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A chronology of climatic downturns through the mid- and late-Holocene: tracing the distant effects of explosive eruptions from palaeoclimatic and historical evidence in northern Europe

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Abstract

Geochronological data of the conifer tree rings in a region sensitive to climatic effects of explosive eruptions were analysed for sudden growth reductions in association with extraordinarily cool reconstructed summer temperatures since 5500 B.C. Tree-ring data came from the stems of living trees and subfossil tree remains collected as increment cores and discs, respectively, from an area of northernmost Finnish Lapland (70–68°N to 30–20°E). Calendar year dates when the tree-ring signatures (i.e., growth reductions and reconstructed temperatures) were concurrent were compared with sulphate data from Greenland ice cores. Previous new evidence are in agreement in demonstrating volcanism behind late-Holocene events in 1601 A.D. and 536 A.D., suggesting that the same causal relationship can be implied further back in time. Our data show that earlier events were found to have occurred in the years 330 B.C., 874 B.C., 1464 B.C., 1584 B.C., 2564 B.C. and 2850 B.C. Interestingly, events of lesser magnitude followed the three major events in 542 A.D., 1453 B.C. and 1579 B.C. by a few years. Natural disasters, and grain crop failures, occurred as a result of these events, as has been documented for the summer of 1601 A.D. through Finnish historical data and broadly in the Northern Hemisphere. Climate has surprised humans during historic and likely pre-historic times, causing sudden alterations in agriculture, ecology and economy, and may do so in the future. We argue that the climate change with the most magnified impacts on society may be a negative temperature anomaly that abruptly decreases resource availability over wide spatial scales.

Climate change may surprise us with unexpected physical changes in the Earth's atmosphere (Streets & Glantz 2000). Surprising events that cause the most damage to human health and property and impact the natural environments are sudden, discontinuous changes leading to extreme weather and climate events (Easterling et al. 2000; Streets & Glantz 2000). Observations and models are in agreement in showing that strong climate anomalies can result from the atmospheric effects of large explosive volcanic eruptions (Chester 1988; Robock 2000; Miles et al. 2004). Indeed, a special characteristic of volcanism is that unusually violent eruptions are known to follow unusually long periods of quiet and that the eruption of volcanoes with no historic activity may lead to the worst disasters (Simkin 1993). A multidisciplinary evaluation is needed to complement the historical record of volcanism. Direct and secondary evidence of the eruptions can be traced from deep ice cores and long tree-ring series, where impurities by volcanic aerosols and growth

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anomalies, respectively, are known to follow such eruptions (Zielinski 2000).

A recent study elucidated the climatic response in Europe following 15 major tropical eruptions over the last half millennium (Fischer et al. 2007) and confirmed the clear pattern of summer temperature cooling during the first and second post-eruption years (Bradley 1988; Robock 2000). Of these, the strongest signal of cooling is found during the year after the eruption. Notably, Finland lies in the area of the most pronounced summer cooling (see figure 1 in Fischer et al. 2007). In the same region, tree rings and their summer temperature reconstructions have been suggested to exhibit volcanic signatures as distant effects of explosive eruptions (Helama, Lindholm et al. 2005; Salzer & Hughes 2007; Helama, Läänelaid et al. 2010). Similar indications have been detected in adjacent areas of northern Europe (Gervais & MacDonald 2001; Salzer & Hughes 2007; Shumilov et al. 2011). By virtue of its geographical location and the sensitivity of tree rings as a proxy for climate, tree-ring material from the region of northern Finland known as Lapland would appear privileged for tracing the climatic effects of past volcanism.

The ongoing task of exploring subfossil wood in Finnish Lapland for dendroclimatic purposes has so far resulted in a chronology spanning 7.6 thousand years (Eronen et al. 1999; Eronen et al. 2002; Helama et al. 2008). The tree-ring chronology of Finnish Lapland has been used to reconstruct the region's past summer temperature variability since 5500 B.C. on annual to millennial scales (Helama, Macias Fauria et al. 2010). This timeframe provides an interglacial window excluding the early Holocene, during which perturbed volcanism followed the deglaciation isostatic adjustment of magma chambers to the associated changes in crustal stresses, as indicated by chemical data yielded by deep ice cores (Zielinski et al. 1994; Zielinski et al. 1997). Compared to the Greenland ice-core chemistry (see Table 1), tree-ring dating allows for greater precision in determining when events occurred (Baillie 2008, 2010).

