2013a). However, the

73 amount of material available is often a constraint because of the short lifespans of trees, the

74 difficulty to separate single and thin rings and obtaining enough cellulose for isotopic analysis

75 (Loader *et al.,* 2013b; Boettger and Friedrich, 2009). To overcome these problems, different

76 sub-sampling methods have been developed such as tree pooling (McCarroll and Loader,

77 2004; Dorado Liñán *et al.,* 2011), serial pooling of consecutive tree rings within an individual

78 tree (Boettger and Friedrich, 2009), and the "offset-pool plus join-point method" (Gagen *et al.,*

79 2012). This last method has permitted constructing a millennial $\delta^{13}C$ series with annual

80 resolution and high replication, while reducing the sampling efforts and laboratory analyses

81 (Gagen *et al.,* 2012). Moreover, a statistical analysis of this method has confirmed its

82 robustness and possible application for the production of millennial $\delta^{18}O$ series (Haupt *et al.,*

83 2014).

84 The present study aims to produce a new paleoclimatic data set based on tree-ring $\delta^{18}O$

85 series covering the last millennium in northeastern North America. For this purpose, we

86 develop a 1010-years long $\delta^{18}O$ series using a combination of living trees and submerged

87 subfossil stems from one site, and reconstruct the summer maximal temperature. We analyze

88 the main characteristics of the climatic series and evaluate its robustness by comparison with

89 other reconstructed temperature series. Finally, we explore the potential impact of natural

90 forcing (solar radiation and volcanic eruptions) on past climatic variability in north-eastern

91 Canada.

92

93 ## 2. **Materials and methods**

94

95 ### 2.1.   **Study area**

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The study site is located at the center of the Quebec-Labrador peninsula in north-eastern Canada (Figure1A). This area is part of the Precambrian Canadian Shield, mainly constituted of granitic and gneissic rocks. The landscape is characterized by a low altitude plateau (400-600 m), with abundant lakes and wetlands (Dyke *et al.,* 1989). Forests of the area are dominated by black spruce (*Picea mariana* (Mill.) BSP) trees, developed as pure open lichen woodlands on well-drained sites, and spruce-moss woodlands in depressions. Balsam fir (*Abies balsamea* (L.) Mill.) and Tamarack (*Larix laricina* (Du Roi) Koch) also grow in this region. Wildfires are the most important natural disturbances with a rotation period estimated between 250 to 500 years (Boulanger *et al.,* 2012).

The climate is continental and subarctic with short, mild summers and long, cold winters. Environment Canada data (Schefferville station) show that the 1949 to 2010 mean monthly temperature is -22.9°C in January and 13.3°C in July with a mean annual temperature of -3.9°C. Total annual precipitation averaged 640 mm with up to 60% falling in summer (June to September). The mean duration of the frost-free period is 75 days from mid-late June to mid-September. The lakes are generally frozen from mid-October to early June.

The selected lake (L20; 54°56'31'' N; 71°24'10'' W) is part of the large network of lakes sampled by our group (Arseneault *et al.,* 2013; Gennaretti *et al.,* 2014a and b). Ecological and morphological criteria have been developed to identify lakes that present the best potential for millennial-long climatic reconstructions (well-preserved subfossil trees) and large stocks of subfossil logs. These lakes are typified by an abrupt lake/forest transition, as well as log accumulation in the lower littoral zone away from ice erosion and waves (Arseneault *et al.,* 2013, Gennaretti *et al.,* 2014a, Gennaretti *et al.,* 2014c). Lake L20 has an altitude of 483 m and an area of 35.1 ha. It is bordered by open spruce-moss with lichen woodlands growing on well-drained podzolic soil and regular slope. The last severe wildfire occurred at about AD 1590 along the southern section of the studied shore segment and more than 1200 years ago along the northern section (Gennaretti *et al.,* 2014c).

123

124   2.2.   **Tree stem selection and sampling strategy**

125

126 We recently demonstrated that isotopic series from different heights along lakeshore trees
127 provide similar isotopic trends, and indicated that the combination of lakeshore black spruce
128 trees with subfossil stem segments does not introduce artefacts in long $\delta^{18}O$ series, thus
129 permitting their combination for climatic reconstruction (Naulier *et al.,* 2014). In the present
130 study, subfossil stems were selected from a large collection of 586 cross-dated specimens
131 from lake L20, also used in the STREC reconstruction (Gennaretti *et al.,*2014b), based on
132 their excellent degree of preservation (Savard *et al.,* 2012), relatively large ring width (>0.2
133 mm) and their life span (Supplementary material, Table I). The development of a millennial
134 isotopic series requires choosing an appropriate method to preserve both high and low
135 climate frequencies, while limiting analytical efforts. We decided to adapt the "offset-pool plus
136 join-point method" (Gagen *et al.,* 2012) in order to obtain an annual resolution with a
137 replication of five trees for each year.

138 According to this sampling method, one cohort is made by selecting segments from five
139 contemporaneous trees such that each cohort overlaps the next one over five years. In our
140 case, five living trees were selected to construct a modern cohort (CV; AD 1860-2006) and 60
141 well-preserved subfossil stems from the lake floor were used to produce 12 subfossil cohorts
142 (C0 to C11; AD 997-1956; Figure 2). Overall, our cohorts cover between 59 and 111-years
143 and the complete suite of cohorts extends from AD 997 to 2006. Additionally, within every
144 cohort, each tree was divided into five-year blocks which were offset by one year among trees
145 (Supplementary material, Table II). As a consequence, the $\delta^{18}O$ value obtained for a specific
146 year is the mean of the isotopic results from five trees, which represents a triangular
147 centralized nine-year moving average.

148

149 ## 2.3. **Laboratory treatment**

150

151 We extracted α-cellulose sub-samples according to a standard protocol modified for small
152 samples at the Delta-lab of the Geological Survey of Canada (Green, 1963; Savard *et al.,*
153 2012). The extracted α-cellulose was dried at 55°C for 12 hours and sub-samples analyzed
154 using peripherals on-line with gas-source isotope ratio mass spectrometers (IRMS). All
155 material was analyzed for $\delta^{18}O$ values with a pyrolysis-CF-IRMS (Delta plus XL). The
156 analytical accuracy of this instrument was 0.2‰ (1σ), as established by using international

157 standards (IAEA-SO-6 and IAEA-NBS-127). All $\delta^{18}O$ measures are reported in permil (‰)
158 relative to the Vienna Standard Mean Ocean Water (VSMOW). A total of 2192 analyses were
159 produced, with, in addition, 24% of all samples retreated and analyzed to determine the
160 external precision (reproducibility) of the complete procedure (0.2‰, 1σ).

