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Invited review

Global biomass burning: a synthesis and review of Holocene paleofire records and their controls

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ABSTRACT

We synthesize existing sedimentary charcoal records to reconstruct Holocene fire history at regional, continental and global scales. The reconstructions are compared with the two potential controls of burning at these broad scales – changes in climate and human activities – to assess their relative importance on trends in biomass burning. Here we consider several hypotheses that have been advanced to explain the Holocene record of fire, including climate, human activities and synergies between the two. Our results suggest that 1) episodes of high fire activity were relatively common in the early Holocene and were consistent with climate changes despite low global temperatures and low levels of biomass burning globally; 2) there is little evidence from the paleofire record to support the Early Anthropocene Hypothesis of human modification of the global carbon cycle; 3) there was a nearly-global increase in fire activity from 3 to 2 ka that is difficult to explain with either climate or humans, but the widespread and synchronous nature of the increase suggests at least a partial climate forcing; and 4) burning during the past century generally decreased but was spatially variable; it declined sharply in many areas, but there were also large increases (e.g., Australia and parts of Europe). Our analysis does not exclude an important role for human activities on global biomass burning during the Holocene, but instead provides evidence for a pervasive influence of climate across multiple spatial and temporal scales.

1. Introduction

There have been large changes in fire during recent decades. In tropical regions, fire has resulted from deforestation and drought (van der Werf et al., 2010). The use of fire in these regions has increased the amount of crop and pasture lands for food production, but has unintentionally caused losses of biodiversity and is responsible for ca 12% of the global increase in CO_2 (van der Werf et al., 2009). In boreal forests, a key carbon reservoir, burned areas have been increasing in some areas and decreasing in others (Gillett et al., 2004; Girardin et al., 2009), but future susceptibility to large fires is very high (Flannigan et al., 2005; Girardin and Mudelsee, 2008), and increases in area burned are expected to account for the vast majority of increases in CO_2 from Canadian boreal forests (Amiro et al., 2009). In temperate regions, fires during the past few decades have grown in size from climate change (Westerling et al., 2006), rural depopulation (Pausas and Fernandez-Munoz, 2012), and, paradoxically, fire suppression (Marlon et al., 2012). Such changes in fire regime, and the possibility of further changes in a warming world, create management challenges throughout the world (Bowman et al., 2011). However, responding to these challenges is difficult because recent changes have been attributed to many factors, and because of our limited understanding of the intertwined nature of human and climate influences on fire.

Fire is episodic in nature, but is influenced by climatological and biological processes that operate over decades and centuries; it is therefore useful to adopt a long-term perspective to understand the

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causes of changing fire regimes. Sedimentary charcoal records can provide information about past changes in fire regimes, such as biomass burned and fire frequency, on multidecadal to millennial scales (Whitlock and Bartlein, 2004). There are hundreds charcoal records that span the Holocene, or past 11,700 years.

Variations in oxygen levels, climate and vegetation controlled fire for hundreds of millions of years (Scott, 2000). Climate has continued to control changes in biomass burning during the late Quaternary (Archibald et al., 2009; Marlon et al., 2009; Daniau et al., 2010). Humans have used fire for about 1 million years (Berna et al., 2012) in myriad ways, but there is debate about whether or when human influences on fire, particularly relating to the expansion of agriculture, exceed those of climate at landscape and broader scales (Vale, 2002; Williams, 2006; Marlon et al., 2008; Dull et al., 2010; Archibald et al., 2012). All the continents (except Antarctica) were settled by the beginning of the Holocene, thus this period provides crucial context for current fire–human–climate interactions and an opportunity to disentangle climate and human influences on changing fire regimes by examining regions with e.g., similar climatic histories and contrasting demographic, cultural and technological changes. Furthermore, the different histories of settlement in different parts of the world during the Holocene allow us to exploit "natural experiments" in the paleorecord, such as the late settlement of islands such as New Zealand (McWethy et al., 2010), for insights into fire history.

The combustion of biomass leads to the release of carbon dioxide (CO_2), methane (CH_4), carbon monoxide (CO), and nitrous oxide (N_2O), as well as aerosols and particulates such as soot. These emissions, particularly CO_2 and CH_4, affect global climate because they are long-lived greenhouse gases (Harrison et al., 2010). Changes in fire also affect the carbon cycle by altering vegetation growth and succession trajectories. As a result, fire–vegetation interactions can determine whether an ecosystem is a net source or sink of carbon, or whether it is in equilibrium with the atmosphere for a given interval (Prentice et al., 2011). Jacobson (Jacobson, 2004) hypothesized that, even with a continual sequestration of carbon after fires through vegetation regrowth, there will be a net accumulation of atmospheric CO_2 from biomass burning over time. One of the earliest continental-scale paleofire data syntheses (Carcaillet et al., 2002) argued that the parallel increase in biomass burning and CO_2 since the LGM supported this idea. However, the authors went further than this by suggesting that, given evidence for intensifying land use and human impact during the Holocene, the observed increase of fire during the Holocene was largely driven by human activities. Ruddiman (2003) has explicitly argued that human use of fire for land clearance during the early Holocene explains increases in CH_4 and CO_2 during the early and mid-Holocene and that these changes would have had a significant impact on climate. Although modeling experiments have been used to explore the potential for land-use changes to explain changes in atmospheric composition during the Holocene (e.g., Joos et al., 2004; Vavrus et al., 2008; Kutzbach et al., 2010) there has been no direct test of the underlying assumption that the fire record reflects the use of fire for land clearance.

Here, we review the primary broad-scale controls of fire and discuss several approaches for disentangling anthropogenic from climatic effects in the Holocene. For simplicity, we separate the Holocene into three intervals, including the early (11.7–8 ka), mid- (8–4 ka), and late (4–0 ka) Holocene. We then exploit existing syntheses of paleofire and archaeological data, and compare these with climate data to investigate the role of climate and anthropogenic impacts on fire. Support for the use of paleofire records based on microscopic and macroscopic charcoal has been rapidly growing in the past decade. Evidence that combining trends in charcoal accumulation rates (also known as "background" trends) from multiple sediment cores produces an index of biomass burned comes from empirically-based regional studies (Marlon et al., 2006; Vannière et al., 2011), simulation models (Higuera et al., 2007; Marlon et al., 2012), and comparisons with independent fire data such as carbon monoxide measurements from ice cores (Wang et al., 2010), historical data (Marlon et al., 2012), and dendrochronological data (Higuera et al., 2011). Furthermore, studies that have compared composite indices of biomass burning from trends in background charcoal with trends in fire frequency from the same records have demonstrated that they produce very similar results, at least in temperate forests (Marlon et al., 2009, 2012); more work is needed to determine whether this is also true in other forest ecosystems, and in grasslands and savannas.

We test the hypothesis that changes in fire during the Holocene correspond more closely with changes in human population or cultivated area rather than with climate changes inferred from reconstructions of temperature and effective moisture. We acknowledge that estimates of population and cultivated area are a poor proxy for the great diversity of human uses of fire during the Holocene, but such complexity should not preclude analyses of available data on changes in population, land-use, climate, and biomass burning. Rather, we argue that such data can provide important insights into human–fire–climate dynamics as well as constraints on those dynamics. We also note that the potential impact of variations in CO_2 on vegetation change and thus fuel availability are not considered. Based on the data presented, we argue that the complexity of human–fire interactions across space and time during the early and mid-Holocene was overridden by stronger climate effects that produced consistent, directional changes in fire activity evident at continental to global scales. In the late Holocene, the data are inconclusive as fire increased despite widespread cooling, yet human activities were presumably not synchronous at continental scales. In the last few centuries, human activity becomes clearly evident at broad scales because strong trends in biomass burning are inconsistent with climate trends.

1.1. Controls of fire

The geographic distribution of fire is driven by the interaction of abiotic and biotic processes that are scale dependent (Chuvieco et al., 2008; Parisien and Moritz, 2009; Krawchuk and Moritz, 2011). Atmospheric circulation patterns and moisture advection on meteorological time scales (i.e. minutes to days) determine the location, incidence and intensity of lightning storms that produce ignitions for the initiation of fire (Bartlein et al., 2008). Weather and vegetation condition also determine surface wind speeds and vapor-pressure gradients, and hence the rates of fuel drying, and soil moisture, which in turn affect the probability of combustion as well as fire spread. At these fine scales, the locations of ignitions and subsequent weather patterns, topography, vegetation, and local controls on fuel load and flammability can interact in complex ways, sometimes producing positive or negative feedbacks on fire behavior (Rollins et al., 2002; Gavin et al., 2006; Balch et al., 2008).

At broad spatial scales and on synoptic climatological time scales (i.e. months to seasons to years), the most important effects of climate on fire are changes in temperature and precipitation, which govern net primary productivity, and the abundance, composition, and structure of fuels (Carmona-Moreno et al., 2005; Moritz et al., 2012). In general, warmer temperatures are associated with increased burning through vegetation productivity and the incidence of fire-promoting climatic conditions (Daniau et al., 2012). The temperature–fire relationship, however, is mediated by precipitation–fire relationships (Kasischke et al., 2002; Balzter et al., 2007; Dennison et al., 2008). Fire responds differently to increases in precipitation depending on whether fuel abundance is initially a limiting factor for fire spread. In arid and semi-arid environments, increases in precipitation tend to increase fire (Veblen and Kitzberger, 2002), whereas in humid environments increased precipitation can reduce fire (Krawchuk and Moritz, 2011).

