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Key Points:
- Ice core analyses demonstrate that boreal fire emissions peak ~2.5 ka
- Global climate drivers fail in explaining the fire maximum
- Humans may have left a quantifiable signature on the climate system ~3 ka

Supporting Information:
- Texts S1–S3 and Figures S1–S15

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Europe on fire three thousand years ago: Arson or climate?

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Abstract The timing of initiation of human impacts on the global climate system is actively debated. Anthropogenic effects on the global climate system are evident since the Industrial Revolution, but humans may have altered biomass burning, and hence the climate system, for millennia. We use the specific biomarker levoglucosan to produce the first high-temporal resolution hemispheric reconstruction of Holocene fire emissions inferred from ice core analyses. Levoglucosan recorded in the Greenland North Greenland Eemian ice core significantly increases since the last glacial, resulting in a maximum around ~2.5 ka and then decreasing until the present. Here we demonstrate that global climate drivers fail to explain late Holocene biomass burning variations and that the levoglucosan maximum centered on ~2.5 ka may be due to anthropogenic land clearance.

1. Introduction

Fire is a key Earth system process and is a major component of carbon cycle [*Bowman et al.*, 2009; *Keywood et al.*, 2013]. Fire affects the climate system principally by releasing particulates, greenhouse gases and carbon [*Bowman et al.*, 2009; *Intergovernmental Panel on Climate Change*, 2013], which would otherwise be stored in woody vegetation in a world without fire [*Bond et al.*, 2005].

Fire and climate influence each other, where climatic conditions are the fundamental driver for the ignition and spread of fire. Sedimentary charcoal records show strong linkages between fire activity and climate [*Marlon et al.*, 2013], generally with low fire activity during cold periods and increased biomass burning in warmer periods [*Daniau et al.*, 2012; *Power et al.*, 2008]. Climate influences fire activity mainly by affecting biomass availability and flammability. Temperatures and atmospheric CO_2 may control plant productivity and thus fuel availability. Precipitation affects fuel flammability, where conditions must be wet enough to allow biomass to grow, yet dry enough to allow combustion [*Daniau et al.*, 2010b; *Marlon et al.*, 2013, 2008; *Pyne*, 2001]. Wet periods can favor biomass growth, while droughts and warming can increase fire ignitions and spread [*Westerling et al.*, 2006].

Human activities may have also influenced fire activity, and thus the climate system, for millennia [*Marlon et al.*, 2008; *Power et al.*, 2008]. Humans and fire have always coexisted, initially for domestic and hunting purposes, wildlife, and vegetation management. A routine use of fire for forest clearance and slash-and-burn agricultural practices only occurred since the Holocene, but when it began on broad scales is unknown and a matter of debate. A fundamental paleoclimate question, then, is when humans began to significantly alter fire regimes and, in turn, the climate system [*Bowman et al.*, 2009]. During the past millennia humans have converted forests and tree-covered lands to cropland and pasture through slash-and-burn techniques and may have unconsciously altered the climate by releasing teragrams of greenhouse gases through clearing forests to create open space.

Most paleoecological studies suggest intensive European forest clearance through fire for agriculture began approximately 3000 years before present [*Marlon et al.*, 2013; *Molinari et al.*, 2013]. Independent evidence, such as charcoal records from sediment cores, indicate a maximum in biomass burning in North America and Europe during the same time period [*Marlon et al*., 2013] (supporting information). While local climate changes may explain the subsequent decrease in fire activity in North America, human activities have played a greater role in Eurasia. When local charcoal records are synthesized at regional to global perspectives, a worldwide increase in fire activity between 2 and 3 ka rise up everywhere except in Australia [*Marlon et al*., 2013], even if climate is relatively stable throughout the Holocene. There is synchronous evidence of agricultural expansion and cultivated areas in Europe, North America, and Asia, therefore suggesting a link between humans and fire.

Models incorporating anthropogenic activities [*Ellis et al*., 2013; *Kaplan et al*., 2009, 2011] support maximum land clearing between ~3 and 2 ka in Europe, where decreased fire activity after 2.5 ka may be due to reduced fuel loads caused by anthropogenic activities such as fragmenting forests with fields and infrastructure. Model simulations based on new data about how land use per capita declines with advances in technology suggest that past attempts to quantify anthropogenic perturbation of the Holocene carbon cycle may have greatly underestimated early human impact on the climate system [*Kaplan et al*., 2009].

Charcoal records are primarily local, limited in number, and not homogeneously distributed and sampled. Differences in geographical abundances of records have to be carefully weighted in regional to global charcoal syntheses. Ice cores provide the possibility of obtaining a hemispheric record of fire activity that includes areas that are not covered by the Global Charcoal Database. Combining ice core and charcoal records provide a robust record of past biomass burning.