161

162 ## 2.4. **Cohort corrections and climatic reconstruction**

163

164 When a long isotopic series is produced from trees coming from a random assemblage,
165 joining two successive cohorts may be difficult due to the existence of isotopic offsets
166 between cohorts. An approach to overcome this problem is to use the mean of $\delta^{18}O$ values
167 coming from several tree segments from the overlap period between two successive cohorts
168 that permit to estimate a correction factor for the offset (Gagen *et al.,* 2012). In the present
169 study, the construction of the millennial $\delta^{18}O$ series required such an adjustment for some
170 cohorts. We have adopted the "join-point method" proposed by Gagen *et al.* (2012). A join-
171 point (hereafter JP), corresponds to the mean results of 5-year blocks from several trees
172 overlapping between two cohorts. We have used all available dated trees (between 10 and
173 24) to produce the required join points (JP 0 to JP 11), and verified several methods to
174 correct for the offsets between cohorts. Hence, every cohort has been corrected by adding
175 the linear regression calculated between the $\delta^{18}O$ values of its two ends.

176 The integrity of the isotopic signal has been verified elsewhere (Naulier, 2015). The $\delta^{18}O$
177 values of lignin and cellulose have been analyzed for three contrasted climatic periods of the
178 millennium (AD. 1141-1164, 1741-1761 and 1886-1909), as detected in previous studies (e.g.
179 Savard *et al.,* 2012). The cellulose isotopic integrity of the subfossil stems has been
180 confirmed by the similarity of the Δ values between living trees and subfossil stems for the
181 1886-1909 period. In addition, the departure between the $\delta^{18}O$ values of lignin and cellulose
182 ($\Delta = \delta^{18}O_{cellulose} - \delta^{18}O_{lignin}$) exhibit no temporal trend.

183 In previous studies, we have analyzed the relationships between $\delta^{18}O$ series of five living
184 trees and various climatic parameters (temperature, precipitation, vapor pressure deficit, etc.).
185 We have established that $\delta^{18}O$ series of black spruce stems sampled at an annual resolution
186 from boreal lakeshore were significantly correlated with June-August (JJA) maximal

187 temperature (Tmax; r=0.54). We have also determined by statistical analysis (Buishand test,
188 Buishand, 1982) that the summers became warmer after 1975 and that the growing season
189 duration and degree-days have increased importantly during the last decade (2000-2010).
190 Moreover, we have demonstrated that during this decade the growing season started sooner
191 and finished later than before, changing the relationship between JJA maximal temperature
192 and $\delta^{18}O$ series (divergence). In other words, this JJA Tmax and $\delta^{18}O$ series relation is stable
193 and strong between 1930 and 2000(r_{mean}= 0.54; 1930-2000), but not after (AD 2000-2010;
194 Naulier *et al.,* 2014; Naulier *et al.,* in press). Therefore, in the present study, we excluded the
195 last decade (divergent years) when calibrating the $\delta^{18}O$ series on temperature data which is
196 assumed to be non-representative of the temperature variation over the last century. The
197 $\delta^{18}O$ series of subfossil cohorts are filtered on 9 years; we have made the choice to pass a 9-
198 year centered-filter on the JJA maximal temperature CRU TS 3.1 in order to use series all
199 treated in the same way for the reconstruction.

200   In a first step, a simple linear regression and a linear-scaling model were calibrated over
201 the entire 1930-2000 period with climatic data. Climate data from the 1900-1929 period were
202 excluded because no meteorological station was then operating at less than 300 km from the
203 study site. The climatic series was separated into two equal periods (AD 1930-1970 and
204 1971-2000; Table 1) in order to test the robustness of the two calibration models, using the
205 non-first-differenced reduction of error (RE), the coefficient of error (CE), the raw mean
206 squared error (RMSE) and the coefficient of determination (r²). The linear regression and the
207 linear-scaling calibration procedures resulted in somewhat different temperature
208 reconstructions of similar robustness with similar RE, CE, r² and RMSE coefficients. In both
209 cases, the model residuals satisfy the standard linear regression assumptions of normality,
210 variance and autocorrelation (not shown), but cannot reproduce all attributes of the measured
211 data. We therefore tested the possibility of averaging the two model results, and this option
212 gave the best reproduction of the measured Tmax. Consequently, we averaged results from
213 the two reconstructions in order to obtain one robust reconstruction (Table 1; Supplementary
214 material, Table III).

215   Then, i-STREC was compared to the only other regional temperature reconstruction
216 (STREC; Gennaretti *et al.,* 2014), which is built from ring width data from 6 lakes, including
217 our site, and with reconstructions based on tree rings from another boreal region
218 (Fennoscandia, Helama *et al.,* 2002; Figure). We also compared i-STREC with independent

219 temperature reconstructions based on other natural archives from North America and the
220 Arctic region (Thomas and Briner, 2009; Kobashi *et al.,* 2011; Luckman and Wilson, 2005;
221 Vinther *et al.,* 2009, Figure 5). Most published reconstructions are based on mean
222 temperatures, except our reconstruction and the one from the Canadian Rockies, which are
223 based on summer maximal temperatures. The influence of climatic forcings was evaluated
224 through the comparison of i-STREC with time series of sulfate emission from volcanic origin
225 (Crowley *et al.,* 2012) and solar radiation series (Bard *et al.,* 2003; Figure 6). The durations of
226 the solar minima have been determined according to existing estimations of solar radiation
227 (e.g., Bard *et al.,* 2003).

228

229 3. **Results and discussion**

230

231 3.1.   **Development of δ^{18}O chronology**

232

233 As the purpose of the reconstruction was to identify contrasted periods and important
234 temperature changes over the last millennium, the choice of the sampling method proposed
235 by Boettger *et al.* (2009) was relevant because it allows for reconstruction of climatic
236 parameters at an annual resolution with a replication of five trees per year. The range of δ^{18}O
237 values of trees is between 19.5 and 22.0‰, and the largest δ^{18}O differences among trees
238 within a junction is obtained for JP5 (3.8‰; Figure 3A). This large inter-tree variability can be
239 explained by a combination of causes, including different growing locations along the
240 lakeshore which influence the water supply of trees (Figure 1B), and inter-tree metabolic
241 variability (0.5‰ in Naulier *et al.,* 2014). Such variability confirms the need to take a large
242 number of trees for a millennial reconstruction in order to capture the site signal (Loader *et al.,*
243 2013a).