From a global perspective, fire thus occupies intermediate environments in terms of climate and vegetation (Archibald et al., 2009; Krawchuk et al., 2009; Parisien and Moritz, 2009). Annual area burned, for example, tends to be highest in regions with net primary productivity between about 400 and 1000 gC m^{-2} and with intermediate levels of moisture availability (between 0.3 and 0.8 on an index of actual to potential evaporation, AE/PE) (Harrison et al., 2010). Extreme environments tend to lack either sufficient fuel, such as in deserts, semi-arid regions and tundra, or sufficient dry periods, such as in tropical rainforests and along temperate oceanic margins, where fuels are usually too wet to burn. What emerges on multiple time scales is a relationship between climate and fire in which biomass burning is generally highest at intermediate levels of moisture at a given temperature.

On multidecadal-to-millennial time scales, climate-fire linkages are governed by changes in the composition of the atmosphere and the latitudinal and seasonal distribution of solar radiation (Millspaugh et al., 2000; Hély et al., 2010). These controls determine latitudinal temperature gradients as well as seasonal variations in land-sea temperature contrast. These gradients and contrast in turn influence large-scale atmospheric circulation patterns and the surface-climate anomalies they determine. Holocene climate variability has been more subtle than glacial–interglacial transitions during the Pleistocene, but climate changes over the past 12,000 years were still large enough to cause biome-scale vegetation and fire responses, such as the expansion and contraction of boreal forests during the Holocene (Ritchie, 1987; Hu et al., 1996; Carcaillet et al., 2006; Viau and Gajewski, 2009) and the greening of the Sahara in the mid-Holocene (Claussen and Gayler, 1997).

Interactions between vegetation and climate changes alter community composition, distribution, and structure – and thus fuel types and loads – to affect fire activity levels in different ways (Whitlock et al., 2010). One of the most important fire–vegetation interactions are those that occur in grasslands and savannas and affect tree–grass interactions (Bond and Keeley, 2005; Staver et al., 2011). Here, biomass productivity controls the occurrence of fire through the cycle of fuel build-up and drying. Fire may have promoted the expansion of grasses in the Miocene at the expense of forests, for example (Keeley and Rundel, 2005). Paleoecological data from North American prairies also documents the importance of fire in mediating tree–grass interactions by preventing the establishment of trees (Nelson and Hu, 2008).

Vegetation changes can also affect fire independently of climate through processes such as post-fire succession (Tinner et al., 2005; Colombaroli et al., 2008). Superposed epoch analysis of charcoal and pollen from two subalpine lakes from the western European Alps, for example, showed that fires only occurred after sufficient fuel build-up. Increases in arolla pine (*Pinus cembra*) always preceded fire, which was typically followed by a pattern of secondary succession that included increases in birch (*Betula*), ericaceous species (25–70 years), and herbs (50–100 years) before pine was re-established (Blarquez and Carcaillet, 2010). Vegetation changes unrelated to succession also affect fire. In some black spruce boreal forests of Alaska, for example, area burned was positively correlated with increasing tree density (Kasischke et al., 2002). The probability of fire also increases with stand age in these forests because fuel accumulation at decadal time scales increases the probability of fire occurrence and spread (Johnson and Gutsell, 1994; Schimmel and Granstrôm, 1997). Longer-term changes in fire independent of climate changes have also been documented in the boreal forest, when fire-prone species replace less flammable species (Hély et al., 2001; Higuera et al., 2009). Similarly, the late-Holocene invasion of Norway spruce *Picea abies* significantly reduced wildfire in humid parts of Scandinavia (Ohlson et al., 2011).

Humans can also affect fire regimes. Archaeological and ethnographic evidence indicates myriad uses of fire, including burning for food and resource procurement, warfare, travel, and clearance of pests and disease vectors in a variety of environments (e.g., grasslands, woodlands, and forests). Fire management was used to prevent uncontrolled wildfires in order to limit damage to settlements and agricultural resources (Iriartea et al., 2012). In southern Europe, fires were effectively suppressed after the establishment of orchards in the late Holocene through laws, manipulation of vegetation and landscape fragmentation (e.g., Tinner et al., 1999; Carcaillet et al., 2009).

Today, fires tend to be most frequent at intermediate levels of population density (Harrison et al., 2010). Fire is rare where populations are very low because such environments tend to have harsh climates and support little vegetation. Likewise, fire is rare in the most densely-populated (urban) areas because fuels are limited, landscapes are highly fragmented, and any fires that do occur are immediately suppressed. Modern studies of fire use in Africa savannas demonstrate that people control the timing, location, size, and spread of burning; such practices have likely been used for centuries. Controlled burns, fuel breaks, and fuel management strategies are used to create landscape mosaics that minimize interannual variability in an ecosystem that might otherwise produce large differences in area burned from year to year (Laris, 2011).

1.2. Disentangling human versus climatic impacts on fire

There is no direct way to distinguish human-set fires from wildfires in the paleorecord. The attribution of changes in fire regime to human impact is therefore largely based on the synchroneity of these changes and indicators of human activity, or more often simply human presence. In the best case, this would be archaeological evidence for the arrival and/or presence of humans in a region or for changes in human use of the environment. In the more general case, however, it relies on identifying signals of "human impact" on the environment such as changes in erosion rates, as recorded in sedimentary records, and changes in vegetation and land use, as recorded in the pollen record.

Archaeological evidence can be used to pinpoint the time of discovery of a region, of colonization, expansion of settlements, and cultural transitions including the beginning of agriculture (e.g., Tsukada et al., 1986; Graves and Addison, 1995). It can also be used to estimate the size of the human population or the intensity of exploitation of natural resources (e.g., Denham and Haberle, 2008; Smith et al., 2008). When such transitions are synchronous with changes in the fire regime, it is hard to escape the idea of a causal link between the two. However, establishing an empirical link depends crucially on the ability to demonstrate synchroneity, which in turn is dependent on having archaeological and fire records with sufficient sampling resolution and a good chronology (Berglund, 2003). Correlation analysis on well-dated, high-resolution charcoal and pollen data from central and southern Europe, for example, has provided clear evidence of the use of fire to open areas for arable and pastoral farming in the late Holocene (Tinner et al., 2005). In contrast, low-resolution sampling and chronological uncertainties have lead to ambiguity and debate in the nature of the fire–human–climate linkages in many places, such as Australia (Lynch et al., 2007; Mooney et al., 2011), China (Han et al., 2012), Central America (Dull et al., 2010; Power et al., 2012), South America (Heusser, 1994; Whitlock et al., 2007), and Madagascar (Bond et al., 2008; Willis, 2008).

Humans altered pre-industrial vegetation cover primarily through land clearance and conversion of land to agriculture (Lagerås and Bartholin, 2003; Parshall et al., 2003), but there are also examples of re-afforestation associated with the abandonment of agricultural land (e.g., Yeloff and Van Geel, 2007; Sköld et al., 2010). These changes are generally diagnosed from the pollen record, through identification of crop pollen (e.g., *Triticum* or *Zea mays*), increases in weedy or ruderal species (such as *Plantago lanceolata*, *Rumex* spp., and *Artemisia*), or decreases in tree species. The presence of crop pollen is diagnostic of human impact, but identifying crop pollen can be problematic. Some features of crop pollen grains, such as surface sculpture, pollen pore diameter, and annulus thickness and diameter are considered diagnostic (Joly et al., 2007). However, using the larger size of crop pollen compared to the natural progenitor – a technique that was used in the past – may not be diagnostic (Beug, 1961, 2004; Andersen, 1979; Poska and Saarse, 2006). Some types of steppe and coastal grasses produce pollen of a similar size to cereals, and some tropical cereal crops produce grains smaller than non-crop grasses (Bonnefille, 1972). Furthermore, cereals tend to release little pollen and have limited dispersal (Edwards, 1979; Berglund, 1985), which makes the identification of the transition to farming more difficult. Pollen of other crops (e.g., *Linum usitatissimum*, *Fagopyrum esculentum*) can be identified with more confidence (Beug, 2004), but are often even less abundant than cereal pollen. Thus increases in weedy species and ruderals have often been taken to confirm the transition to agriculture (Faegri and Iversen, 1989). Increases in weedy plants and ruderals occur in response to any kind of disturbance (e.g., storms, droughts, natural fires), however, and changes in the abundance of forest taxa can similarly be induced by climatic factors or changes in fire regimes, independent of human interventions. Thus, whereas the association of increased burning with evidence for rice paddy cultivation (Zong et al., 2007) provides a clear-cut example of human impact on local fire regimes, changes in the forest cover or the abundance of weeds are more equivocal. Furthermore, climate-induced changes in vegetation cover can trigger changes in other environmental conditions that have been used as indicators of human impacts, such soil erosion (Goudie, 2000). Where there are several lines of evidence that could reflect human impact, or when increases in fire occur when climate conditions are unfavorable to such increases, the attribution of changes in fire regimes to human impacts is strengthened (e.g., Edwards et al., 2006; Bradshaw et al., 2010). Nevertheless, caution needs to be exercised to avoid circularity in attributing changes to human impacts alone.