Here we introduce and investigate a high-resolution Holocene record of changes in Northern Hemisphere fire emissions in the North Greenland Eemian (NEEM) ice core using measurements of the biomarker levoglucosan. This broad-scale fire reconstruction from the glacial transition until the present allows examining possible anthropogenic influences on variations in past fire activity. We examine potential fire activity controls through comparisons of biomass burning emissions recorded in the Greenland NEEM ice core with independent reconstructions of fire history from charcoal data, with modeled area burned with and without anthropogenic interference, and with climate indicators.

2. Methods

The deep NEEM ice core (Figure S1 in the supporting information) was drilled from 2008 to 2012 in Greenland (77.49°N; 51.2°W, 2480 m above sea level (asl)), where the ice core was processed in the field using a Continuous Flow Analysis (CFA) system [*Zennaro et al*., 2014]. Samples for levoglucosan determination are from the inner core section, where a single CFA sample contains a swath of 1.10 m of core. Frozen samples were transported to Ca' Foscari University of Venice and stored in a −20°C cold room before being analyzed. Deep glacial ice is subjected to considerable stress, where the pressure inside the highly compressed bubbles may cause cracking of the ice, due to the strong pressure change after drilling. The presence of fractures promotes contamination by the drilling fluid and limits the use of the CFA system. We integrated CFA samples with 154 additional "brittle ice samples" (from 566.50 to 1279.85 m depth, 2820–8903 years B.P.), where each sample is approximately a 5–10 cm piece of ice core. We analyzed a total of 629 samples from 4.95 to 1493.25 m depth (1–14,962 years B.P.). We use the decontamination protocol and the analytical procedure presented in *Zennaro et al*. [2014], which slightly modifies the method for determining levoglucosan using liquid chromatography/negative ion electrospray ionization-tandem mass spectrometry (HPLC/(−)ESI-MS/MS) reported by *Gambaro et al*. [2008]. Blanks are consistently less than the lowest levoglucosan concentrations (blanks < 10 pg mL^{-1}). NEEM samples were analyzed in a random order; thus, potential contamination or fluctuations in instrumental precision and accuracy are not expected to influence the centennial to millennial trends of levoglucosan flux.

Brittle ice samples were transported in a frozen state to the University of Venice laboratory, where they were stored in a −20°C cold room until decontamination. We decontaminated brittle ice samples rinsing each sample 3 times with ultra pure water using a 1 L wash bottle and TFE clamps under a Class-100 clean bench located in a Class-10,000 clean room at the University of Venice until reaching a net loss of 1 cm of ice in each dimension. The cleaned ice was then melted into 125 mL low-density polyethylene (LDPE) wide-mouth bottles under the clean bench and stored into 15 mL LDPE bottles.

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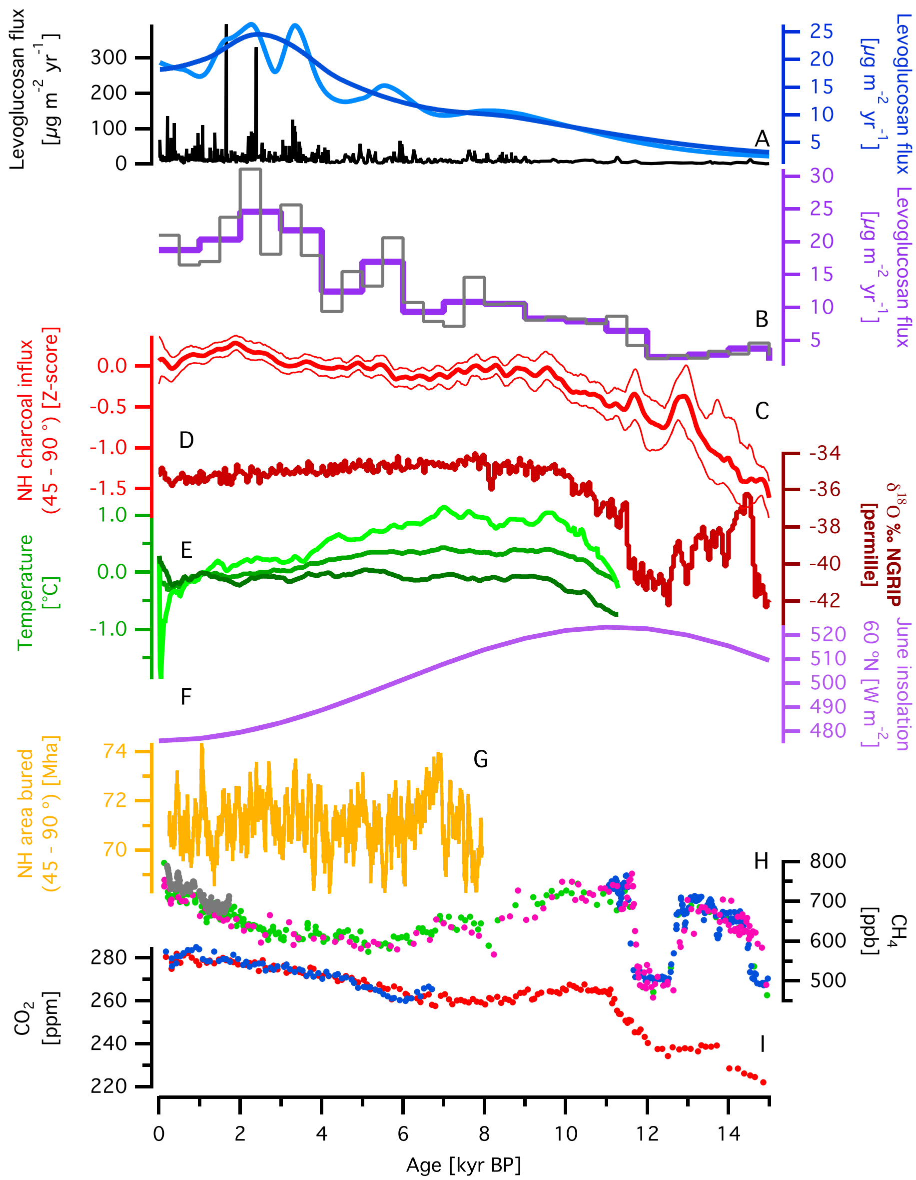