244   The δ^{18}O values at intersections between two successive cohorts and of the JP δ^{18}O
245 means are surprisingly matching in most cases (except at the C8/C9 and C3/C4 junctions;
246 figure 3). These observations suggest that offset correction between cohorts is not always
247 necessary. However, we have determined that a modification of the JP adjustment procedure
248 published by Gagen *et al.* (2012) would optimize the correction while conserving the isotopic

249 variability and trends over the millennium (Naulier, 2015). Hence, we have used the mean of

250 $\delta^{18}O$ values of JP from overlapping cohorts to calculate the required adjustment, this mean

251 being considered as "the adjustment value". Correcting cohort $\delta^{18}O$ series with this method

252 increases the number of trees considered (20 to 33 trees instead of 10 to 23 trees if only JP

253 are used). After correction, the mean of the millennial $\delta^{18}O$ series is 20.8‰. The strong

254 correlation between the $\delta^{18}O$ series of living trees and subfossil stems (r^2=0.70) over their

255 overlapping period (1860-1956) confirms the isotopic integrity of the subfossil stems, and

256 ensures that the climatic reconstruction can be performed over the rest of the millennial $\delta^{18}O$

257 series.

258   Although the cohort sampling method presents many positive points, it is important to

259 highlight some of its flaws. Indeed, the sampling strategy produces a $\delta^{18}O$ series smoothed

260 with a centered 9-year filter. This smoothing leads in some cases to series requiring more

261 precaution than non-smoothed series before they can be interpreted or used. For instance,

262 the calibration of our smoothed $\delta^{18}O$ series required a centered 9-year filtering of the climatic

263 series. Consequently, correlations between isotopic and climatic series are improved by

264 smoothing due to the sampling method. Nevertheless, these correlations represent solid and

265 real links, and do not create artefacts (see also section 3.2.1, and Naulier *et al.,* 2014).

266

267 3.2.   **Model validation and millennial climatic reconstruction**

268

269   3.2.1. Model validation

270

271   We have compared two methods of sampling and production of the ring series from living

272 trees prior to calibration: (1) separation at an annual resolution without pooling (e.g., Haupt *et*

273 *al.,* 2013); and (2) sampling with the cohort approach. We have found that the second was

274 best suited because it slightly improved the correlation between $\delta^{18}O$ series and maximal

275 temperatures (r^2=0.64 vs 0.54 with annual resolution), and it allows using $\delta^{18}O$ series

276 compatible with the one used for the millennial subfossil series (9-year moving average).

277 Consequently, the $\delta^{18}O$ series of the living-tree cohort (CV) was used to calibrate the model

278 and to provide a robust reconstruction as it show a strong correlation with the new CRU

279 series (r^2=0.64) for the entire calibration period (1930-2000), which confirmed that the

280　reconstruction of past temperature with the average $\delta^{18}O$ series is suitable (Figure 4A). The
281　RMSE is 0.26, with a RE and a CE of 0.60. However, it appears that the calibration and
282　verification coefficients are significantly changing depending on the period selected for the
283　statistical analysis. This observation implies that the correlation between Tmax and $\delta^{18}O$
284　series is not stable over the last century even if the correlation stays significant (Naulier *et al.*
285　in press).

286　　These statistical results confirm that the summer temperature reconstructed based on $\delta^{18}O$
287　values (i-STREC) are representative of the natural variability that existed in north-eastern
288　Canada.

289

290　　　　　3.2.2. Millennial temperature trends in north-eastern Canada
291

292　　The i-STREC shows a 0.6°C decrease of maximum summer temperature over the past
293　millennium (Figure 4B), whereas a millennial 0.2°C cooling is roughly estimated for the mean
294　temperature of northern hemisphere (PAGES 2k consortium, 2013). However, our local
295　temperature decrease is in the same order of magnitude as the decrease of summer-months
296　mean temperature reconstructed using pollen in the North American tundra (Viau *et al.,*
297　2012), and in northern regions of high latitudes (sediment, tree-ring widths, ice core in
298　Kaufman *et al.,* 2009 and tree-ring widths in Esper *et al.,* 2012). This temperature decline
299　over the last millennium is generally attributed to orbital forcing (PAGES 2k consortium,
300　2013).

301　　The reconstruction suggests that the maximum summer temperature has varied from a
302　maximum of 17.3°C around AD 1008-1010 to a minimum of 14.8°C around AD 1670-1674.
303　The twentieth century was generally cold (mean of 16°C) with an abrupt warming trend during
304　the last three decades (+0.2°C/10 years between AD 1900-1980 and +0.8°C/decade between
305　AD 1980-2010; CRU TS 3.1 data; Figure 4A). Furthermore, two major climatic episodes were
306　also revealed by i-STREC: a warm period during the eleventh and twelfth centuries (mean of
307　16.5°C between 1000 and 1250) and a cold period from the early fifteenth to the end of the
308　nineteenth centuries (mean of 15.8°C between 1450 and 1880; Figure 4B). These periods are
309　in agreement with the general knowledge of the temperature trends observed globally for the
310　last millennium and correspond to, the Medieval Climate Anomaly (MCA) and the LIA,

311 respectively (IPCC 2013, PAGES 2k consortium, 2013). Based on i-STREC data, we
312 associate these two climatic episodes to the ~ AD 1000-1250 and ~ AD 1450-1880 time
313 periods, respectively.

314

315   3.2.3. Evidences of contrasted climatic periods
316

317   The high summer temperatures of the 11th century (AD ~ 1000-1250; Figure 4B) coincide
318 with peaks previously observed in our study area based on tree-ring width (Gennaretti *et al.,*
319 2014b; STREC), as well as in Greenland ice cores (Kobashi *et al.,* 2011; Vinther *et al.,* 2009,
320 2010), tree-ring series from the Canadian Rockies (Luckman and Wilson, 2005) and
321 Fennoscandia (Helama *et al.,* 2002; Figure 5), and large-scale reconstructions (Mann *et al.,*
322 2009, Ljungqvist *et al.,* 2012, Kaufman *et al.,* 2009, Trouet *et al.,* 2013). Several hypotheses
323 on the forcing of this warm anomaly have been proposed, including a prolonged tendency
324 towards a positive-phase of the North Atlantic Oscillation (NAO; Trouet *et al.,* 2009; Trouet *et*
325 *al.,* 2012) or a synchronicity between La Niña phase and a warm phase in the Atlantic
326 Multidecadal Oscillation (AMO; Feng *et al.,* 2011; Mann *et al.,* 2009). In northeastern Canada,
327 the NAO has an important impact on winter temperatures but not for summer (Hurrell *et al.,*
328 2003). In contrast, the AMO influences spring and summer temperatures (Fortin and
329 Lamoureux, 2009) and is partly responsible for the recent sea surface temperature warming
330 of northeastern Canada (Ding *et al.,* 2014). However, the state of the AMO at the beginning of
331 the millennium and its potential influence on climate during the MCA are unknown. Recently,
332 Sicre *et al.* (2014) have demonstrated that during the MCA, the Northern Annular Mode
333 (NAM) was effective concomitantly with a strong ice-loaded Labrador Current (LC). This
334 combination could be responsible for a decrease of fresh air from the Arctic to eastern
335 Canada, and consequently, for an increased temperature along the continent.