The signature years of the new palaeoclimatic reconstruction (Helama, Macias Fauria et al. 2010) have not so far been systematically examined. To analyse the data, we employ robust methodology based on multiple lines of evidence. First, our analyses are constrained by the knowledge of the effects in the region of explosive eruptions that take place far away (Fischer et al. 2007). Ensuring the robustness of our dendrochronological analyses, two different tree-ring methods were used to examine the initial dendroclimatic data producing the chronologies. In addition, we combined a predetermined

definition of the volcanic signature in the dendrochronological data (Gervais & MacDonald 2001) with a comparison to the chronologically constrained evidence of the Greenland ice-core chemistry (Table 1). For the late-Holocene, it was possible to superpose the palaeoclimate results upon the historical data and review the consequences in context of human ecology (Holopainen & Helama 2009). These analyses paved the way for our new record of climatic downturns, with linkages to the distant effects of volcanism on northern Europe through the present interglacial.

Material and methods

Initial tree-ring chronology

Data comprising the series of tree rings were derived from the stems of living trees and trunks of subfossil tree remains collected as increment cores and discs, respectively, from an area of northernmost Finnish Lapland located at 70–68°N to 30–20°E (Eronen et al. 1999; Eronen et al. 2002; Helama et al. 2008). The widths of the rings were measured under the microscope and the resulting tree-ring series cross-dated using established dendrochronological techniques (Aniol 1983; Holmes 1983; Van Deusen 1990). In this routine method, the series are examined for errors and the inter-correlative sample series are progressively added into the mean chronology (Fritts 1976). In addition to cross-dating within the tree-ring data of Finnish Lapland, the tree-ring chronologies of other parts of Finland (Helama, Lindholm et al. 2005) and the adjacent area in northern Sweden (Grudd et al. 2002) have been successfully constructed (Eronen et al. 2002). The Finnish Lapland chronology comprises 69 living and 1249 subfossil tree-ring series of Scots pine (*Pinus sylvestris* L.) and the combined chronology covers the interval from 5633 B.C. until the present-day (Helama, Macias Fauria et al. 2010).