Figure 1. Levoglucosan flux (black) during the past 15 kyr over the NEEM ice core and LOWESS smoothing with SPAN parameter (f) 0.2 (light blue) and 0.5 (blue) of levoglucosan flux (A); Levoglucosan flux averages for 1000 year (violet) and 500 year (gray) windows during the past 15 kyr over the NEEM ice core (B); *Z* scores of high-latitude (45–90°N) Northern Hemisphere charcoal influx (500 year LOWESS smoothing) with 95% bootstrap confidence interval (red) (C). Fifty year mean NGRIP $d^{18}O$ (bordeau), as reported by *NGRIP Project Members* [2004] (D); Northern Hemisphere (30–90°N) (light green), global (green), and North Atlantic (dark green) temperature over the past 11.3 kyr, as reported in *Marcott et al*. [2013] (E); June insolation at 60°N [*Berger and Loutre*, 1991] (F); Modeled burned area for high-latitude (45–90°N) Northern Hemisphere (G); Preindustrial CH_4 analyzed in the Arctic GISP2 (black) [*Kobashi et al*., 2007], GRIP [*Blunier et al*., 1995] (green), NGRIP (blue) [*Baumgartner et al*., 2012], GISP2 (purple) [*Brook et al*., 1996], and NEEM (gray) [*Rhodes et al*., 2013] (H); Preindustrial atmospheric CO_2 concentrations analyzed in the Antarctic Taylor Dome (blue) [*Indermuhle et al*., 1999] and EDC (red) [*Monnin et al*., 2004] ice cores (I).

Accumulation in Greenland was not constant during the past 15 kyr [*Rasmussen et al*., 2013]. These differences in local accumulation change the wet deposition of aerosol in Greenland [*Fischer et al*., 2007], and we therefore use levoglucosan fluxes rather than concentrations. We corrected concentrations for Greenland snow accumulation data by multiplying each 1.10 m levoglucosan concentration by the accumulation rate for its respective depth interval after applying the NEEM ice core chronology