336   Following the MCA, i-STREC emphasizes a cold period between ~ AD 1450 and 1880,
337 which can be attributed to the Little Ice Age (LIA; Figure 4B). It is worth noting that a short
338 warming phase occurred between 1510 and 1590 (Figure 4B). Such warm phase also
339 occurred at the eastern Canadian treeline and included the expansion of upright tree growth
340 forms in lichen-spruce woodland (Payette *et al.,* 1989). Overall, the LIA is recorded in several
341 Northern hemisphere temperature reconstructions based on various proxies, even if its length

342   vary among regions (PAGES 2k consortium, 2013). At the hemispheric scale, the LIA is a
343   well-documented cool period (Moberg *et al.,* 2005; Mann *et al.,* 2009; Hegerl *et al.,* 2007), and
344   several causes may have concurred to trigger its occurrence, including a succession of strong
345   volcanic eruptions (Crowley, 2000; Miller *et al.,* 2012, Gennaretti *et al.,* 2014b), millennial
346   orbital cooling (Kaufman *et al.,* 2009; Esper *et al.,* 2012), and low solar radiation (Bard *et al.,*
347   1997).

348   It is without surprise that the MCA was warmer than the LIA (+0.4±0.3°C) in the L20 area.
349   We have furthermore compared maximal temperature of the MCA with the modern warming,
350   following the approach used in the IPCC report (Masson-Delmotte *et al.,* 2013). The running
351   50-years averages between 1000 and 1100 were higher (+0.2±0.1°C) than the measured
352   temperature of the last 50 years (1959-2009). It is worth nothing that when considering 30-
353   years running averages, the 1979-2009 period appears to be the warmest (average of
354   +0.6°C±0.4) over the last millennium. Considering these results, one can infer that the MCA
355   and recent warming show similar average maximal temperature in the study area.

356   These results contrast somewhat with Northern hemisphere temperature reconstructions
357   that have determined that the mean annual temperature of the modern period was the
358   warmest in northern Canada (Mann *et al.,* 2009; Ljungqvist *et al.,* 2012). Indeed, the data
359   available for these hemispheric reconstructions in the last IPCC report are scarce for north-
360   eastern Canada (Viau *et al.,* 2012). Clearly, the i-STREC results indicate that the MCA in
361   northeastern Canada has been as warm as the modern period of the last millennium (Figure
362   5). The similarities between MCA and the modern period were expected considering that the
363   MCA has been widely studied for its similarities with the modern warming period.
364   Nevertheless, the causes that triggered these similar climatic periods are likely different (i.e.,
365   Landrum *et al.,* 2013; Way and Viau, 2014). Indeed, if the MCA is solely controlled by natural
366   processes, it seems that the warming during the modern period results from a combination of
367   natural and anthropogenic causes (i.e., Mann *et al.*, 2009; Viau *et al.*, in press). By using
368   empirical statistical modeling and global climate models for the 1881-2011 period in Labrador,
369   Way and Viau (2014) have shown that up to 65% of the variance in annual air temperature
370   was explained when also including anthropogenic forcing in the model. Even if summer
371   temperature has increased at a lower rate compared to annual air temperature in Labrador,
372   the observed warming (+1.9°C) between 1970 and 2000 in the region of L20 is one of the

373 fastest over the last millennium. In the next decades, if warming continues at this rate,
374 temperature will reach a new record for the last millennium.

375

376   3.2.4. Climatic forcings of the last millennium

377

378 Contrary to previous studies, our isotopic series does not emphasize an abrupt LIA onset in
379 response to volcanic forcing such as the AD 1257 Samalas event (Lavigne *et al.,* 2013;
380 Figure 6). Instead, our data suggest that solar radiation was the most influential forcing on
381 Tmax changes in the studied region. Indeed, the most important cooling phases of i-STREC
382 occurred during periods of low solar activity like the Oort (AD 1040-1080), Dalton (AD 1800-
383 1850), Maunder (AD 1600-1650) and Spörer (AD 1410-1480) minima (Figure 6). A simple re-
384 sampling method involving 1000 iterations of re-sampling (bootstrapped) has demonstrated
385 that the low temperature periods were always associated to low solar radiation periods
386 (p<0.05). Proposing that solar radiation represents an important control on temperature in
387 north-eastern Canada is in agreement with the hypothesis that the solar forcing was important
388 during the last millennium (AD 1000 to ~1900), except during the modern period
389 (Breitenmoser *et al.,* 2012; Keller *et al.,* 2004), implying that recent anthropogenic impact is
390 the main control at that time. However, even i-STREC is not significantly influenced by
391 volcanism, as determined by superimposed epoch analysis (results not shown), the possibility
392 that successive strong volcanic eruptions combined with solar minima could have contributed
393 to the important LIA cooling in Northeastern Canada cannot be discarded. Strong eruptions
394 and solar minima coincide during the Maunder minimum with the Kuwae eruption, and the
395 Dalton minimum with the unknown (1809), Tambora (1815) and Cosiguïna (1835) eruptions.
396 The role of coinciding natural forcings is also invoked in other paleoclimatic studies that have
397 compared northern hemisphere reconstructions with solar radiation series (e.g., Bard *et al.,*
398 2006; Breitenmoser *et al.,* 2012; Crowley *et al.,* 2000; Lean *et al.,* 1995; Shindell *et al.,* 2003).
399 These studies have shown that temperature changes were largely due to solar forcing alone
400 during the first part of the last millennium, and to volcanic and solar forcings (i.e,
401 Breitenmoser *et al.,* 2012), or to volcanic eruptions (i.e., Crowley *et al.,* 2000; Keller *et al.,*
402 2004) during the end of the LIA (after 1600). In addition, Tingley *et al.* (2014) have
403 demonstrated, by analyzing the ring density in trees growing at high latitude, that the trees

404 recorded not only volcanic eruptions but also variations in light intensity. This finding indicates
405 that both isotopes and density of trees can record changes in solar radiations.

406   The other temperature reconstruction produced for the studied region (STREC) contains a
407 stronger volcanic signal than i-STREC (Gennaretti *et al.,* 2014b). Considering that the two
408 reconstructions are statistically robust, we can assume that they both reflect real trends. In
409 addition, calibrating STREC using the same approach than for i-STREC (i.e. calibration on
410 maximum temperature over the 1930-2000 time period) indicates that methods cannot
411 account for the main differences between the two reconstructions. Consequently, differences
412 in thermal trends between i-STREC and STREC must be caused by their respective
413 sensitivity to climatic triggers and control mechanisms, ring width and $\delta^{18}O$ values. The first
414 important point to bear in mind is that temperature is the main control on changes in ring
415 widths and $\delta^{18}O$ values, but not the only one. Consequently, other climatic parameters (i.e.,
416 precipitations, vapor pressure deficit) have also generated short and medium variations on
417 the two series, creating an important "climatic noise" at high and medium frequencies,
418 possibly explaining the differences between the reconstructions (Naulier *et al.,* 2014).
419 However, the ring width and $\delta^{18}O$ series used to generate STREC and i-STREC display
420 similar long-term climatic trends. This last point is quite important, considering that our main
421 purpose was to identify long climatic tendencies over the last millennium in northeastern
422 Canada.