Tree-ring sensitivity

The non-climatic trends in the tree-ring growth were removed prior to the targeted analyses using conservative growth trend modelling. A modified negative exponential curve (Fritts et al. 1969) or linear regression with negative or zero slope was fitted for each individual ring width series. The tree-ring indices were computed as ratios between the measured and modelled widths. This method of standardization (Fritts 1976) is known to be well suited to Scots pine tree-ring width data (Lindholm
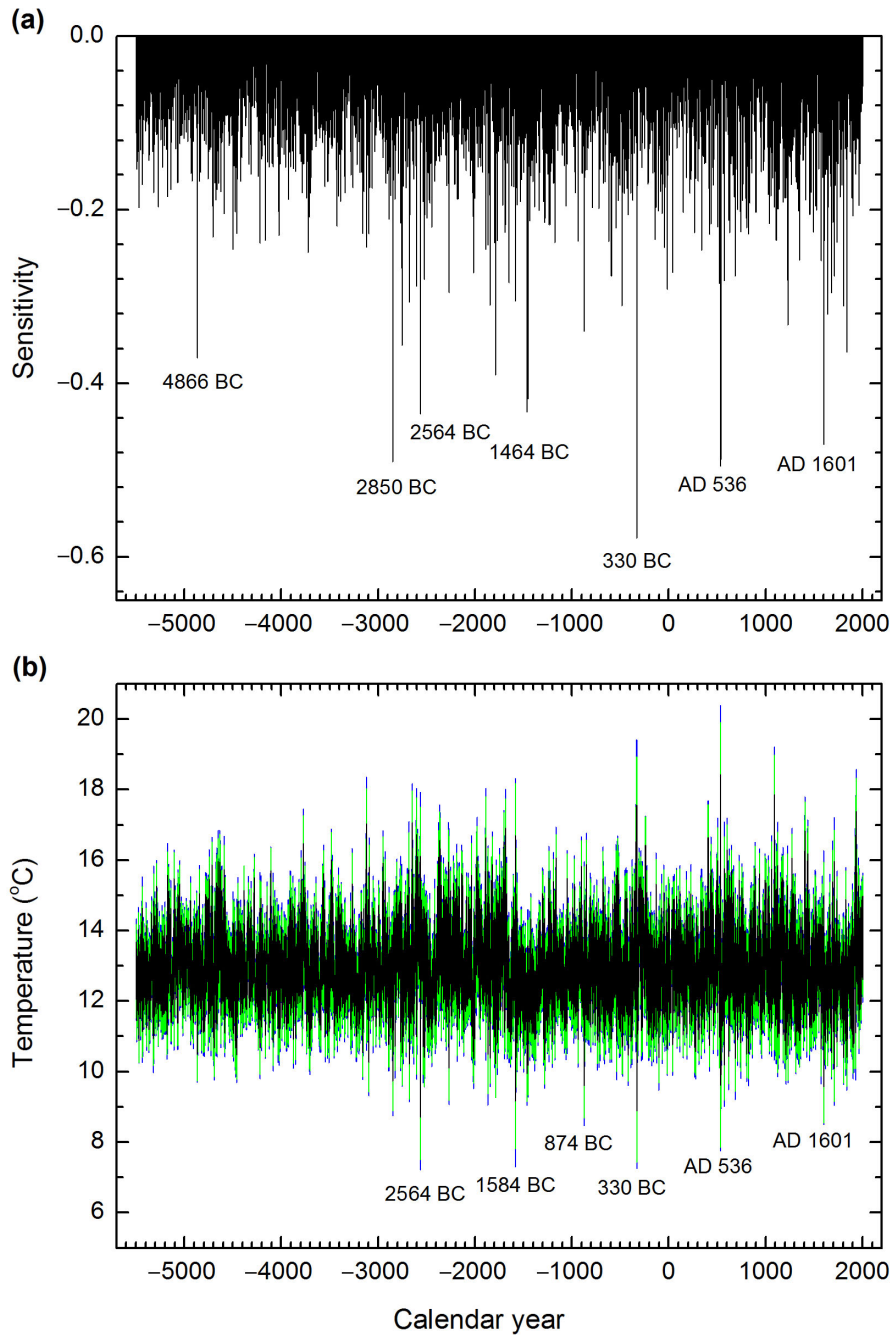
Fig. 1 A mid- and late-Holocene chronology of climatic downturns. (a) Tree-ring sensitivity (i.e., sudden change in growth conditions). Please note that only negative departures are given, the values therefore indicating growth reductions. (b) Reconstructed summer (July) temperature variability (black line) with the green and blue areas indicating the 95% and 99% confidence intervals of the reconstruction. The study period was 5500 B.C. through 2005 A.D. The years discussed in the text are shown as tree-ring dated calendar years B.C. and A.D.

1996; Helama, Lindholm et al. 2004). The mean chronology was calculated using a biweight robust estimation (Cook et al. 1990). The variance of the chronology was stabilized using the method of Osborn et al.

(1997), using a time-independent estimate of mean interseries correlation ($r = 0.344$; Helama, Lindholm et al. 2004). Following the previously applied protocol (Gervais & MacDonald 2001), the chronology was

Table 1 The Greenland ice cores referred to in this study.

Ice core	References
Dye 3 (part of the Greenland Ice Sheet Project)	Clausen et al. (1997), Vinther et al. (2006), Larsen et al. (2008)
Greenland Ice Core Project (GRIP)	Clausen et al. (1997), Southon (2002), Vinther et al. (2006), Larsen et al. (2008)
North Greenland Ice Core Project (NGRIP)	Rasmussen et al. (2006), Vinther et al. (2006), Larsen et al. (2008)
Greenland Ice Sheet Project 2 (GISP2)	Alley et al. (1993, 1997), Meese et al. (1997), Zielinski et al. (1994), Southon (2002)

characterized for the amplitude of the inter-annual growth variations, calculated using the sensitivity (Fritts 1976). The sensitivity was determined as $(x_t - x_{t-1})/(x_t + x_{t-1})$, where x is the mean index of year t (Gervais & MacDonald 2001). The tree-ring signature of large-scale climatic effects from explosive eruptions can lag the actual volcanic event (Jones et al. 1995; Briffa et al. 1998). That is, the excessively negative values in year t could be expected from the explosive eruptions from the same year t or the previous year t-n, that is, during the first and second post-eruption years (Bradley 1988; Fischer et al. 2007).