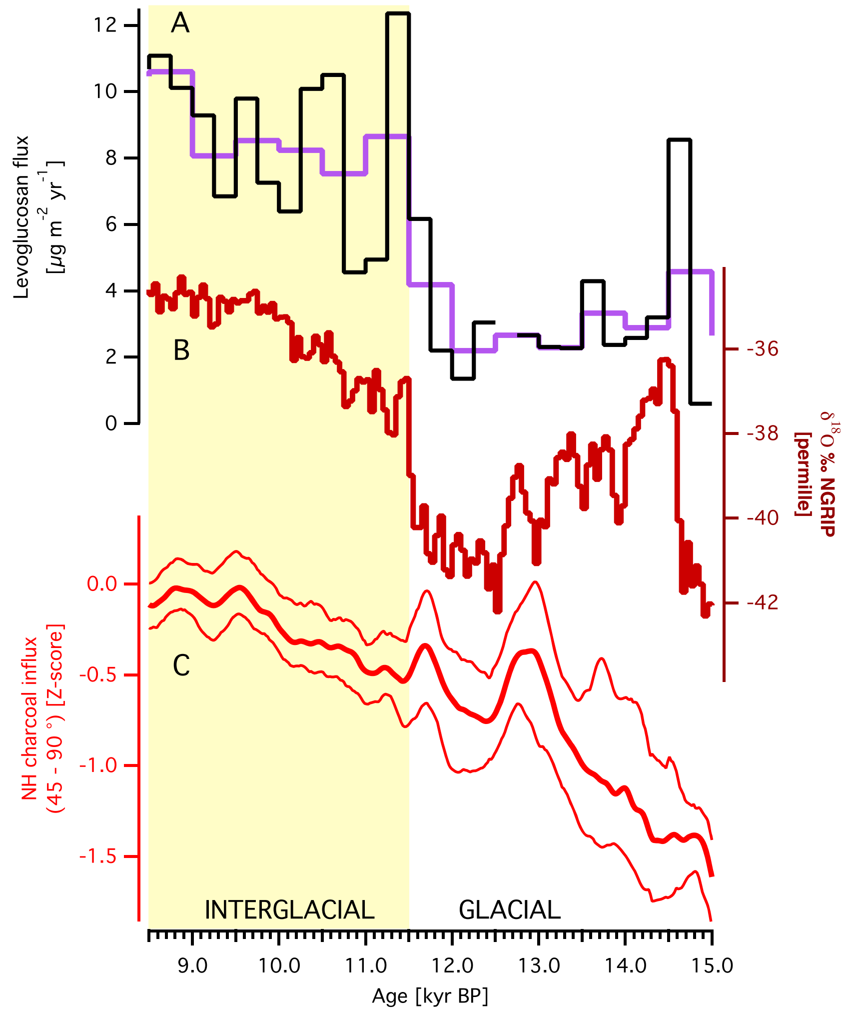

Figure 2. Levoglucosan flux averages for 500 year (violet) and 250 year (black) windows during the glacial-interglacial transition (A); 50 year mean NGRIP d^{18}O (bordeau), as reported by *NGRIP Project Members* [2004] (B); *Z* scores of high-latitude (45–90°N) Northern Hemisphere charcoal influx (500 year LOWESS smoothing) with 95% bootstrap confidence interval (red) (C).

[*Rasmussen et al*., 2013]. Long-term fire activity was reconstructed from levoglucosan fluxes (Figure 1A) by smoothing the data using a locally weighted scatterplot smoothing (LOWESS) smoothing tricube function (Figure 1B), as discussed in *Zennaro et al*. [2014].

3. The Levoglucosan Record

The NEEM levoglucosan record demonstrates an increase in fire activity throughout much of the Holocene, peaking at ~2.5 ka (Figure 1A) even when using multiple statistical thresholds to remove outlying values (supporting information). LOWESS smoothing (Figure 1B), 250 year and 500 year means (Figure 2) of the levoglucosan flux indicate that the lowest levoglucosan values in our record occurred ~15 ka when the flux was ~3 $\mu g\,m^{-2}\,yr^{-1}$. These values generally remained lower than 5 $\mu g\,m^{-2}\,yr^{-1}$ until ~12 ka, before increasing approximately threefold during the glacial-interglacial transition. The levoglucosan flux increased from previous levels and remained substantially above or close to 8 $\mu g\,m^{-2}\,yr^{-1}$ during the past 10 kyr. A peak is recorded ~5.5 ka, when the average flux was 16.9 $\mu g\,m^{-2}\,yr^{-1}$ between 5 and 6 ka. The levoglucosan flux abruptly increased ~4 ka when the maximum is recorded between 3.5 and 1.5 ka, with the average flux of 24.6 $\mu g\,m^{-2}\,yr^{-1}$ between 2 and 3 ka (Figure 1B).

The degradation kinetics or microbial decomposition of levoglucosan trapped in ice cores is currently unknown. If microbial degradation occurred, the levoglucosan loss would be proportional to the elapsed time, resulting in consistently declining levoglucosan concentrations in the older ice core sections. However, the peak levoglucosan concentrations are centered on ~2.5 ka (Figure 1). Thus, degradation processes cannot explain the levoglucosan fire maximum ~2.5 ka, and the Holocene levoglucosan concentrations are not a simple degradation curve.

Zennaro et al. [2014] suggest that North American fires are the main sources of pyrogenic aerosols transported to NEEM during the past two millennia due to their proximity and the modern boreal forest extent. Asian forest fire contributions are also likely important, especially during decadal-scale droughts. Temperature is the main control on forest fire activity on centennial timescales, but on multidecadal or shorter timescales precipitation may be the primary influence on boreal fires, as suggested by increased levoglucosan peaks during dry periods [*Zennaro et al*., 2014].