423   The second important aspect to consider is that the temperature-linked processes
424 responsible for the variations of ring widths and $\delta^{18}O$ values slightly differ. In the studied
425 region, rings widths are directly influenced by photosynthetic rates, which generally increase
426 with ambient temperatures. In addition, volcanic aerosols blocking light after a major volcanic
427 eruption may also reduce ring growth concomitantly to reduced temperature, explaining the
428 strong influence of major volcanic events on ring width. In contrast, one of the main controls
429 on the final tree-ring $\delta^{18}O$ values is the temperature prevailing regionally during cloud mass
430 distillation, as registered in the raindrop signal and transferred to the source water in soils,
431 then through the root system, to the tree. Moreover, the temperature effects on fractionation
432 during distillation and precipitation (Rayleigh process) are not limited to a temperature range,
433 and may record temperature lows modulated by solar minimums. When strong volcanic
434 events are combined with minimal solar radiations, the strong influence on regional
435 temperature is therefore detected by $\delta^{18}O$ values of rain drops. These key differences in

436   mechanisms controlling temperature recorded in ring widths and $\delta^{18}O$ values imply that the

437   two proxies may emphasize forcings in a complementary way.

438

439   As a summary, it appears that ring widths or $\delta^{18}O$ series have strengths and weaknesses as

440   proxy of past climatic conditions. However, the climatic data that can be extracted from the

441   two series can generate complementary information, permitting to highlight several climatic

442   forcings and identify the main regional control on past, present and future temperatures.

443   Nevertheless, there are still needs for further understanding the differences between

444   processes influencing isotopic assimilation and ring-width growth. Such information would be

445   useful for future climatic reconstruction using a multi-indicator approach.

446

447 4. **Conclusion**

448

449   1. The cohort sampling method allows reconstructing climatic variability of medium and low

450   frequencies by using fewer samples than other sampling methods, with a high temporal

451   resolution and analytical replication. Our adjustment of the method of joining cohorts, the JP-

452   adjustment method, permits the preservation of $\delta^{18}O$ variability between segments of trees

453   without biasing the millennial $\delta^{18}O$ series.

454   2. The combination of two statistical models (linear scaling and simple linear regression) has

455   permitted an adequate reproduction of the measured regional temperature, and allowed

456   reconstructing maximum temperature over the last millennium.

457   3. i-STREC is complementary to the only other reconstruction in the study region (STREC,

458   based on tree-ring width). These two reconstructions should be combined within a multi-

459   parameter approach to increase the proportions of variance explained.

460   5. i-STREC suggests that the main climatic forcing at play during the last millennium in the

461   studied region was solar activity, but we remain cautious because we base this hypothesis

462   solely on an apparent correlation between reconstructed Tmax and the curve for solar

463   radiations. Clearly, coldest episodes in the L20 area coincide with low solar radiation (Oort,

464   Spörer, Maunder and Dalton), with the exception of an episode in the nineteenth century,

during which low solar radiations (Dalton minima) were combined with two successive and strong volcanic eruptions (unknown 1809 and Tambora 1815 eruptions).

6. Overall, i-STREC shows that the Medieval Climate Anomaly (997-1250) was characterized by a temperature range similar to the one of the modern period in the study region. However, the sudden and rapid temperature increase during the last three decades is one of the fastest over the last millennium (+1.9°C between 1970 and 2000) and if this rapid warming rate persists, the future climate in northeastern Canada may become an issue of concern.

Acknowledgments

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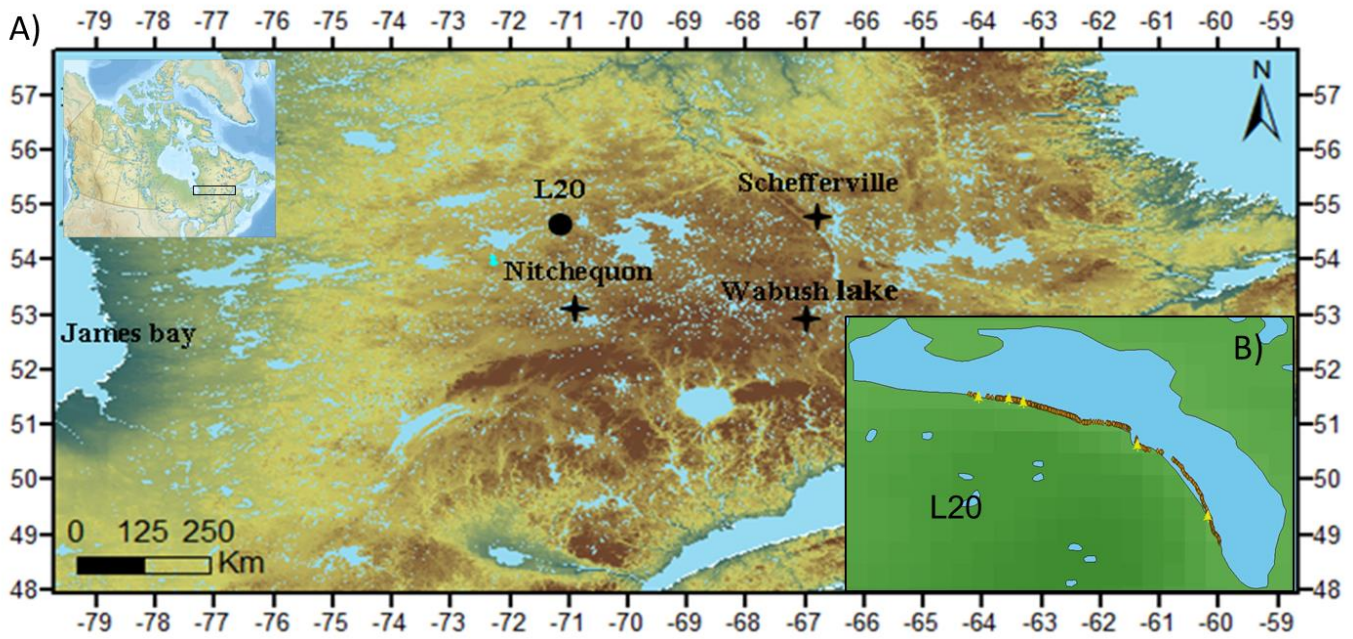
Figures and table