Indications of climatic cooling

A palaeoclimatic reconstruction (Helama, Macias Fauria et al. 2010) was produced using a regression of the regional summer temperature on the tree-ring width chronology. First, an empirically designed procedure resembling regional curve standardization (Briffa et al. 1996) was applied, with the exception that the concavity of the growth trend was adjusted for past changes in pine population density in the study region (Helama, Timonen et al. 2005; Helama, Macias Fauria et al. 2010). That is, the non-climatic variations were removed from the individual tree-ring series with a modified negative exponential curve (Fritts et al. 1969) tuned to account for palaeoecological information about the non-climatic changes in the trend shape (Helama, Macias Fauria et al. 2010). The tree-ring indices were derived as ratios between the observed and reference values, and the chronology was averaged using a biweight robust mean to produce annual index values. The variance of the chronology was stabilized using the method of Osborn et al. (1997), using a time-independent estimate of mean interseries correlation ($r = 0.330$) (Helama, Lindholm et al. 2004).

Second, the tree-ring chronology was regressed against instrumental climate data from a weather station in Karasjok, northern Norway (69°28′ N; 25°31′ E) to depict temperature variations on a Celsius scale (Helama, Macias

Fauria et al. 2010). This method was used here as it was previously evaluated as a robust palaeoclimatic model for similar data (Helama et al. 2009). The reconstruction accounts for more than 40% of the total observed variability ($R^2 = 0.43$; $R_{ADJ}^2 = 0.41$) over the full calibration period (1877–2004 A.D.). Moreover, our previous analysis showed that not only tree rings of living pines but also the tree-ring chronology of purely subfossil logs correlate positively with summer temperatures (Helama, Holopainen et al. 2004). Applying the obtained transfer function for the tree-ring data over the pre-calibration period, the chronology was transformed into yearly estimates of summer temperatures for 5500 B.C. through 2005 A.D. Confidence intervals for the final reconstruction were computed from the autoregressive structure of the residuals of the linear association between the reconstruction and instrumental data and the years 1877 A.D. and 2004 A.D., based on 1000 Monte Carlo simulations (Macias Fauria et al. 2010; Macias-Fauria et al. 2012). The extremely cool summers following explosive volcanic events with large-scale climatic effects were identified from the reconstruction as the most negative temperature departures from the long-term mean of the reconstruction.

Results

A look at the plot of the tree-ring variability reveals a characteristic paucity of anomalously poor growth in terms of dendrochronological sensitivity (Fig. 1a). The most negative years of growth are not clustered within a limited period but are spread over several millennia. Considering the late-Holocene (here, 1–2005 A.D.), the tree-ring record shows extreme drops in growth as having occurred in 1601 A.D. and 536 A.D. These years were reconstructed as having been exceptionally cool (Fig. 1b). For both of these years, the summer temperatures were reconstructed to have been cooler than 10°C on average, which is more than three standard deviations from the reconstructed mean of 13°C. The year 536 A.D. was followed by another year of reduced growth, in 542 A.D., during which the temperatures are reconstructed to have been nearly as cool as six years before (Fig. 2).

Tree-ring data for the B.C. era indicate that there were two years of reduced growth—in 330 B.C. and 2850 B.C.—that were comparable to the two late-Holocene events described above (Fig. 1a). The years 874 B.C., 1584 B.C. and 2564 B.C. were also reconstructed to be extremely cool years. However, the years 874 B.C. and 1584 B.C. did not show comparable sudden growth suppression, implying that the reconstructed coolness during these events was caused by a progressive cooling
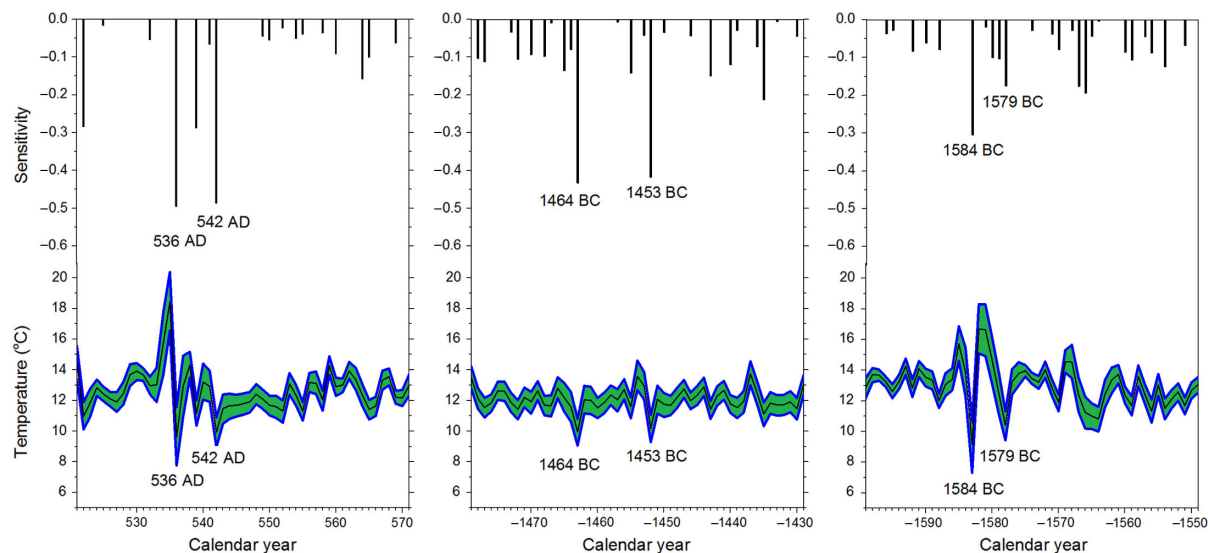
Fig. 2 Recurring events in growth and climate conditions. The duets occurred in 536/542 A.D., 1464/1453 B.C. and 1584/1579 B.C., shown here as sensitivities and reconstructed temperatures. Please note that only negative departures are given for sensitivity. The reconstructed summer (July) temperature variability (black line) is shown with the green and blue areas indicating the 95% and 99% confidence intervals of the reconstruction.