A positive shift in methane concentrations is observed in the Greenland Ice Sheet Project 2 (GISP2) (Greenland) ice core between ~ 3.2 and 2.8 ka [*Brook et al*., 1996], but not in other ice cores (Figure 1) [*Blunier et al*., 1995], together with an isotopic excursion of $\delta^{13}CH_4$ toward less negative values around 3 ka [*Sowers*, 2010]. This evidence may suggest pyrogenic inputs to the hemispheric methane budget. However, the lack of $\delta^{13}CH_4$ shifts during other levoglucosan oscillations (i.e., 5.5–6.0 ka) suggests regional (i.e., Europe) rather than hemispheric-scale biomass burning changes recorded in levoglucosan-based fire reconstructions.

The Global Charcoal Database is a robust syntheses of widespread observations, where the ~3–2 kyr B.P. fire maximum is analyzed in terms of human influence in a recent review of fire controls at regional to global scales [*Marlon et al*., 2013]. Multiproxy reconstructions in continental sediment cores (i.e., pollen and charcoal; synthetized in the supporting information) suggest strong human influence and disturbances on fire activity beginning ~4 ka [i.e., *Molinari et al*., 2013]. Model results independently simulate extensive European deforestation ~3 ka [*Kaplan et al*., 2009] or do not match with observed fire reconstruction (in Europe) due to the absence of anthropogenic forcing in the model [*Brücher et al*., 2014]. While the levoglucosan data and charcoal data peak at the same time, they have low-correlation values (supporting information). Levoglucosan and charcoal are different types of fire activity records in the sense that ice cores are single “receptors” that integrate atmospheric inputs over wide geographic areas, while individual charcoal sites provide primarily local data that are then synthesized into composite records that represent some areas better than others, depending on where sites exist.

Our Holocene levoglucosan results agree with regional and high-latitude (45–90°N) Northern Hemisphere charcoal syntheses demonstrating a preindustrial regional and global maximum in biomass burning between ~3 and 2 ka, followed by a decline during the following millennia (Figure 1C and Figure S7 and Text S2 in the supporting information) [*Marlon et al*., 2013]. We compare the NEEM levoglucosan profile with the available regional charcoal syntheses (supporting information) from potential boreal source areas [*Zennaro et al*., 2014] and farther south in nonboreal areas due to potentially increased forest extent thousands of years ago. The regional charcoal syntheses demonstrate that the 3–2 ka fire peak is evident in both North America and Europe, where the fire peak is generally ascribed to climatic and anthropogenic influences, respectively (supporting information). As an example of this, data-model analysis of 156 sedimentary charcoal records across Europe and paleoclimate data was used to explore spatial and temporal patterns, and potential drivers of biomass burning, also suggesting a progressive increase in fire frequency ~3.5 ka. Vegetation, precipitation, and temperature have a primary role during the early Holocene, while the mid-Holocene increase in fire activity is best explained by anthropogenic land cover change [*Molinari et al*., 2013].

3.1. Influences During the Late Glacial and Holocene

Fire regimes, in terms of frequency, size, and intensity of fire, are the result of changes in climate, vegetation, and anthropogenic burning, as well as combinations of these factors [*Daniau et al*., 2012; *Krawchuk and Moritz*, 2011; *Krawchuk et al*., 2009; *Marlon et al*., 2008; *Whitlock et al*., 2010]. Temperature, insolation, and atmospheric CO_2 levels influence vegetation productivity and thus fuel availability with warmer conditions generally correspond with increased fire activity [*Daniau et al*., 2010a, 2012; *Marlon et al*., 2013, 2008; *Whitlock et al*., 2010]. Precipitation primarily affects fuel flammability, where dry conditions increase flammability, but the conditions must also be sufficiently humid to allow vegetation growth [*Daniau et al*., 2012; *Krawchuk et al*., 2009; *Whitlock et al*., 2010].

Variations in the transport of biomass burning emissions to Greenland caused by climate changes since the Last Glacial (circulation patterns, transport efficiency, and forest locations) may also have influenced levoglucosan variations during the late Glacial and Holocene. Retreating continental ice sheets during the late Glacial, for example, were associated with an overall expansion and poleward shift in the extent and location of boreal forests, thereby increasing biomass burning sources and emissions. Greenland levoglucosan fluxes appear sensitive to this changing boreal ice extent [*Clark et al*., 2009; *Peteet et al*., 2012] insofar as levoglucosan quantities gradually increased from the onset of deglaciation ~15 ka until the present, as the extent of forests, fuel, and presumably biomass burning expanded while ice-covered lands were reduced [*Bigelow et al*., 2003; *Elias*, 2006; *Prentice et al*., 2000].