A second approach for separating climate and human impacts is to assess the synchrony of environmental changes (Gavin et al., 2006; Ali et al., 2009), based on the assumption that similar patterns of disturbance at widely-separated sites (i.e., hundreds or thousands of kilometers apart) with different environmental and human histories are unlikely to occur unless driven by factors that are similar at such scales, such as changes in climate conditions or vegetation composition and structure. Haberle and Ledru (2001), for example, compared charcoal records from Indonesia and Papua New Guinea and from Central and South America with climate data and information about human occupation; they concluded that patterns of synchrony and asynchrony in the records indicated that climate was the dominant forcing of fire activity in the two regions. Inverse correlation occurred during the Younger Dryas chronozone, when the tropics were strongly influenced by extratropical climate conditions, and positive correlations increased during the late Holocene, when tropical ENSO activity strengthened. One problem with this approach, as shown by examples from central and southern Europe, is that climate can also influence crop production at decadal to centennial scales, resulting in regionally-synchronous changes in land use and human populations (Tinner et al., 2003, 2009).

Other approaches include the process of elimination and modeling experiments. For example, in the western USA, Marlon et al. (2012) used the process of elimination to show that climate alone is sufficient to explain observed changes in fire. Temperature and drought data alone could predict 85% of the variability in charcoal levels during the past 1500 years in the western USA, including a pre-industrial maximum in burning during the Medieval Climate Anomaly and minimum during the Little Ice Age. This suggests that Native American impacts on fire regimes in western forests were secondary to those caused by climate changes. Dynamic landscape modeling has also recently been used to simulate natural vs. anthropogenic ignition during the Holocene; here fire simulations were compared to sedimentary evidence of burning and to palynological human indicators to demonstrate anthropogenic impacts on fire in the Alps during the late Holocene (Colombaroli et al., 2010).

Here we examine biomass burning trends at regional and continental scales, focusing on the trends in biomass burning in regions with different environmental, human and climate histories. We compare biomass burning reconstructions with available data on population and climate for each region. Our analysis does not provide a strong test of whether climate or people controlled fire in each region, but rather details the patterns in Holocene fire occurrence around the world and emphasizes features that are consistent with existing data on climatic and anthropogenic changes.

1.3. Approach and data

We analyze Holocene trends in biomass burning using 703 charcoal records (Fig. 1), including 658 from the Global Charcoal Database (GCD) version 2.0 (Daniau et al., 2012), 20 sites from Mooney et al. (2011) and 25 additional sites from published literature. Our aim is to explore the relationship between regional changes in biomass burning and potential drivers of those changes, specifically climate changes that act on fire directly or through changing vegetation and changes in land-use and/or fire-use resulting from human population changes or from cultural and technological innovations.

The original charcoal data were transformed following standard procedures described in Power et al. (2010): calculation of influx rates (i.e., the number, area or weight of particles/cm^2/year), standardization using a Box–Cox power transformation, and conversion to z-scores using a base period of 12,000–200 years before present (i.e., excluding the industrial period). As illustrated by the record from Lake Tilla, Nigeria (Salzmann et al., 2002), charcoal concentrations by depth (Fig. 2a) can look different when presented as influx values (Fig. 2b) due to variations in sedimentation rates (Goring et al., submitted for publication). Standardizing charcoal data using a power transformation accentuates small-scale variations and lessens the largest variations, but does not affect the timing of events (Fig. 2c). In order to construct composite curves for specific regions, each record was sampled at 20-year intervals (without interpolation) to reduce the influence of very high resolution records on the composite curves, some of which have annual resolution. Standardized, sampled records were then aggregated and smoothed using a lowess curve with 250 and 500-year moving windows. Bootstrap confidence intervals (95%) were constructed by sampling the records by site with replacement. Variability in the trend lines reflects changes in the average level of biomass burning for a given region, whereas variability in the width of the confidence intervals reflects variations in the number of available

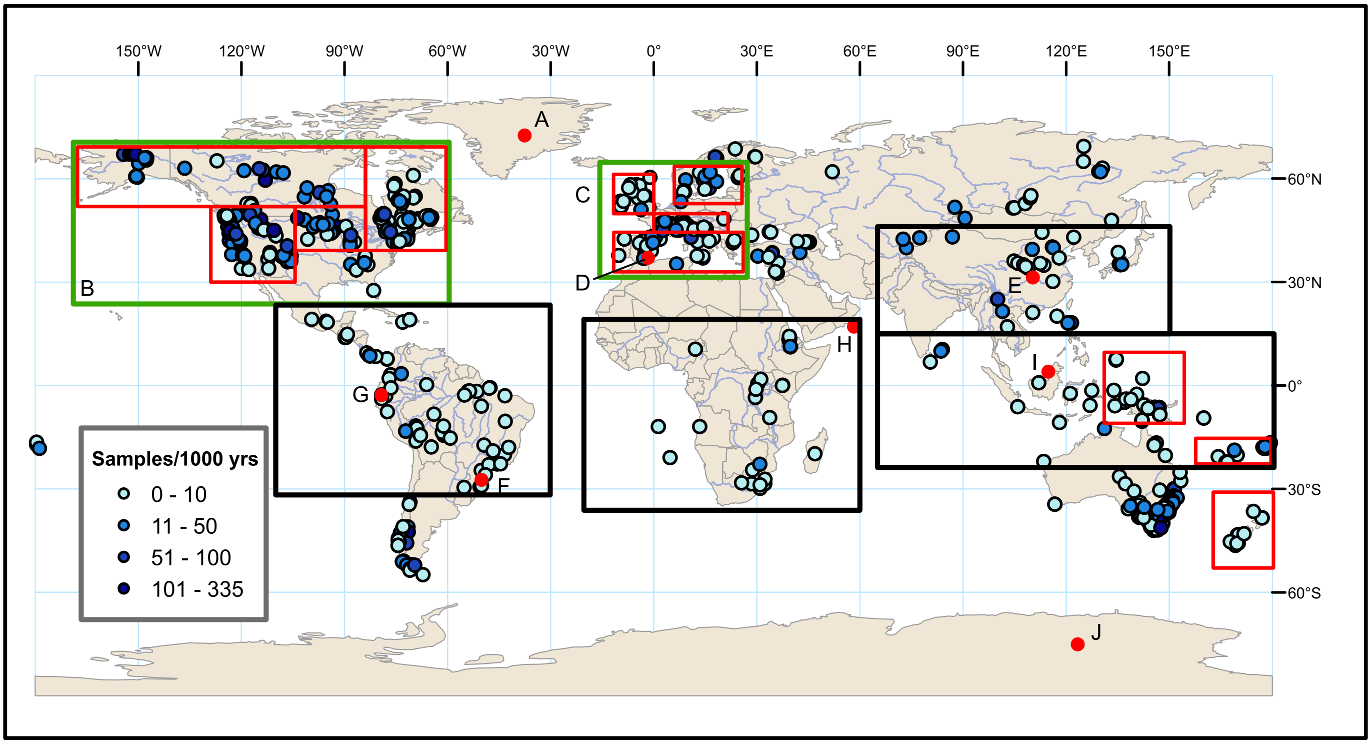

Fig. 1. Locations of charcoal records from the Global Charcoal Database v2 supplemented by data from Mooney et al. (2011). Blue symbols are colored by the number of samples per 1000 years in the record. Locations of paleoclimate records area marked by red circles, including A) NGRIP $\delta^{18}O$ ice core record (Andersen et al., 2004); B) North American pollen-based temperature reconstruction (green rectangle) (Viau et al., 2006); C) European pollen-based temperature reconstruction (green rectangle) (Davis et al., 2003); D) Sea-surface temperature (Cacho et al., 1999) reconstruction from the Alboran Sea; E) Sanbao Cave, China speleothem record (Wang et al., 2008; Dong et al., 2010); F) Botuvera Cave speleothem record (Cruz et al., 2005); G) Sediment color-based ENSO record from Laguna Pallcacocha in the southern Ecuadorian Andes (Moy et al., 2002); H) Qunf Cave, Oman speleothem record (Fleitmann et al., 2007); I) Northern Borneo speleothem record (Partin et al., 2007); and J) EPICA ice core records of deuterium, CO_2, and CH_4 (Jouzel et al., 2007; Loulergue et al., 2008; Lüthi et al., 2008).

records and the degree of dissimilarity between charcoal values at different sites for a given time.

We divided the world into regions with similar modern climates and based on the availability of 1) sufficient charcoal records to provide statistically meaningful composites; 2) independent paleoclimate reconstructions; and 3) distinct human histories. Examination of Pacific Islands, where the timing of the arrival of humans is relatively well known, provides a test of whether fire activity changes after colonization. Such a test does not account for changes in the density, duration, size, culture and technologies of human settlements after colonization, which would preferable, but such data do not exist yet at broad scales. We also examined four regions of North America and four regions of Europe to contrast areas where the timing of the spread of agriculture and forest clearance was different. Analysis of continental-scale fire reconstructions in extratropical and tropical regions provides further tests of the importance of human and climatic influences on fire history. Finally, we examine the global trend in fire for all 706 records during the Holocene to explore the possibility of a human influence through fire feedbacks on global atmospheric composition.