over two or more consecutive years. A quiescence of the earliest first two-and-a-half millennia of the record was contrasted with an exception of growth reduction in 4866 B.C. (Fig. 1a).

Of note, the events of 1464 B.C. and 1584 B.C. were followed by cool years 11 and five years later, respectively—in 1453 B.C. and 1579 B.C. These anomalies bring to mind the events in 536 A.D. and 542 A.D. (Fig. 2).

Discussion

Late-Holocene events

The signature years of the late-Holocene occurred in 1601 A.D. and 536 A.D. A closer look at these events is essential as they allow for the natural proxy evidence to be compared to historical documentation. Historical descriptions from Sweden, Norway and Iceland indicate that the sun was obscured in 1601 A.D. (Kalela-Brundin 1997). The Greenland Ice Core Project (GRIP; Clausen et al. 1997) and Greenland Ice Sheet Project 2 (GISP2; Zielinski et al. 1994) ice-core data suggest volcanic signatures in 1601 A.D. and 1604 A.D., respectively (Table 2). Later studies have assigned the coolness of the year 1601 A.D. to the eruption of Huaynaputina, in Peru, that had occurred the previous year (Briffa et al. 1998; de Silva & Zielinski 1998).

The geographical location of the study region poses challenges for agriculture because of the northern climate (Parry 1975; Parry & Carter 1985), and one would expect to see climatic effects on past agricultural production in

the region, as proposed by Solantie (1997). Documentary evidence of agricultural success (Tornberg 1989) reveals marked variability of rye and barley crops over the interval of time in which we are interested, and seems to indicate years of crop failure in southern Finland (ca. 60°N; 22°E) in 1554 A.D., 1577 A.D. and 1601 A.D. (Fig. 3). Historical harvests of rye and barley have been analysed on the basis of continuous records between 1551 A.D. and 1609 A.D. (Holopainen & Helama 2009). Recalculating their data, we find that the average grain-figure (ratio between sown and harvested grain) for the period of 1592–1600 A.D. was 5.5, whereas for the year of 1601 A.D., it was as low as 2.8 (Fig. 3). At much broader spatial scales, Atwell (2001) and Verosub & Lippman (2008) reviewed evidence from documentary and natural archives that show aggregations of anomalously cool weather and famine in several countries and continents.