Cold and dry climate conditions during the Glacial favored extensive tundra and steppe vegetation [*Prentice et al*., 2000] in areas where forests now exist. We attribute the increase in levoglucosan flux (Figure 1B) from the Glacial to the mid-Holocene to changes in ice extent and in biomass productivity, where the combination of colder and drier glacial climate, lower CO_2 concentrations, and decreased insolation may have limited biomass growth and vegetation density [*Daniau et al*., 2012; *Marlon et al*., 2008; *Prentice et al*., 2000; *Whitlock et al*., 2010], and hence the fire activity recorded in Greenland during the end of the glacial period (until ~12 ka) [*Behl*, 2011; *Williams*, 2003].

3.2. Major Global Climate Parameters Alone Cannot Explain the 3.5–1.5 ka Maximum

During the glacial-interglacial transition (~12 ka), changes in levoglucosan flux in the NEEM ice core (Figures 1 and 2) generally follow changes in North Greenland Ice Core Project (NGRIP) δ^{18}O‰ (Figure 1D), including the abrupt increases in fire activity, temperature, and δ^{18}O‰ from ~12 until ~10 ka (Figure 2 and supporting information) [*NGRIP Project Members*, 2004]. Insolation and temperature are major drivers of biomass burning, where fire activity often increases with warming [*Daniau et al*., 2012; *Marlon et al*., 2013; *Power et al*., 2008]. The summer solar insolation over the high-latitude Northern Hemisphere (60°N) peaked ~11 ka, then decreased until the present day (Figure 1F) [*Berger and Loutre*, 1991]. This increased irradiation caused warmer summers in the Northern Hemisphere, with the warmest temperatures ~7 ka, and then steadily cooled until the start of preindustrial era [*Marcott et al*., 2013; *Renssen et al*., 2009]. The warmest temperatures at 7 ka are due to the final disappearance of ice sheets and the still-high (but decreasing) summer insolation. If temperature was the primary control, this cooling should have produced a continuous decrease in fire activity from ~7 ka to the preindustrial era. During the early to middle Holocene fire activity continually increased, even though insolation and summer temperatures decreased in the Northern Hemisphere. While summer insolation declined during the late Holocene, however, transient climate simulations of temperature changes [*Liu et al*., 2009] suggest that spring and fall temperatures may have increased in Beringia, northeastern North America, and elsewhere in the Northern Hemisphere [*Bartlein et al*., 2014], potentially explaining some of the late Holocene increases in fire activity.

Levoglucosan flux dramatically increases at ~4 ka and reaches its highest mean value between ~3.5–1.5 ka, even though the Holocene δ^{18}O‰ trend is stable and temperatures and insolation declined in the Northern Hemisphere [*Berger and Loutre*, 1991; *Marcott et al*., 2013]. Insolation and temperature changes are likely responsible for the rapid increase of levoglucosan during the glacial-interglacial transition, but these major climate drivers cannot explain the observed pyrogenic maximum ~2.5 ka and the subsequent decrease (supporting information). Many charcoal records also indicate higher than present day fire levels 3 ka in North America and Europe [*Marlon et al*., 2013].

We used an Earth System Model of intermediate complexity (CLIMate and BiosphERe-CLIMBER-2) and a land surface model (JSBACH), which dynamically represent vegetation, to simulate natural fire [*Brücher et al*., 2014]. Then, we compared fire activity trends with aggregated values of modeled burned area. The model simulation comprises the last 8 kyr and includes natural fire activity during the Holocene from potential levoglucosan source regions (supporting information). Simulations show constant biomass burning at all latitudes in Europe (Figure S12). However, all other continental-scale locations show increasing biomass burning during the past 4 kyr when burned area is simulated at 0–90°N and 30–90°N, whereas the higher latitude (45–90° N) locations show constant biomass burning. High-latitude simulations (45–90°N) encompass all sources of biomass burning plumes reaching Greenland, including the closest eastern Canadian source. However, simulations relative to potential North American and Eurasian sources do not show any features similar to the NEEM levoglucosan profile. These simulations (supporting information) further demonstrate that major global climate forcings do not support a maximum in fire activity nor a decrease in land burned during the late Holocene.