Figure 1. A) Site location (black circle) and meteorological stations (black stars). B) Representation of lake L20, also illustrating the sampling location of subfossil stems (brown marks) and living trees (yellow marks).
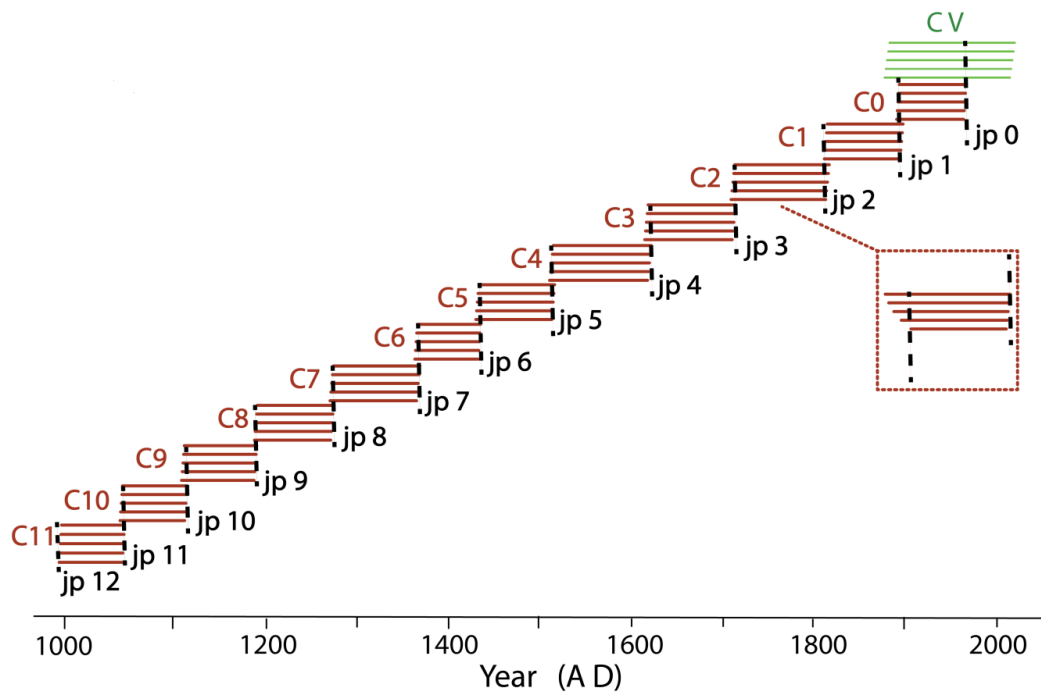
Figure 2. Illustration of the sampling strategy. From year 1000 to 2006, cohorts are represented in red for subfossil stem segments (C0-C11), and in green for living trees (CV). The join points are in black on the overlapping periods (JP0 to JP12).
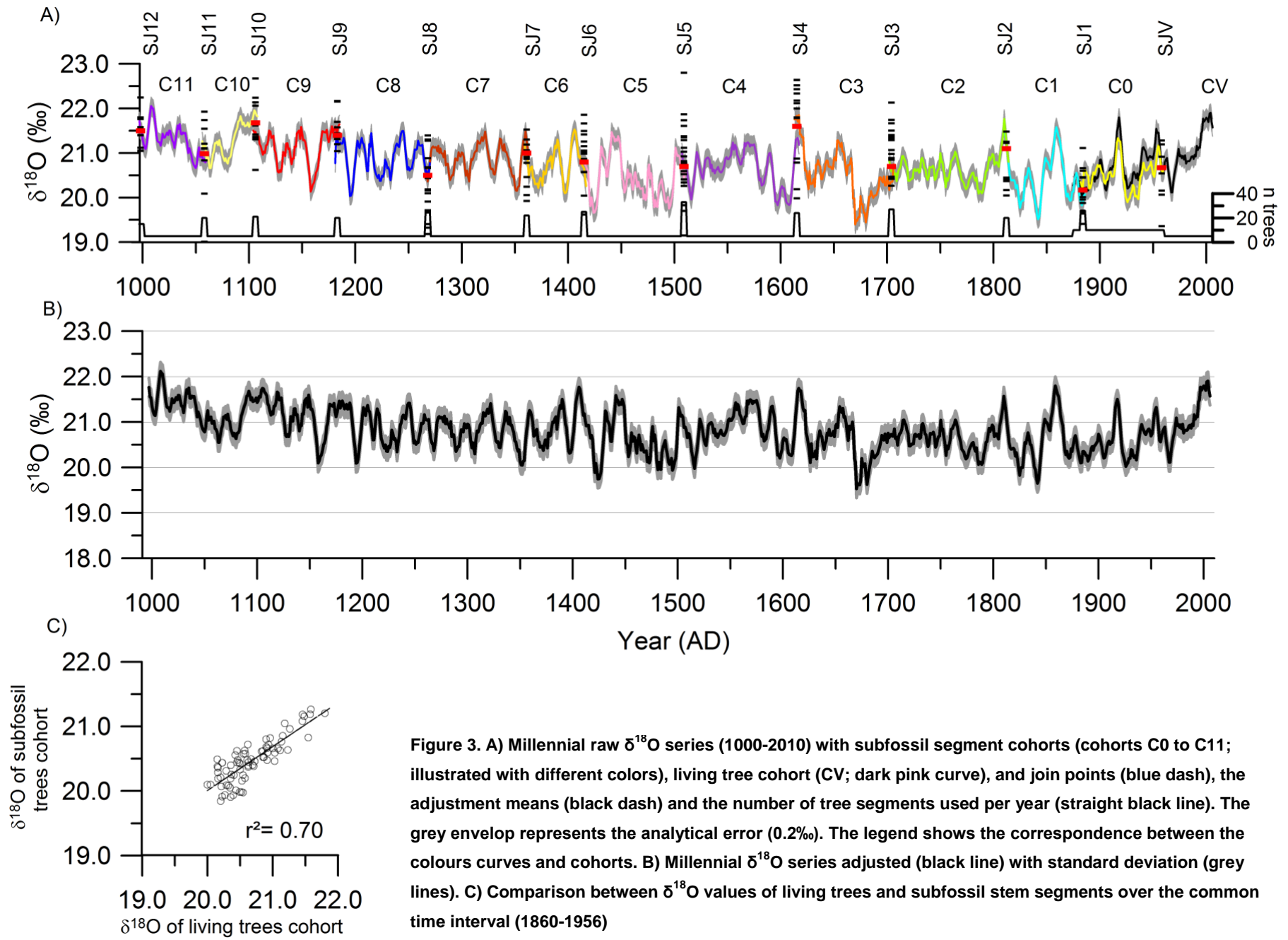
Figure 3. A) Millennial raw δ^{18}O series (1000-2010) with subfossil segment cohorts (cohorts C0 to C11; illustrated with different colors), living tree cohort (CV; dark pink curve), and join points (blue dash), the adjustment means (black dash) and the number of tree segments used per year (straight black line). The grey envelop represents the analytical error (0.2‰). The legend shows the correspondence between the colours curves and cohorts. B) Millennial δ^{18}O series adjusted (black line) with standard deviation (grey lines). C) Comparison between δ^{18}O values of living trees and subfossil stem segments over the common time interval (1860-1956)
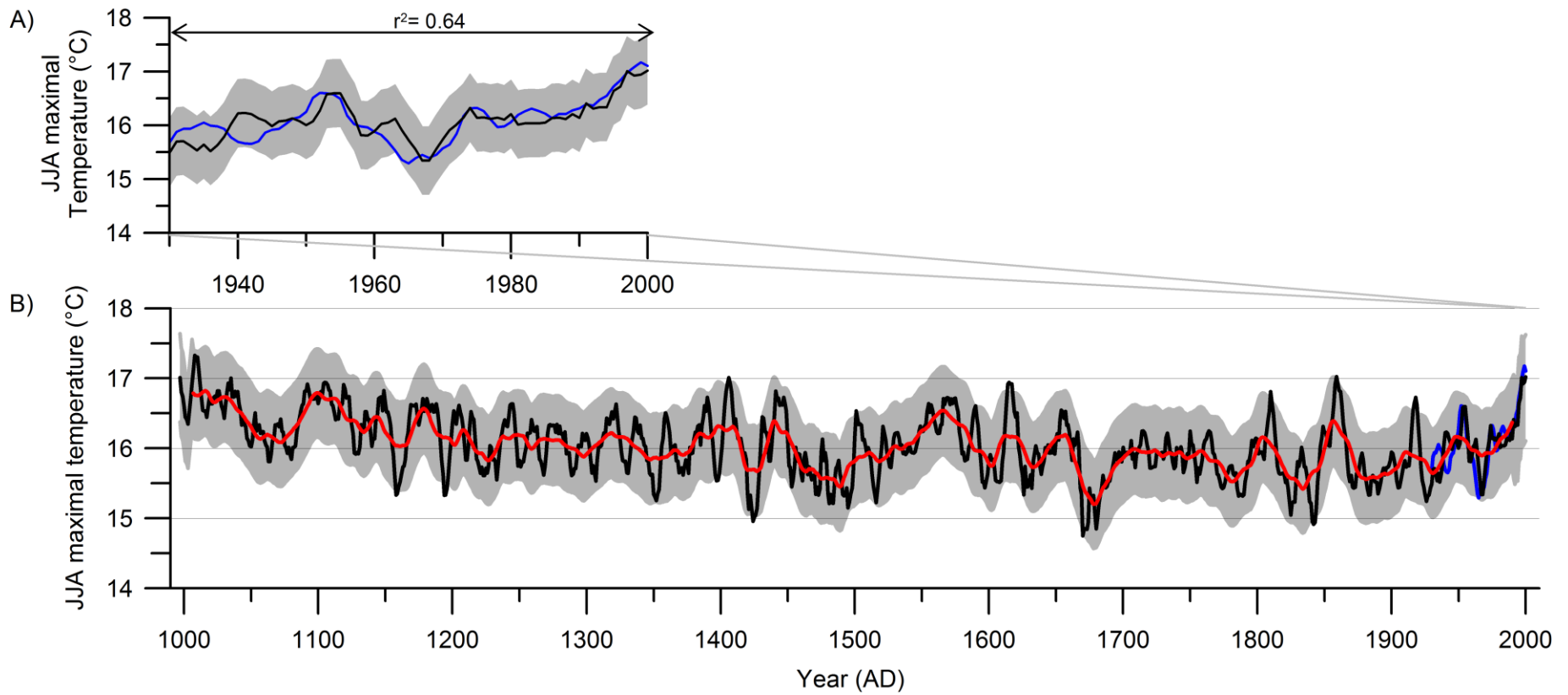