Area-weighted estimates of population and total cultivated area (the sum of crop and pasture land) were calculated from the HYDE dataset (Klein Goldewijk et al., 2010) and are presented on log scales for the regional and continental summaries. For the global analysis, we also provide an alternative reconstruction of Holocene changes in cultivated area based on the conceptual model of Ruddiman and Ellis (2009). The HYDE data assumes that total cultivated area is related to population, and that the amount of land required per person is constant. Ruddiman and Ellis (2009) have suggested that the amount of cultivated land required per person was greater in the mid-Holocene than in the late Holocene because cultivation efficiency has increased over time. Thus, in considering potential feedbacks from human fire-use on greenhouse gases, we scaled the HYDE estimates of cultivated area using the Ruddiman and Ellis (2009) model of decreasing land area used per capita over time. The resulting reconstruction of total cultivated area has a similar shape to the HYDE reconstruction, but in the adjusted curve the total area increases much faster in the mid- and late-Holocene, and variations after 2 ka are more pronounced.

Paleoclimate reconstructions representative of regional climate changes were obtained from published literature. Mean annual temperature reconstructions for Europe (Davis et al., 2003) are based on pollen records. The Davis et al. (2003) reconstruction does not capture the climate history of the western Mediterranean region very well, and so we compare the charcoal data from this regions with a record of mean annual sea surface temperatures (SST) from the western Mediterranean (Cacho et al., 1999). Viau et al. (2006) provide pollen-based reconstructions of July temperature for North America through the Holocene. Unfortunately, there are relatively few pollen-based quantitative climate reconstructions through the Holocene from other regions of the world. Consequently, we use $\delta^{18}O$ values in speleothem records that can be interpreted as changes in effective moisture for Asia (Wang et al., 2008; Dong et al., 2010), the American tropics (Botuvera, Cruz et al., 2005), sub-Saharan Africa (Qunf Cave, Fleitmann et al., 2007), and Australasia (Borneo, Partin et al., 2007).

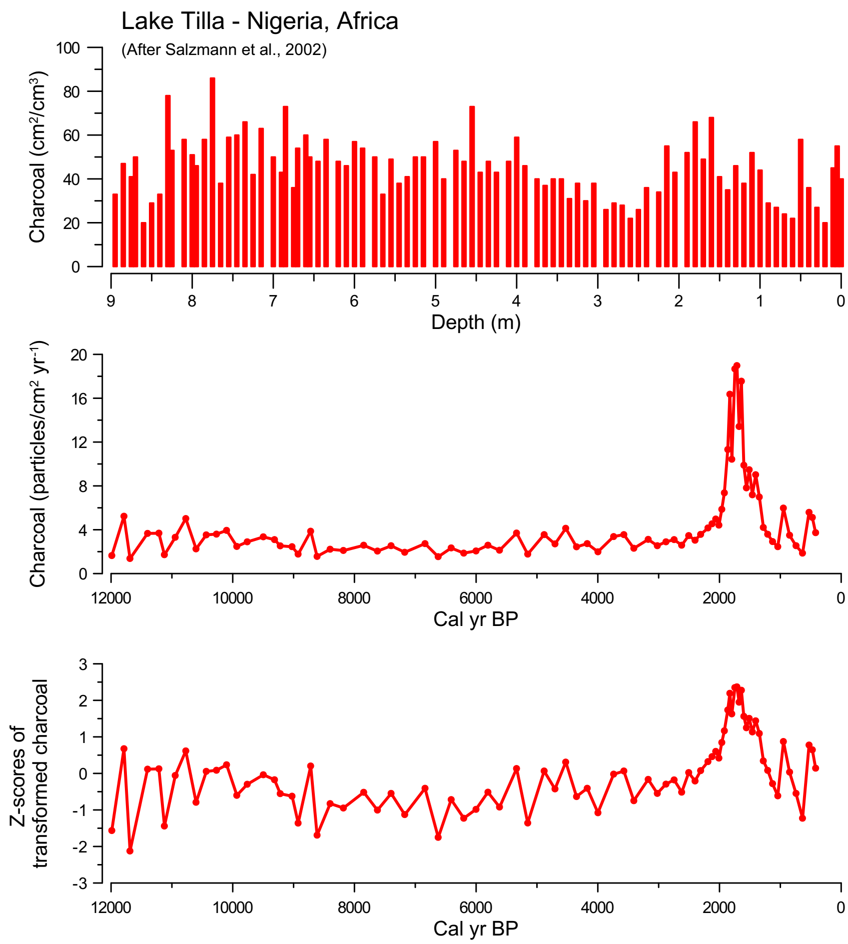

Fig. 2. Charcoal concentrations, charcoal influx, and standardized charcoal influx values from Lake Tilla, Nigeria (Salzmann et al., 2002).

2. Results & discussion

2.1. Pacific Islands

Late-settled islands offer an opportunity to distinguish pre-industrial human from climate impacts on burning, if the dates of settlement are known. Human colonization of New Guinea pre-dates the Holocene, but occurred ca 3 ka in Fiji and New Caledonia (Hope et al., 2009), and ca 1.2 ka in New Zealand (Wilmshurst et al., 2008). If expanding human populations led to increased fire in the Pacific Islands, we would expect fire to increase at the time of settlement in Fiji, New Caledonia and New Zealand, and that the pattern of changes would be different from that recorded in New Guinea. The climate history of this region is not well constrained. Two speleothem records from Indonesia (Partin et al., 2007; Griffiths et al., 2010) show different patterns, although both show an increase in effective moisture during the early Holocene and less millennial-scale variability subsequently. ENSO activity apparently increases from the mid- to late-Holocene (Clement et al., 2000; Moy et al., 2002; Rodo and Rodriguez-Arias, 2004). As a result, if climate were the primary control of fire in the Pacific Islands we would expect to see relatively high levels of burning in the early Holocene due to rapidly changing climate conditions and potentially again in the late Holocene due to increasing ENSO activity, with less burning in the mid-Holocene when temperature and monsoon-related precipitation are more stable and climate variability may have been reduced.

The charcoal record from New Guinea shows no overall trend during the Holocene. Biomass burning decreases during the early Holocene, increases abruptly at 8 ka, declines from 8 to 5.5 ka, increases abruptly again 5 ka, and then declines to about 0.5 ka. Local charcoal maxima occur at ca 11.7, 7.8 and 5 ka and local minima at 8 and 6 ka (Fig. 3a). The strongest feature of the New Guinea record is the large increase in biomass burning in the past two centuries. The records from Fiji and New Caledonia show a general upward trend in fire from 10 ka onwards (Fig. 3b–c). Biomass burning is variable but low prior to 4 ka and higher thereafter with local minima ca 2.8 and 1.2 ka, the latter date coinciding with human arrival (Fig. 3b). The past 1000 years of the Fiji and New Caledonia record is characterized by a strong increase in charcoal. Biomass burning in New Zealand is low and declining at the beginning of the Holocene. Several records show fire in the early Holocene, but almost none show peaks in burning in the mid-Holocene (Fig. 3c). There is an increase in biomass burning during the late Holocene, with peaks at 4 ka and 2 ka. There is a local minimum in burning just prior to Māori settlement and a strong increase since then (ca 1.2 ka).

A synchronous peak in fire also occurs in all island regions ca 7.5 ka; this predates the first record of relatively continuous forest disturbance associated with agricultural burning in Irian Jaya (Haberle et al., 1991). The establishment of cultivation at 5 ka is coincident with a large increase in fire in New Guinea (Fig. 3d), and also with local evidence of anthropogenic fire use on the island (Haberle et al., 1991; Haberle and David, 2004). However fire subsequently declines despite evidence for the intensification of agriculture, at least in the highlands (Roberts, 1998). Increased fire ca 5 ka in New Guinea has also been attributed to increased ENSO activity (Lynch et al., 2007). Similar increases in fire occur at this