Table 2 Likely correspondences between the tree-ring dates presented in this study and previously published evidence from the Greenland ice cores.

| Year	Ice-core evidence
1601 A.D.	1604 (Zielinski et al. 1994); 1601 (Clausen et al. 1997)
536 A.D.	533–543±2 (Larsen et al. 2008)
330 B.C.	365 (Zielinski et al. 1994)
874 B.C.	864 (Zielinski et al. 1994), 888 (Clausen et al. 1997)
1464 B.C.	1442, 1454, 1459 (Zielinski et al. 1994); 1457, 1463 (Clausen et al. 1997)
1584 B.C.	1577, 1594, 1600 (Zielinski et al. 1994)
2564 B.C.	No acidity peaks around this dendrochronological date
2850 B.C.	2815 (Zielinski et al. 1994)
4866 B.C.	4893 (Zielinski et al. 1994)
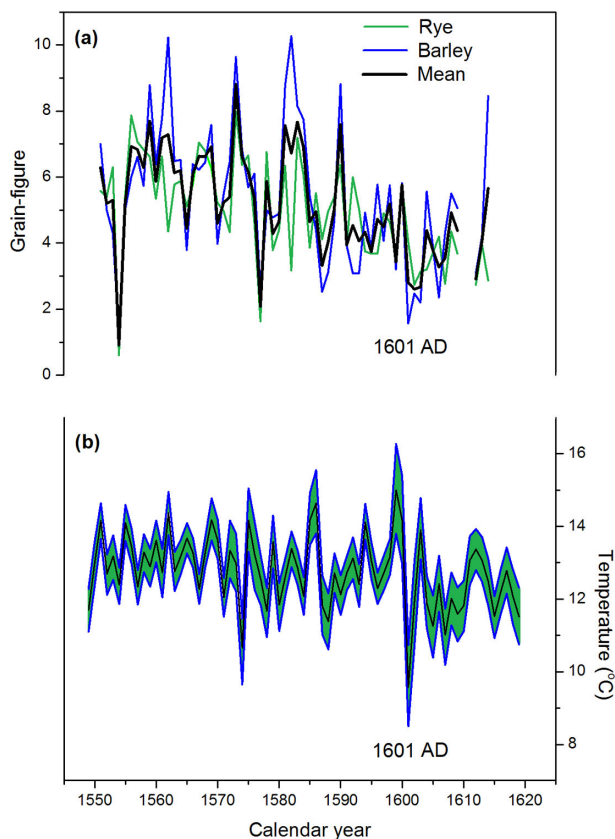
Fig. 3 Chronologies of harvests and climate variability. (a) Rye and barley grain-figures in south-western Finland for the period 1549 A.D. through 1619 A.D. shown with (b) the reconstructed summer temperature variability of this study (black line) with the green and blue areas indicating the 95% and 99% confidence intervals of the reconstruction. The grain-figure data were recalculated from the original study (Holopainen & Helama 2009) stabilizing the variance of the grain-figures using a method identical to tree-ring analyses (Osborn et al. 1997).

The 536 A.D. event has long intrigued scientists of various disciplines (Stothers 1984, 1999; Briffa et al. 1990; Zetterberg et al. 1994; Arjava 2005; Baillie 2008; Larsen et al. 2008; Woods 2010). The event with the associated evidence inferred from tree rings has been previously associated with the Rabaul volcano on the island of New Britain, Papua New Guinea (Stothers & Rampino 1983a, 1983b; Stothers 1984). New evidence obtained by radiocarbon dating of charcoal collected from the basal units of the local eruption deposits has suggested, however, that the Rabaul eruption occurred during the 7th century A.D. (McKee et al. 2011). The lack of volcanic deposits in ice cores (Clausen et al. 1997) has also led to the suggestion that a comet exploding in the upper atmosphere could result in the deposition of debris in the upper atmosphere, which would reduce the penetration of sunlight and cause a climatic

downturn (Rigby et al. 2004). However, the comparison of sulphate from multiple ice cores—Dye 3 (part of the Greenland Ice Sheet Project), GRIP and North Greenland Ice Core Project (NGRIP)—unveiled a signal consistent with the volcanic hypothesis for the event (Larsen et al. 2008). The temporal pattern of sulphate variations in the ice cores further anchored the ice-core and tree-ring indications of volcanic dust and climatic downturns in the years 522 A.D., 532 A.D., 536 A.D., 542 A.D. and 574 A.D. (Baillie 2008). The data shown here support our understanding of 536 A.D. and 542 A.D. as having been among the coolest summers in the Holocene (Fig. 2).