Figure 4. A) Comparison between reconstructed (i-STREC, black line) and observed (blue line) JJA maximal temperature and mean square values on the entire period of calibration. B) i-STREC (black line) with 21-year moving average (red line) and observed JJA maximal temperature series from CRU TS3.1 (blue line). In both cases, the dark grey shading represents uncertainty with ±1 RMSE calculated on the 21-year filter values.
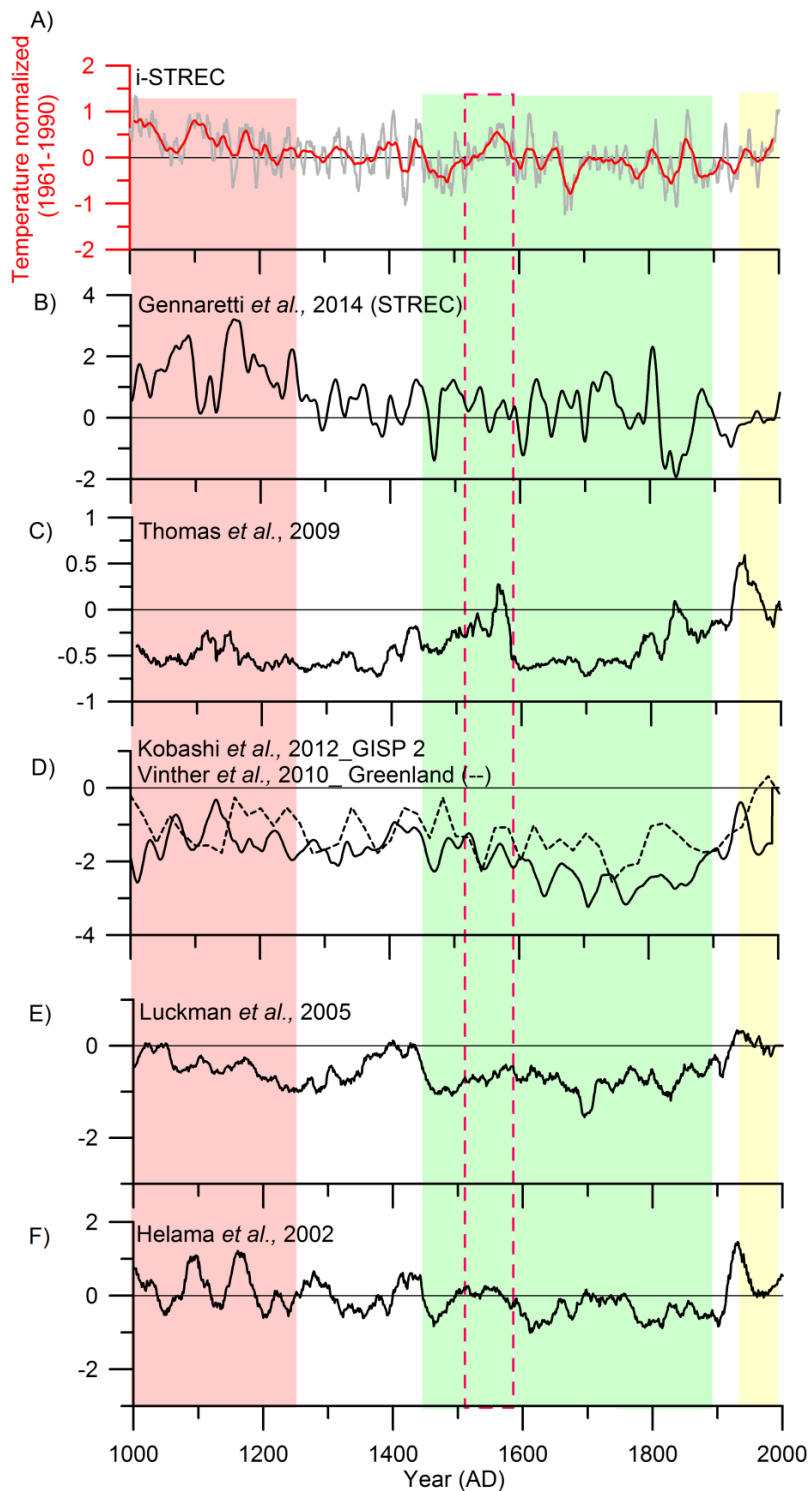
Figure 5. Comparison between i-STREC and other temperature reconstructions (data obtained from NOAA). (A) i-STREC from in north-eastern Canada. B) STREC from tree-ring width, in the same region. C) July-September temperature from varved sediments, Baffin Island, Arctic Canada. D) annual surface temperature from GISP 2 ice core, in Greenland. E) May-August maximum temperature from maximum latewood density and tree-ring width, Canadian Rockies. F) July temperature from tree-ring width, northern Finland. Shading based on i-STREC is shown to ease comparison with the other reconstructions compiled: warmer (pink, colder (green) and modern periods (yellow). All reconstructions have been smoothed with a 21-years filter and normalized (1960-1991).
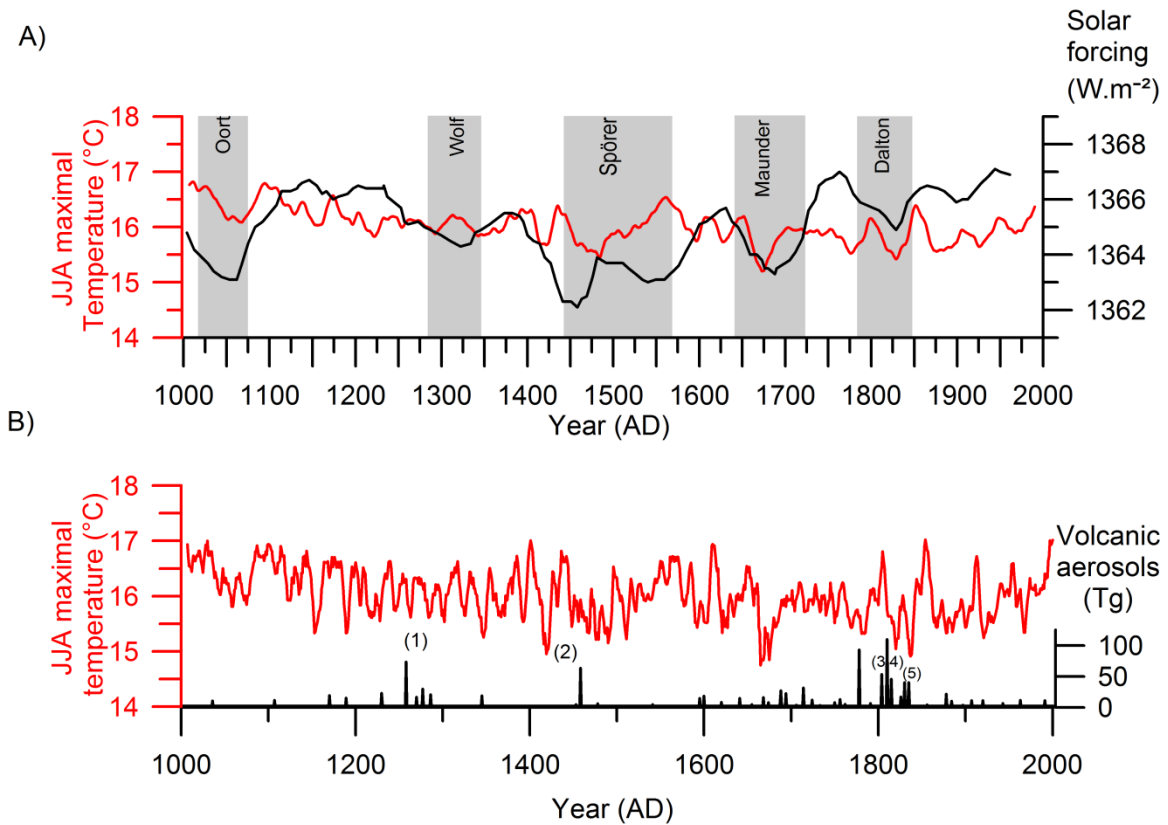