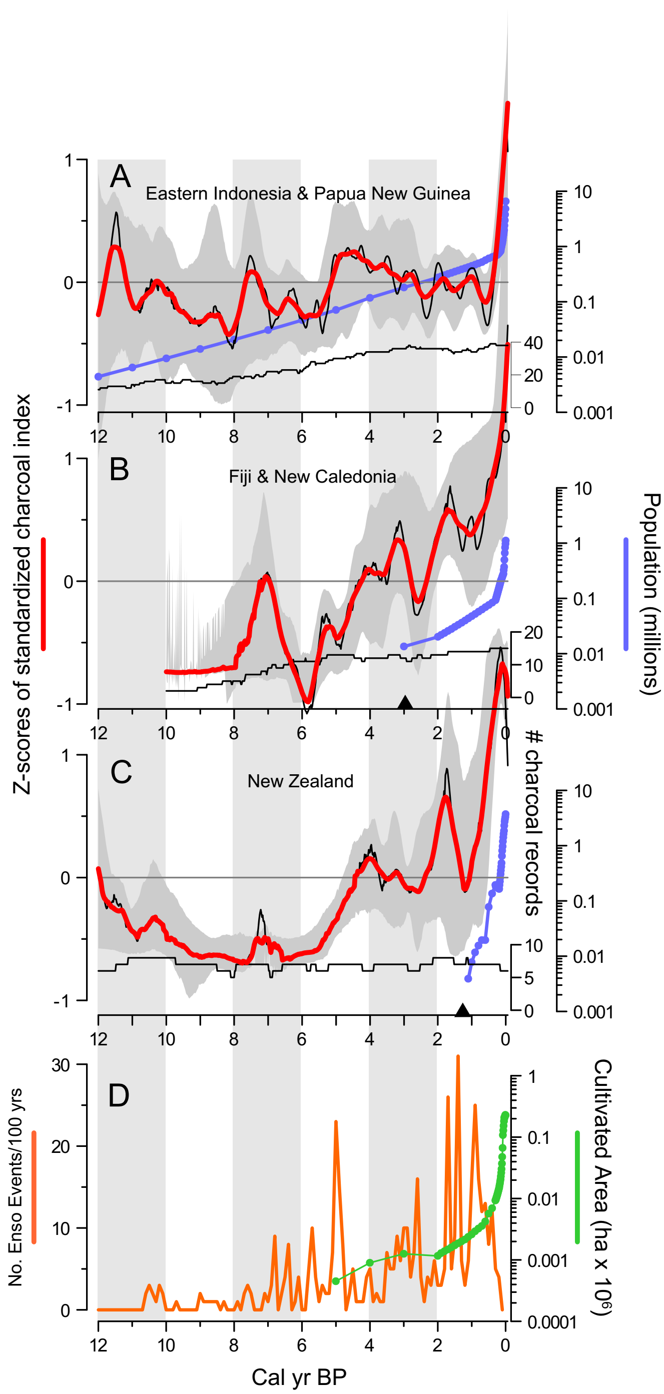

Fig. 3. Biomass burning trends for A) New Guinea, B) Fiji and New Caledonia, and C) New Zealand, all smoothed with a 500-year (red) and 250-year (black) line and shown with 95% bootstrap confidence intervals. Area-weighted population estimates (blue) are shown for each region. D) A reconstruction of the frequency of El Niño-Southern Oscillation (ENSO) events (Moy et al., 2002).

time in Fiji and New Caledonia, and also in New Zealand, despite the absence of people there. The late-Holocene increase in burning in New Zealand has been attributed to more frequent droughts associated with the intensification of ENSO circulation patterns (McGlone et al., 1992). Synchronous peaks in fire also occur across the Pacific Islands ca 2 ka, which post-dates settlement in New Guinea, Fiji and New Caledonia, but pre-dates settlement of New Zealand. Trends in fire during the past ~3000 years in Fiji and New Caledonia have been attributed to increased forest clearance and disturbance associated with human colonization (Hope et al., 2009). A paleoecological record from Lac Saint Louis in New Caledonia, for example, records an increase in charcoal accumulation, a decline in mangrove pollen, and increasing coastal savanna taxa that are contemporaneous with archaeological evidence of human activity at 3 ka (Stevenson, 2004). Most records from Fiji document increases in charcoal coupled with the permanent disappearance of primary forest taxa within the past 2000 years, although the timing of the impacts varies locally. The initial settlement date of 3 ka for Fiji and New Caledonia, however, is not marked by a large increase in burning as might be expected, but rather by an abrupt decline in burning (Fig. 3b). In contrast, the post-colonization increase in fire is very pronounced. McWethy et al. (2010) have shown that a dramatic increase in burning occurred within 200 years of Māori arrival at all but the wettest, high elevation sites; the composite curve (Fig. 3c) is consistent with this and shows an immediate increase in fire after Māori arrival.

In general, charcoal records from the Pacific Islands share a common pattern of moderate but declining levels of burning during the first half of the Holocene followed by low burning in the mid-Holocene and an increase in fire in the late Holocene. The regional nature of this pattern suggests it reflects climate changes. Synchronous increases in biomass burning occur ca 7.5 ka, 5.5 ka, and 2 ka, regardless of whether people are present or not. An abrupt increase in fire is coincident with the arrival of humans in New Zealand but not in Fiji and New Caledonia. The last two centuries are marked by exceptionally high levels of biomass burning in every region, which suggests that European settlement in the Pacific region had an important impact on fire regimes. However, recent decades are marked by a downturn in burning in New Zealand, consistent with the hypothesis that agricultural expansion eventually leads to forest landscape fragmentation and reduced fire.

2.2. Europe

Anthropogenic fire use in Europe is inferred from scattered records since the Mesolithic ca 8 ka. Animal husbandry and agriculture across Europe from the southeast after 8 ka, reaching the British Isles ca 5.7 ka (Roberts, 1998) (Fig. 4e). Early agriculture generally involved modification to existing woodland ecosystems rather than the complete replacement of forest by agroecosystems (Roberts, 1998); the latter occurred in the late Holocene. Thus, if we hypothesize that humans are responsible for Holocene increases in biomass burning in Europe, we would expect to observe increases in charcoal since 8–7 ka in southern and Central Europe and since ca 6 ka in the British Isles and in northeastern Europe, with more widespread increases in burning (associated with the initial destruction of forests for agroecosystems) after ~4 ka in all regions. Population growth also shows different patterns in different parts of Europe (Fig. 4). Population is estimated at about 1.2 million in the Mediterranean and 0.5 million in Central Europe at the end of the early Holocene, whereas similar levels were only reached in the British Isles and in Northeastern Europe during the late Holocene. Population (Fig. 4a–e) gradually increases in Europe until just after 2 ka, followed by a decline (minimal in northeastern Europe) reaching a local minima ca 1.5 ka, and a second decline between 600 and 700 years ago associated with the Black Death. Cultivated area for Europe increases most rapidly from 9 to 6 ka then more slowly from 6 to 2 ka; two short-lived declines after 2 ka parallel population changes but cultivated area generally increases again after ca 1.5 ka. Declines in burning are expected to be coincident with population declines in the past two millennia associated with disease outbreaks. Given the long-settlement history of the region, and the conversion of most of the land to agroecosystems and thus

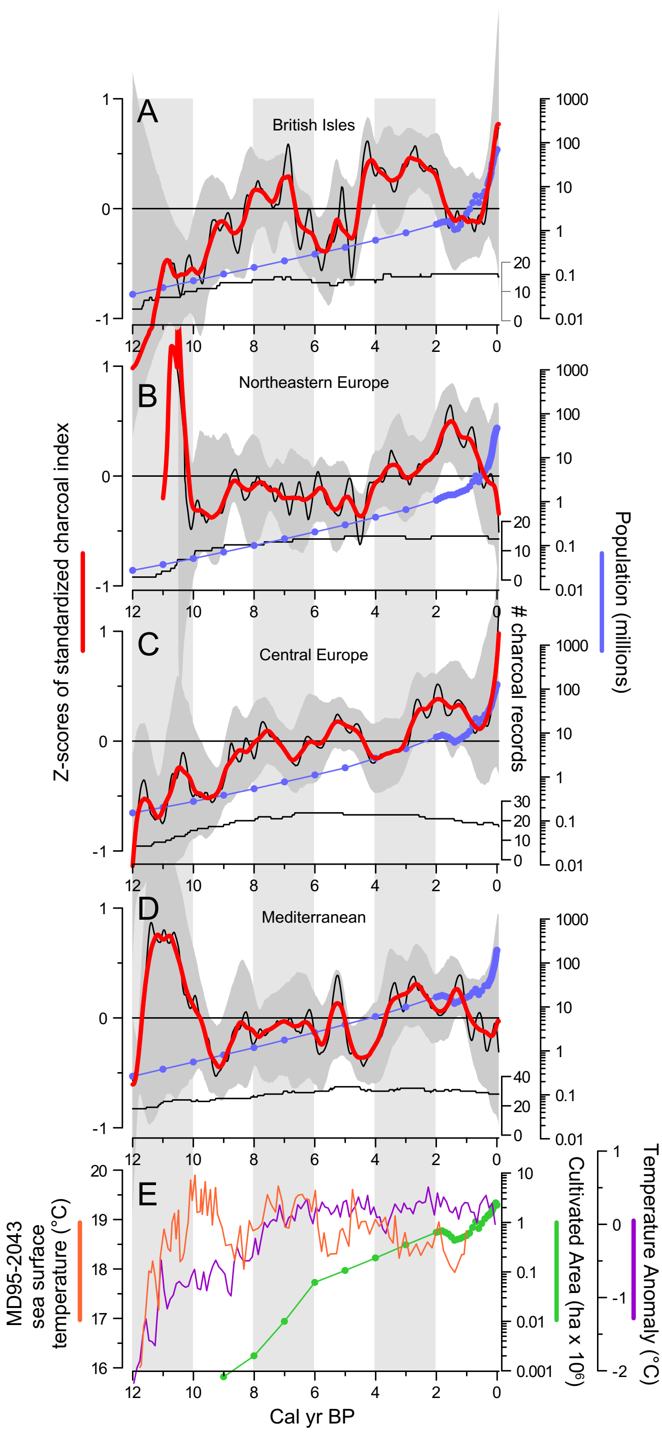

Fig. 4. European biomass burning trends for the A) the British Isles, B) northeastern Europe, C) Central Europe, and D) the Mediterranean, all smoothed with a 500-year (red) and 250-year (black) line and shown with 95% bootstrap confidence intervals. Area-weighted population estimates (blue) are shown for each region. E) Sea surface temperature reconstruction (orange) for the western Mediterranean (Cacho et al., 1999).

the loss of fuel for wildfires, it might be supposed that the incidence of fire will decline at some point in the late Holocene. However, this fire suppression signal is unlikely to be synchronous across Europe.