Comparisons over the B.C. era

Efforts to link tree-ring and ice-core data become more complex as one looks further back in time. Several studies examining individual and multiple ice cores have indicated increasing errors in layer counting with older ice materials (Clausen et al. 1997; Meese et al. 1997; Southon 2002; Rasmussen et al. 2006; Vinther et al. 2006). The estimations have shown a maximum dating error of 2% in the Holocene segments of the Greenland ice chronologies (Meese et al. 1997; Rasmussen et al. 2006). An additional 80-year offset was found close to 5300–5400 years ago when comparing the GRIP and GISP2 chronologies (Southon 2002). Moreover, our tree-ring results during the B.C. era did not always compare well with other tree-ring estimates (Salzer & Hughes 2007). With respect to different tree species and geographical locations, the deviations can originate from correspondingly dissimilar ecological circumstances. Differences can also result from discrepancies in methods (Baillie & Munro 1988; Salzer & Hughes 2007) to derive tree-ring information.

An indication of the 330 B.C. event in our tree-ring records could be linked with a GISP2 sulphate peak in 365 B.C. (Zielinski et al. 1994) (Table 2). Though the apparent difference of 35 years is large, errors in the ice chronology (Meese et al. 1997) may accommodate it. Broadly coinciding with these dates, the eruption of Atacazo occurred in the Western Cordillera of Ecuador (Siebert & Simkin 2002). Described from a geological perspective by Hidalgo et al. (2008), the eruptions are estimated to have had a large explosive magnitude. The calibrated radiocarbon dates average to 390–260 (1-sigma confidence level [CI]) and 400–23 calibrated years B.C. (2-sigma CI) (Hidalgo et al. 2008). In addition, our indication of the year 874 B.C. may match the sulphate signal for 864 B.C. shown in the GRIP2 ice core (Zielinski et al. 1994) and that for 888 B.C. in

the GRIP core (Clausen et al. 1997). The latter could point to a major volcanic eruption in the Northern Hemisphere as the deposited amount of volcanic acid was estimated to be three times more than the mean annual background acid deposition rate (Clausen et al. 1997).

The advantages of anchoring the tree-ring and ice-core evidence of volcanism by multiple recurring events has previously been demonstrated (Baillie 2008, 2010). In this regard, there is tree-ring evidence for a two-stage event involving the years 1464 B.C. (stronger) and 1453 B.C., with an 11-year interval. A similar pair of events was observed in tree rings in 1584 B.C. (stronger) and 1579 B.C. The 1584/1579 B.C. events would show an interesting correspondence with the GISP2 sulphate anomalies dated to 1600 B.C. and 1594 B.C., with a difference of six years (Zielinski et al. 1994). Yet a single sulphate anomaly nearest to the tree-ring dates would suggest an ice-core date of 1577 B.C. (Zielinski et al. 1994). Regarding the 1464/1453 B.C. events, the tree-ring data corresponded with the GISP2 sulphate peaks dated to 1454 B.C. and 1442 B.C., with a 12-year interval (Zielinski et al. 1994). Moreover, the Dye 3 ice core indicated chemical signatures for the years of 1463 B.C. and 1457 B.C. (Clausen et al. 1997). Thermoluminescence dating of volcanic samples from the island of Yiali in Greece (distinguishable from tephras from the more extensively studied Mediterranean eruption of the Santorini [Thera] volcano) derived a mean age of 1460 B.C. (Liritzis et al. 1996).

Dendrochronological dates of 1628 B.C. from European oak tree-ring widths (Baillie & Munro 1988), 1627 B.C. as a frost-damaged ring (LaMarche & Hirschboeck 1984), a growth minimum in 1626 B.C. (Salzer & Hughes 2007) from the North American bristlecone pine (*Pinus longaeva* and *Pinus aristata*) data, and the sulphate event in 1623 B.C. from GISP2 ice-core data (Zielinski et al. 1994) have been suggested as traces of the Santorini eruption. Our data did not indicate anomalous conditions during the period of 1627–1623 B.C. Therefore, the paired events in 1584 and 1579 B.C. found in this study represent yet a new line of evidence for the highly unstable climate in the mid-second millennium B.C.

The new evidence for the events of potentially volcanic origin in 2564 B.C. did not have a potential match in the ice-core results (Table 2). The tree-ring date of 2850 B.C. appeared within the uncertainty boundaries of the ice-core event in 2815 B.C. (Zielinski et al. 1994). The earliest tree-ring indication of volcanism in 4866 B.C. could match the GISP2 evidence of high sulphate values in 4893 B.C. (Zielinski et al. 1994).