Figure 6. Volcanic and solar forcings. A) I-STREC (reconstructed summer temperature, 21-years smoothed; red line), compared with the well-known solar minima (grey bands) and the solar forcing series (black line; Bard *et al.,* 2003). B) I-STREC reconstructed summer temperature (red line) compared with the volcanic aerosols sulfates (Sigl *et al.,* 2013). The major eruptions are marked: (1) 1257/1258= Samalas, (2) 1456= Kuwae, (3) 1783= Laki, (4) 1809= unknown and 1815= Tambora and (5) 1835= Cosigüina).

1 **Table 1. Summary of the verification statistics for calibrations using the linear scaling and simple**
2 **linear regression methods for different periods, and using the measured maximal temperature**
3 **series.**

	Calibration (1930-1970) Linear scaling/ simple linear regression	Calibration (1971-2000) Linear scaling/ simple linear regression	Calibration (1930-2000) Linear scaling/ simple linear regression
Average °C± SD	15.9 ± 0.3	16.4 ± 0.4	16.1 ± 0.4
R² calibration	0.27/ 0.43	0.85/ 0.85	
R² verification	0.85/ 0.85	0.27/ 0.43	
R² total period			0.62/ 0.64
RMSE verification, °C	0.14/ 0.33	0.34/ 0.33	
RMSE calibration, °C	0.33/ 0.16	0.15/ 0.14	
RMSE total period, °C			0.27/ 0.25
RE	0.94/ 0.19	0.64/ 0.07	0.57/ 0.64
CE	0.84/ 0.05	-0.02/ 0.21	0.57/ 0.64

4 SD is the standard deviation, r² is the coefficient of determination (R squared), RMSE the raw
5 mean squared error, RE the reduction of error and CE the coefficient of error.