Pollen-based mean annual temperature reconstructions show a steady increase prior to 6 ka and stable temperatures thereafter in northern Europe (Fig. 4e) (Davis et al., 2003). In southern Europe, the early Holocene was characterized by high seasonal contrast, summer warmth and cold winters that lessened during the mid-Holocene, with conditions becoming drier in the late Holocene (Jalut et al., 2009). Sea-surface temperature (SST) in the western Mediterranean become cooler during the Holocene, consistent with the increasing aridity (Cacho et al., 1999) (Fig. 4e). The highly seasonal nature of Mediterranean climates, coupled with warm, dry summers, means they are more susceptible to fire than regions farther north. Increased precipitation, for example in the mid-Holocene, is therefore expected to reduce fire in the southern Mediterranean more dramatically than in northern Europe (Vannière et al., 2011). Holocene climate changes in Europe are expected to produce different patterns of change in fire regimes in northern and southern regions. In northern areas, low temperatures in the early Holocene are expected to produce low (but increasing) levels of biomass burning, whereas high climate variability due to changes associated with deglaciation may increase fire in some areas. In southern areas, increasing temperatures where fuels were not limited should lead to high biomass burning. Furthermore, short-term SST fluctuations (e.g., ca 9, 5.5, and 1.5 ka) indicate significant short-term climate variability, which could have influenced fire activity in the south throughout the Holocene.

In the British Isles, charcoal increases during the Holocene from very low levels at 12 ka to high levels in the first half of the late Holocene (Fig. 4a). Two intervals, from ca 7 to 5 ka and from 2 to 0.5 ka, show reduced charcoal abundances. Charcoal levels increase rapidly from 5 to 4 ka, and are highest between 5 and 2 ka. Charcoal accumulations also rise sharply in the past ~500 years. Fire, population and temperatures all increase during the early Holocene in the British Isles. The lack of local evidence for widespread human impacts on burning prior to 8 ka (Huang, 2002; Fyfe et al., 2003; Innes and Blackford, 2003) and a warming trend that is consistent with increased burning suggests that climate was the most important driver of fire. An interval of low fire activity regionally from ca 7 to 5 ka is linked to the *Ulmus* decline, but it is not clear whether low fire is due to climate or anthropogenic activities (Cayless and Tipping, 2002). Increased burning in the British Isles after 5 ka, however, is typically linked with human use of fire to open woodlands and improve grazing, and also with the expansion of heathlands (Fossitt, 1994; Innes and Simmons, 2000; Fyfe et al., 2003; Froyd, 2006; Ryan and Blackford, 2010). It is unclear, however, whether increased fire frequency due to human-caused ignitions caused heathlands to expand, or whether heath expanded as a result of grazing and thus increased fires. A decline in fire during the past 1000 years in the British Isles has been linked to increasing cultivation and changes in land-use (Fyfe et al., 2003).

Burning biomass is extremely high just before 10 ka in northeastern Europe, but then declines and remains low until 4 ka. Biomass burning increases from 4 ka to ca 1.8 ka, and then declines to the present (Fig. 4b). High early Holocene fire activity in the Northeast is registered at a few sites in Finland, Sweden and Denmark, and may reflect increased burning due to the reorganization of climate and vegetation at this time. Fire activity in this region in the early- and mid-Holocene is attributed to climate change, although a stronger human influence on fire is inferred during the late-Holocene (Sarmaja-Korjonen, 1998; Carcaillet et al., 2009; Olsson et al., 2010).

In Central Europe, charcoal generally increases throughout the Holocene (Fig. 4c). There are local maxima in fire at ca 8, 5, 2 ka, and at present, and local minima occur at 9, 7, 4, and 0.5 ka. Fire, population and temperatures increase during the early Holocene in Central Europe. The lack of local evidence for human impacts on burning coupled with a general warming trend (Fig. 4e) suggests that climate changes are the most likely cause of increased burning in the early Holocene. There is evidence for widespread human impacts in Central Europe from the mid-Holocene to present; many studies show a rise in fire accompanying a decline in forest cover and increased archaeological or archaeobotanical evidence of human activities (Clark et al., 1989; Carcaillet, 1998; Tinner et al., 1999, 2005; Fyfe et al., 2003; Andrič, 2007; Wehrli et al., 2007; Kaltenrieder et al., 2010; Kaal et al., 2011). The overall levels of burning in Central Europe are moderate during the mid-Holocene, however, suggesting that human impacts may have been intensifying but remained localized. Climate conditions may also have limited widespread burning in the mid-Holocene. Some sites show human impacts on burning only in the late Holocene (Pini, 2002), and many sites show a general intensification of anthropogenic disturbance ca 3–2 ka (Tinner et al., 2003).

Increasing fire from ca 4 to 2 ka in both Central and northeastern Europe is not consistent with declining or stable temperatures in the northern latitudes since the mid Holocene, and thus contradicts the idea that temperatures are the primary control on fire here. The "Roman Warm Period" (RWP, Lamb, 1977) ca 2 ka is broadly associated with warm, dry conditions in Europe (Harrison et al., 1993; McDermott et al., 2001; Patterson et al., 2010; Luterbacher et al., 2012) and could have triggered high levels of fire activity then, but obviously cannot account for the increase in fire starting at 4 ka. The decline in burning that occurs from the RWP to the LIA is not synchronous everywhere, but generally occurs a few hundred years after the reduction in population and cultivated area ca 1.5 ka. A second smaller and shorter decline in population and cultivated area occurs ca 0.5 ka, coincident with the LIA. Such patterns suggest that climate continued to influence fire here in the late Holocene, particularly through an increase in fire during the RWP and a reduction during the LIA. Biomass burning also increases rapidly in recent centuries but the cause is unknown.

Mediterranean records (Fig. 4d) show a peak in burning before 10 ka and low fire in the mid-Holocene. Fire increases in the Mediterranean in the late Holocene and declines during the past 1000 years. High early Holocene fire activity in the Mediterranean is consistent with the expansion of forests and increased summer warmth (Carrion et al., 2001) (Fig. 4e). An interval of low fire ca 9 ka is observed in the Mediterranean and parallels a shift to cooler SSTs in the marine sediment core from the western Mediterranean sea (Fig. 4d), however such changes appear to have been accompanied by relatively humid conditions and reduced fire on the continent (Jalut et al., 2000). Relatively low levels of burning in the mid-Holocene in the Mediterranean is consistent with humid summers and winters in southern Europe (Brewer et al., 2009; Roberts et al., 2011). A short but pronounced reduction in fire in the Mediterranean ca 4 ka is accompanied by high lake levels, suggesting wetter conditions (Vannière et al., 2008), but limited data and complex fire–vegetation–climate dynamics in this region warrant more careful and detailed analysis (e.g., Vannière et al., 2011).

The charcoal data from Europe as a whole show strong correlations with changes in temperature (Fig. 7b, e), suggesting that climate changes controlled biomass burning in the early Holocene, and continued to have a strong influence in the mid- and late-Holocene. Variations in human impacts on fire in Europe may provide an explanation for short-term shifts in biomass burning during the mid- and late-Holocene. The similarity between late-Holocene trends in fire in the Mediterranean and other European areas suggests a strong climatic influence on fire, although it is most often attributed to increased anthropogenic burning. If the widespread increase in fire in Europe from 4 to 2 ka and the subsequent pre-industrial decline is strongly influenced by human activities, then there must have been a shift in the relationship between population and fire ca 2 ka because the subsequent decline in fire is accompanied by a continued increase in population. Human activities may have reduced fire after 2 ka through landscape fragmentation, reduced forests and fuels, and less