3.3. Local Climate May Partially Explain the Fire Maximum

Global climate changes do not explain the biomass burning signature recorded by levoglucosan flux. However, hemispheric to local climates may partially explain the biomass burning trend. Paleofire reconstructions encompassing all of boreal North America [*Ali et al*., 2009, 2012; *Girardin et al*., 2013; *Hély et al*., 2010] identified a gradual reduction in fire frequency during the last 3 kyr. Temperatures in Québec, Canada, peaked ~3.2 ka while maximum temperatures in Labrador are recorded ~4 ka before declining during the late Holocene [*Viau and Gajewski*, 2009]. Evidence supporting a shift toward wetter conditions were inferred at ~3 ka in Northern Québec [*van Bellen et al*., 2011; *Van Bellen et al*., 2013] and Labrador [*Viau and Gajewski*, 2009], and ~3.5 kyr in Southern Québec after a dry period between 4.8 and 3.4 ka, together with decreasing temperatures during the past 4.5 kyr [*Muller et al*., 2003]. Precipitation peaked ~1.5–2.0 ka in Labrador and Québec, while decreasing during the past 5.5 kyr in central Canada [*Viau and Gajewski*, 2009]. Western Canada demonstrates colder and wetter climate conditions with a decrease in fire frequency between 3.5 and 2.4 ka [*Hallett et al*., 2003]. In general, the literature suggests shifts toward wetter conditions in North America during the last approximately three millennia [*Ali et al*., 2012; *Girardin et al*., 2013; *Hély et al*., 2010].

The northern boundary of the boreal forest in North America and Europe retreated southward by ~200 km over the last 6 kyr [*Bigelow et al*., 2003; *Prentice et al*., 2000] where the shift is strongly asymmetrical around the pole. For example, in some sectors such as Québec-Labrador the tree line was south of the present position [*Bigelow et al*., 2003; *Prentice et al*., 2000]. This southward boreal forest shift during the last 6 kyr would have reduced the far northern source of fire, in part explaining the levoglucosan flux decrease during the last ~2.5 kyr, but does not explain the entire middle to late Holocene trend.

Modeled climate conditions also support the decrease of fire regimes in boreal North America during the late Holocene [*Ali et al*., 2012; *Girardin et al*., 2013; *Hély et al*., 2010]. *Hély et al*. [2010] argue that the downturn in fire frequency during the past ~5 kyr inferred from charcoal records from boreal North America are due to increasing available moisture and decreasing summer insolation (shorter fire season) caused by orbital forcing. *Ali et al*. [2012] conclude that the decrease in fire season length during the middle and late Holocene caused by the decrease of NH summer insolation, and the wetter climate (increase in precipitation) that characterized the northeastern North American boreal forests induced a sharp decrease in biomass burning ~3.0 ka. Therefore, a combination of regional late Holocene precipitation increases, superimposed on millennial-scale summer cooling [*Marcott et al*., 2013; *Viau and Gajewski*, 2009] may explain the decrease in Northern American fire activity during the last 3 kyr.

A shift to generally dry conditions is recorded across the Asian monsoon region beginning ~5 ka [*Li et al*., 2009] and more specifically was recorded at ~4 ka in China, followed by droughts extending until 3 ka, synchronous with the fall of many Chinese Neolithic cultures [see *Liu and Feng*, 2012, and references therein]. Vegetation reconstructions based on pollen analyses support steppe forests under cold and dry conditions between 4.0 and 2.8 ka in the southern Mongolian Plateau, with dry steppe vegetation under cooler and drier conditions until 1.4 kyr B.P. [*Sun and Feng*, 2013].

3.4. Did Human Activity Influence Biomass Burning?

Boreal forests in North America and Eurasia are likely the most important sources of fire products in Greenland ice cores during the present day and the past two millennia [*Zennaro et al*., 2014]. Currently, forests at Northern Hemisphere midlatitudes are not a major fire source due to few remaining forests and dense populations. Millennia ago midlatitude Northern Hemisphere forests were still extensive and humans may have burned this vegetation, even in the upper midlatitude regions (north of 45°N) of Europe and China [*Kaplan et al*., 2009, 2011]. Global circulation remained relatively stable during the middle to late Holocene, as suggested by stable oxygen isotope data demonstrating that circulation patterns of air masses reaching the Greenland ice sheet did not significantly change during the past 10 kyr [*NGRIP Project Members*, 2004; *Vinther et al*., 2006]. However, the previous southern extension of boreal forests several thousand years ago changed the location of large-scale fire sources where fire products may have followed trajectories starting from sources farther south. The possibility of such transport from the midlatitudes is supported by Chernobyl radiation products [*Davidson et al*., 1987; *Dibb*, 1989] and volcanic tephra from Vesuvius [*Barbante et al*., 2013] which are traced from their source well into the Arctic.

The scientific literature supports the possibility that forest clearance rates culminated 2.5–2.0 ka in much of Europe [*Ellis et al*., 2013; *Kaplan et al*., 2009, 2011]. A pollen-based reconstruction of land cover change demonstrates quantifiable forest clearance across much of Europe beginning ~4 ka [*Fyfe et al*., 2015]. Forests were reduced to near-modern levels by 2500 years B.P. in England [*Woodbridge et al*., 2014] and 2.2 ka in France [*Fyfe et al*., 2015]. This drop in the forest clearance rate in northwest Europe during the last 2500–2000 years is consistent with the decreasing levoglucosan flux after 2.5 ka.