Geochronology with ecological implications

Tree-ring data sensitive to summer temperature fluctuations were examined for anomalously negative proxy departures in a region where distant effects of explosive eruptions have been shown to produce a pattern of cooling for the first and second years after the volcanic events (Fischer et al. 2007). This was shown for the year 1601 A.D., with strong impacts on the agricultural production as reconstructed from historical documents from southern Finland (Fig. 3). The interannual variability in crop yields in this region is among the lowest in Finland (Mukula & Rantanen 1989a, b), which suggests that larger year-to-year variations may have been experienced elsewhere in Finland. The association is plausible as both the harvests of modern (Mukula & Rantanen 1987, 1989a,b) and historical (Holopainen & Helama 2009) agriculture are known to depend on the growing season climate.

Increased human mortality in the region in association with the crop failures has been documented for the first years of the 17th century A.D. (Jutikkala 1958, 2003a, 2003b). We acknowledge that linking climate phenomena with past eruptions involves suppositions (Sadler & Grattan 1999), but we nevertheless suggest this connection for the event in 1601 A.D. in northern Finland. The climate–crop association may have played a role in the agriculture-dependent economic changes that occurred around that time (Holopainen et al. 2012) and affected the reproductive success among the poor families living in the region (Rickard et al. 2010).

The temporal association of tree-ring anomalies with late-Holocene eruptions is suggestive of a relationship that can be extended back in time. In this regard, pollen data indicate sporadic cultivation activity in parts of Finland since the Bronze Age and thus during at least the past three millennia (Taavitsainen et al. 1998). The summer temperature driven environmental productivity has also been shown to influence hunter–gatherer population size fluctuations in the region (Tallavaara et al. 2010; Tallavaara & Seppä 2011). A clear picture of this type was derived via proxy comparison between the summed probability distribution of archaeological radiocarbon dates (dates-as-data) and varve thicknesses of organic sediment layers from a lake within the study region (Tallavaara & Seppä 2011).

Conclusions

This paper has presented new evidence for abrupt reductions in climatic cooling during the mid- and late-Holocene. The backbone of this study was the 7.6

thousand year-long tree-ring chronology from Finnish Lapland. Previous tree-ring and ice-core studies, as well as historical documentary evidence, were also drawn upon. The most severe of the late-Holocene events (1601 A.D. and 536 A.D.) clearly illustrated the distant effects of explosive eruptions in the study region of northern Europe. In particular, the event in 1601 A.D. exemplified reduced survival potential of historic humans when subjected to a volcanic-induced climatic surprise. Lack of coherence between events inferred from tree-ring data and ice-core sulphate data became evident over the B.C. era. It may be fair to state that these inconsistencies result from increasing ice-layer counting errors and problems related to annual layer identification in the ice cores. On the other hand, the tree-ring data yield more accurate dates for cooling events but do not prove whether a volcanic eruption was the cause. Nonetheless, the majority of the tree-ring dates over the B.C. era can be linked with sulphate peaks in multiple Greenland ice cores. Overall, the comparisons produced added evidence for linking the tree-ring and ice-core signals of volcanism over the present interglacial. This paper also supports the idea of analysing and interpreting the human ecology data in the context of geochronological evidence. Such analyses will help us to outline the effects that our society will encounter with the next sudden event of volcanic-induced large-scale cooling.

Large and explosive volcanic eruptions have occurred several times in the past and it is highly likely the geological activity of our planet will produce similar events in the future, though the timing of eruptions and the intensity of their climatic and environmental effects may again surprise people (Streets & Glantz 2000). Examining violent climate changes in the Holocene shows that they have low-probability of occurring in the short-term but that the recurrence of similar new events is unavoidable in the long-term. Society is now supposed to be preparing for gradual warming as the current climatic trends are predicted to continue over the next century (Bergström et al. 2011), which could reduce our ability to cope when a sudden event of volcanic-induced cooling takes place. In the perspective of human ecology, disasters occur when the buffering capacity of society is exceeded by natural events (Kates 1971; McLaughlin & Dietz 2008). Clearly, the climate phenomena described in this paper fulfil the criteria of anthropologically-defined disasters, involving a combination of natural and techno-logical features and a population in a condition of vulner-ability (Oliver-Smith 1996). Our research has indicated the adverse effects of abrupt negative temperature anoma-lies on food production, which has obvious ecological and economic implications (Holopainen & Helama 2009;

Rickard et al. 2010; Holopainen et al. 2012). Our findings may be seen parallel with archaeological (Gamble et al. 2005) and palaeontological (Foley 1994) research which shows that climate changes strongly disturb human (and hominin) populations, which respond by contracting.

Acknowledgements

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