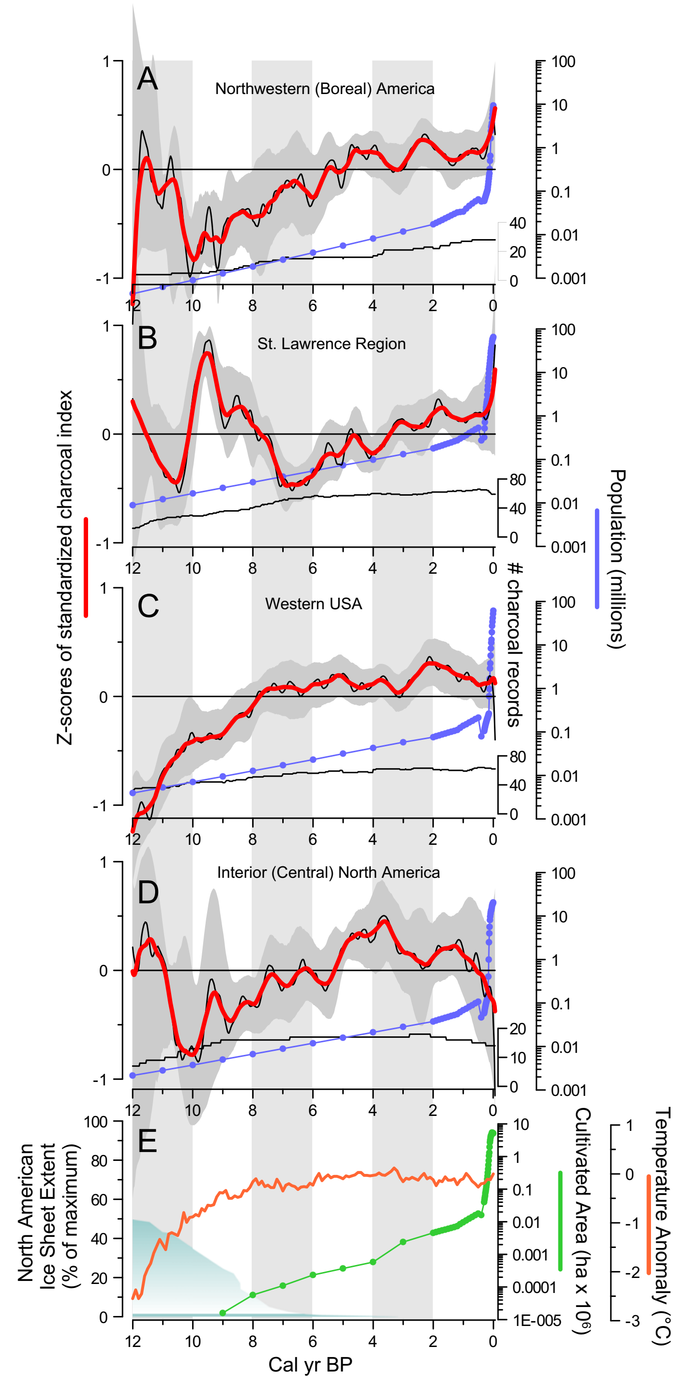

Fig. 5. North American biomass burning trends for the A) boreal region, B) St. Lawrence region, C) western USA, and D) continental interior, all smoothed with a 500-year (red) and 250-year (black) line and shown with 95% bootstrap confidence intervals. Area-weighted population estimates (blue) are shown for each region. E) Pollen-based July temperature reconstruction (orange) for North America (Viau et al., 2006), cultivated area of North America (green), and extent of the North American ice sheet (% of maximum, blue shading) from Dyke et al. (2002).

reliance on fire for land management. Reductions in fire globally in the past century have been attributed to the inadvertent consequences of landscape fragmentation through agricultural growth coupled with more active fire management (Marlon et al., 2008), and it is possible that in Europe this transition began much earlier (Carcaillet et al., 2009; Feurdean et al., 2012). This is not obvious from the data, however, because the relationships between population and biomass burning are different in different regions.

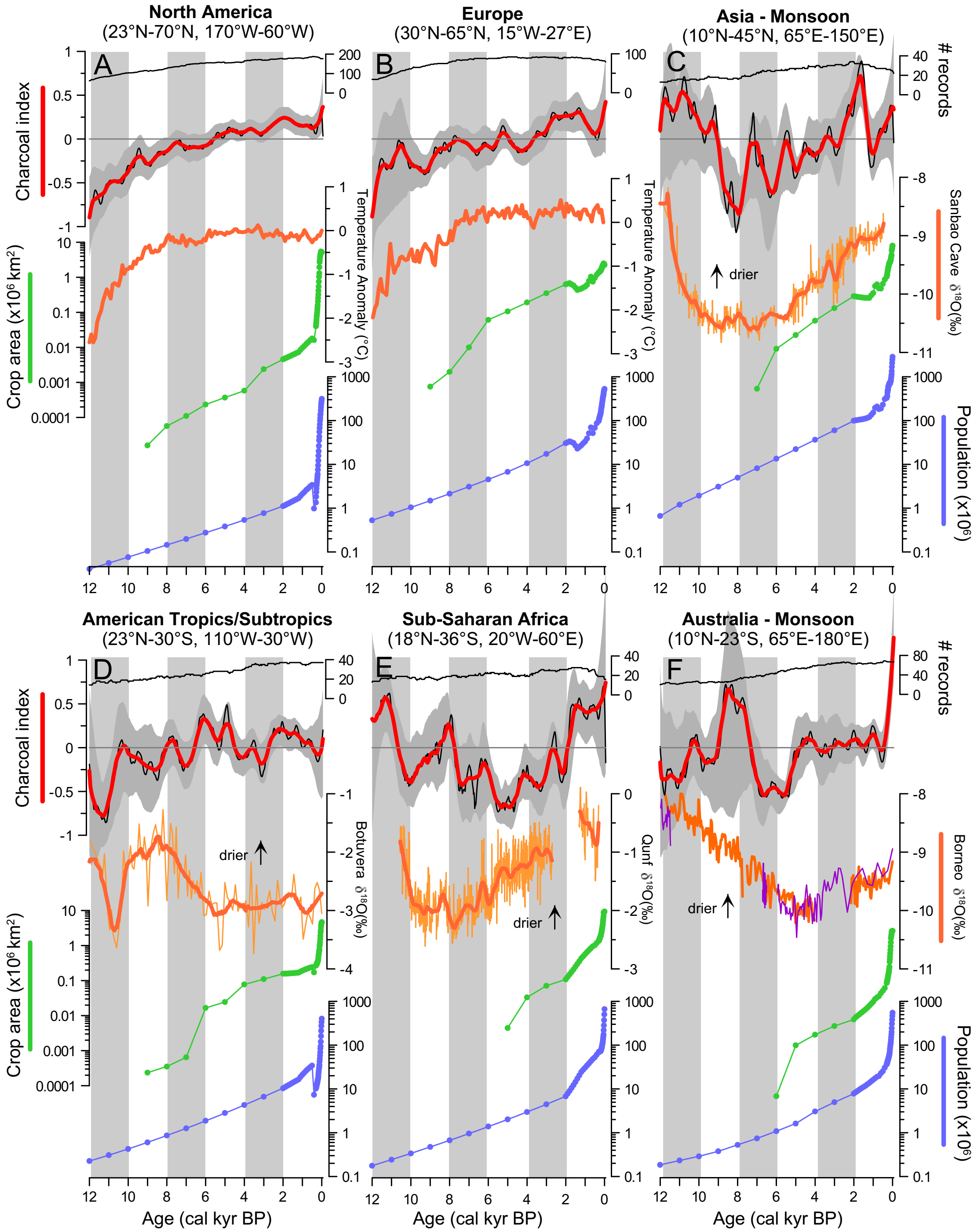

Fig. 6. Continental biomass burning trends smoothed with a 500-year (red) and 250-year (black) line and shown with 95% bootstrap confidence intervals for A) North America, B) Europe, C) the Asian monsoon area, D) the American tropics, E) Sub-Saharan Africa, and F) the Australian monsoon area. Paleoclimate data (orange) include pollen-inferred temperatures for North America (Viau et al., 2006), pollen-inferred mean annual temperatures for Europe (Davis et al., 2003), $\delta^{18}O$ values in speleothem records from Sanbao Cave, China (Wang et al., 2008; Dong et al., 2010), Botuvera Cave, Brazil (Cruz et al., 2005), Qunf Cave, Oman (Fleitmann et al., 2007), and northern Boreno caves (Borneo, Partin et al., 2007). Speleothem data were smoothed similarly to the charcoal data with a 500-year window (thick orange line). Area-weighted cultivated area (green), and area-weighted population estimates (blue) are also shown for each continent.

Unusually warm conditions ca 2 ka followed by unusually cold conditions ca 0.5 ka, potentially provide a coherent explanation for the general trajectories in fire across Europe as a whole even during the late Holocene. Trends in fire during the most recent centuries in Europe show strong regional contrasts, with sharp increases in fire in the British Isles and Central Europe, a sharp decline in North-eastern Europe, and more moderate changes in the Mediterranean. The 20th century reduction in fire in northeastern Europe, which occurred before active fire suppression and in the absence of grazing, has been attributed to reduced human ignitions inferred from the economic and cultural transition to modern agriculture and forestry (Wallenius, 2011). This transition to modern agri-culture and forestry appears to have been associated with *increased* fire in many records from the rest of Europe, for example in Hungary (Willis et al., 1997), Slovenia (Andrič, 2007), Switzerland (Conedera et al., 2009), and the Mediterranean countries Greece,