Land cover change model results suggest that trends in fire history inferred from charcoal records are best explained by a substantial human influence on the climate system from biomass burning starting from the mid-Holocene [*Kaplan et al*., 2009, 2011]. When models take into account the increasing efficiency of land use [*Kaplan et al*., 2009], deforestation rates are higher between 3.0 and 1.5 ka than model results that do not incorporate technological changes and that indicate greater deforestation levels after ~1.5 ka. In a newer work, *Kaplan et al*. [2011] compare results from two land cover change models, one based on the HYDE 3.1 land use database [*Goldewijk et al*., 2011] and one based on KK10 [*Kaplan et al*., 2011] in which per capita land use declines with population density. Results substantially differ between the two models; the KK10 scenario estimates greater deforestation when land use per capita declines as population density increases, whereas the HYDE database predicts relatively low land use until the industrial era (10% of global land area is used at A.D. 1850). In the HYDE scenario land clearing in northeastern China, the Middle East, Europe, and South America began about 3 ka, while according to the KK10 scenario about 40% of land in these regions was already influenced by humans [*Kaplan et al*., 2011]. At ~2 ka about 60% of the total land area of Europe and China was cleared of forest vegetation in KK10, while in the HYDE scenario anthropogenically induced land cover change only slightly increased [*Kaplan et al*., 2011].

3.5. After the Fire Maximum: The Last Two Millennia

The relationship between population density and fire activity is nonlinear [*Guyette et al*., 2002] and nonmonotonic [*Bistinas et al*., 2013]. Data and models suggest that fire activity peaks when population density is at intermediate levels and decreases when population density exceeds a regionally varying threshold [*Bistinas et al*., 2013; *Syphard et al*., 2007]. Thus, the net impact of increasing population density can reduce fire frequency [*Knorr et al*., 2014]. Low population densities exist where climate is harsh and vegetation is low, while in densely populated areas fuel is limited and vegetation and landscapes are fragmented, thereby limiting the spreading of fire [*Archibald et al*., 2012, 2009; *Molinari et al*., 2013]. Humans may reduce fuel composition and density by introducing intensive agriculture and grazing, prescribed fire and substantial fire-suppression efforts. Increasing portions of the landscape occupied by settlements and infrastructure including road networks fragment fuel loads, reducing connections between fire-prone habitats and inhibiting the spread of fires that would otherwise burn larger areas in a continuous fuel bed [*Archibald et al*., 2009; *Guyette et al*., 2002; *Molinari et al*., 2013; *Prentice*, 2010; *van der Werf et al*., 2013]. Forest clearance rates may have dropped after 2 ka primarily because many Eurasian forests had already been cut. Landscape fragmentation may have further diminished fire activity ~2 ka thereby causing the decreasing trend [*Marlon et al*., 2013; *Molinari et al*., 2013].

4. Conclusions

Climate changes including summer North Hemisphere insolation and temperature affect boreal fire activity over millennial timescales. The levoglucosan flux reaching the NEEM site encompasses both North American and Eurasian boreal sources. Our results are consistent with charcoal-based evidence of broad-scale biomass burning throughout the Holocene. Regional charcoal syntheses, pollen analyses, and historical evidence generally ascribe the 3–2 ka fire peak evident in North America to climatic influences, while the fire peak in Europe is often linked to anthropogenic activities [*Marlon et al*., 2013; *Molinari et al*., 2013].

Northern Hemisphere temperatures and especially summer fire season temperatures remain stable or decrease between 3 and 2 ka. Therefore, major climate parameters and environmental changes (i.e., vegetation cover and bioproductivity) alone cannot explain the levoglucosan flux reaching Greenland during the middle to late Holocene. Given the lack of a plausible climate control for this pattern, coupled with the absence of paleoclimate evidence for any synchronous global climate change at this time, we argue that human activity associated with agriculture and land clearance provide the best explanation for observed trends in fire activity during the late Holocene. Millennial-scale variations in seasonal climate impacts on burned area and other fire regime characteristics, however, require further investigation.

Extensive deforestation in Europe between 2.5 and 2 ka [*Kaplan et al.*, 2009, 2011; *Molinari et al.*, 2013] is synchronous with the NEEM levoglucosan fire peak. Charcoal records and model results demonstrate both increased fire activity and the likelihood of anthropogenic forest clearing by biomass burning. The superimposition of human activities on increased fire activity due to regional climate factors results in a fire peak between 3.5 and 1.5 ka in the NEEM ice core, demonstrating a quantifiable early human impact on the climate system beginning about 4000 years ago.

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