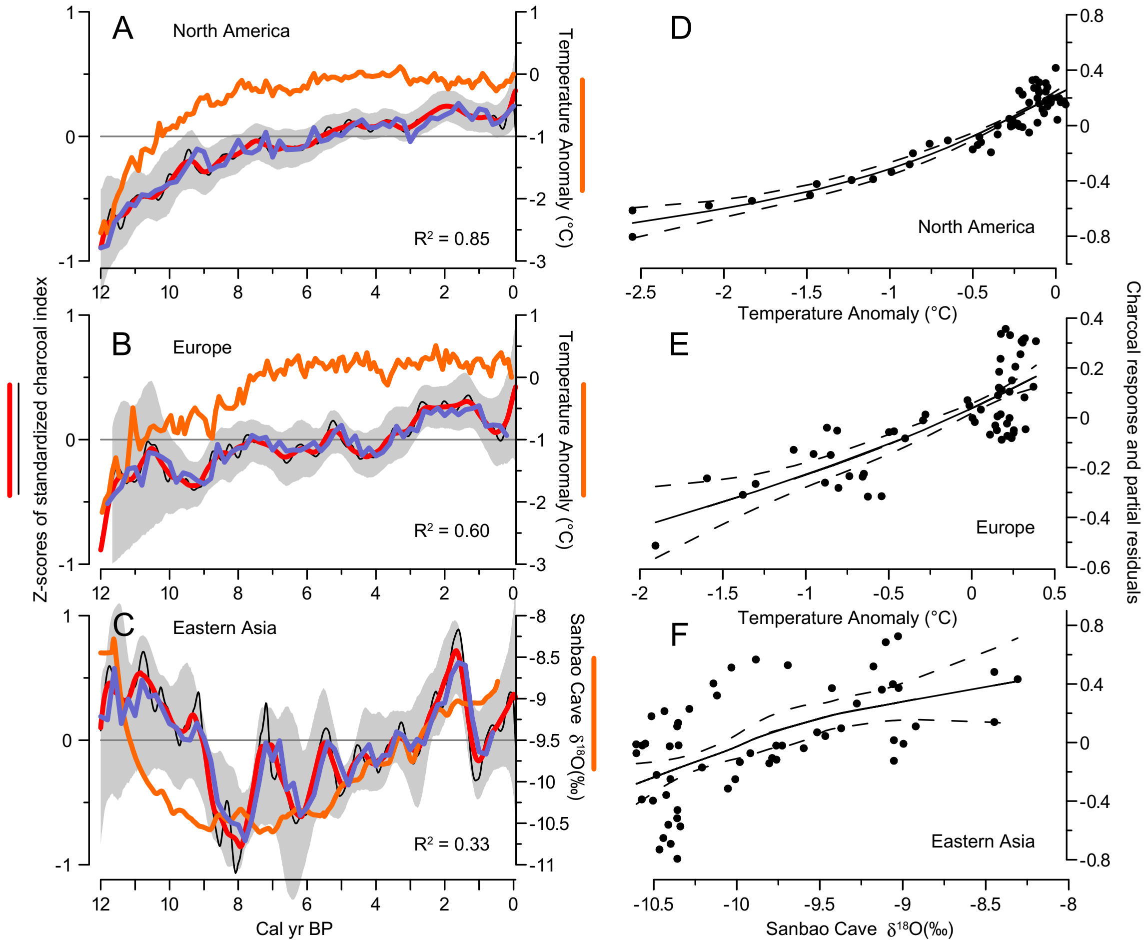

Fig. 7. Continental biomass burning trends smoothed with a 500-year (red) and 250-year (black) line with 95% bootstrap confidence intervals (gray) for A) North America, B) Europe, and C) the Eastern Asia monsoon area. Temperature data (orange) are from Viau et al. (2006) for North America and from Davis et al. (2003) for Europe. The smoothed speleothem record for Eastern Asia is from Sanbao Cave, China (Wang et al., 2008; Dong et al., 2010). Fitted values for the combined (GAM plus ARIMA) models appear in blue. Partial residual plots for each GAM are shown for D) North America, E) Europe, and F) Eastern Asia show the characteristic response of regional biomass burning: biomass burned increases nonlinearly with temperature, increasing more rapidly at higher temperatures; in Eastern Asia, biomass burned increases nonlinearly with decreasing moisture.

Italy, Portugal and Spain (Moreno et al., 1998). In the western Mediterranean, there is evidence that fire has been increasing in recent decades due to rural depopulation and farm abandonment that has led to reforestation and increased fuel loads (Pausas and Fernandez-Munoz, 2012).

The disentangling of climate, vegetation succession and human impacts on fire regimes is hampered by a paucity of review studies compiling local information, particularly about the timing of changes in human populations and cultural practices. Our analyses emphasize the importance of both climate and humans in driving changes in fire regimes, and suggest that a thorough and analytical reappraisal of the role of humans on past fire regimes is required. Such a reappraisal necessitates the compilation of archaeological and archaeobotanical evidence at regional scales, and the application of more quantitative analytical methods such as multivariate time-series and ordination analyses, generalized additive models (GAM) or dynamic fire modeling (Colombaroli et al., 2009, 2010).

2.3. North America

Trends in biomass burning show large regional differences in the early Holocene but are broadly similar in the middle and late Holocene. Intervals of high charcoal accumulation in the early Holocene are first recorded by sites in the Interior (Central) part of the continent, then in the Northwestern (Boreal) region, and later in the St. Lawrence region (Fig. 5a,b,d). In contrast, charcoal abundances increase steadily in the western USA from the early to mid-Holocene (Fig. 5e). In the mid- and late-Holocene, biomass burning increases gradually in the Boreal, St. Lawrence and Western USA regions to a pre-industrial maximum ca 2 ka, whereas the Central region shows a peak in fire ca 4 ka and a subsequent decline toward present. In the industrial era, biomass burning increases in the Boreal and St. Lawrence regions and declines in the West and Interior.

At a continental scale, July temperatures (Viau et al., 2006) rise rapidly from ca 12 to 10 ka and then more gradually from 10 to 8 ka, after which they remain stable (Fig. 5e). However, the time series show different trends in different regions (Viau et al., 2006). July temperatures were relatively high in the American midwest in the early Holocene, for example, and peaked ca 8 ka in northern Quebec before declining to lower levels during the mid- and late-Holocene. Several pollen and lake-level studies from both western and eastern North America also suggest that climate became cooler and wetter during the late-Holocene (Shuman et al., 2004, 2009a; Brown et al., 2006), but these trends are not evident in the Viau et al. (2006) reconstructions for these regions.

Population estimates for North America increase exponentially from less than 100,000 at 12 ka to 3.2 million by 1750 AD, with a short-lived decline in indigenous populations about 500 years ago following European colonization (Fig. 5). At 8 ka, population estimates are less than 2000 people for the boreal region, ~30,000 for the St. Lawrence, ~7000 for the interior, and ~13,000 for the western USA. By 1750 AD, estimated populations increased to ~80,000 for the boreal region, 1.3 million for the St. Lawrence region, ~150,000 for the western USA, and ~90,000 for the interior. Cultivated area increased from nearly zero at 8 ka to ~600 ha at 4 ka, to ~86,000 ha at 1750 AD, with the largest area in the western USA.

Spatially and temporally variable intervals of increased fire activity in North America in the early Holocene are consistent with the large, rapid reorganizations of atmospheric circulation and vegetation communities that occurred at that time (Shuman et al., 2002a, 2009b). High fire activity in the early Holocene in the St. Lawrence region (Carcaillet et al., 2002) has been attributed to the expansion of conifer forests (Richard, 1994; Williams, 2002) that resulted from warmer, drier conditions associated with a shift from anti-cyclonic-dominated circulation patterns to more zonal circulation after the collapse of the Laurentide Ice Sheet (Carcaillet and Richard, 2000). Climate-induced changes in vegetation composition are also implicated in the early Holocene increase in burning in the Brooks Range, Alaska (Higuera et al., 2009). In particular, early Holocene fire frequency increased with increasing warmth and with shifts from herb to shrub tundra; a subsequent decline in fire frequency occurred as shrub tundra shifted to deciduous woodlands. The gradual increase in biomass burning in the western USA has been attributed to increasing temperatures and the expansion of forests in e.g., the Pacific Northwest (Marlon et al., 2006; Whitlock et al., 2008). However, some sites in interior "summer wet" areas that receive moisture from monsoon systems show more abrupt increases in fire activity in the early Holocene (Brunelle et al., 2005; Anderson et al., 2008). Large shifts in biomass burning during the early Holocene in the continental interior also match a peak in July temperatures inferred from the regional pollen data in Viau et al. (2006). Drying in the Midwest during the early Holocene was time-transgressive and either caused ecotones to shift eastward or progressive drying from west to east (Williams et al., 2010).

A gradual rise in biomass burning occurs from the mid- to late-Holocene across North America, reaching a maximum ca 2 ka in all regions except the Central region, which reached a maximum ca 4 ka (Fig. 5a–d). Population and cultivated area increase most rapidly in the late Holocene but remain low compared with levels in Europe and Asia (Fig. 6). There is no known evidence for the human use of fire in the Boreal region even in the late Holocene. Conditions became warmer and effectively wetter during the mid- to late-Holocene transition in both the Boreal and St. Lawrence region and apparently remained that way during the late Holocene (Lynch et al., 2004; Viau and Gajewski, 2009). It is therefore difficult to explain relatively high biomass burning in the Boreal region during the late Holocene, including the peak at 2 ka and the subsequent pre-industrial decline with available climate or population data.

In the St. Lawrence region, Bremond et al. (2010) show that most of the burning derive from the Taiga Shield in the late Holocene, which is attributed to increased seasonal drought (Carcaillet and Richard, 2000). Abrams and Nowacki, 2008 argue that burning by indigenous populations played an important role in maintaining vegetation communities throughout eastern North America in the late Holocene. However, paleofire studies from the northeastern US do not support this (Parshall and Foster, 2002; Oswald et al., 2010). Munoz et al. (2010) synthesized archaeological, pollen, climate and charcoal data in the northeastern US during the Holocene and found a correlation between major environmental–climatic transitions and cultural and demographic shifts, with the last shift occurring ca 3 ka, coincident with the transition from Late Archaic to Woodland cultures. The authors inferred a cultural response to changing environmental conditions, but their data indicate that increased burning occurred during a population decline ca 3–2 ka. The brief Early Woodland phase coincides with the Holocene maximum in burning in the St. Lawrence region, and may have contributed to increased fire activity at this time.

The Holocene maximum in burning in the western USA also occurs ca 2 ka (Fig. 5c). There is some evidence for Native American fire use in the late Holocene from paleofire studies in coastal regions (Brown and Hebda, 2002) and river valleys (Walsh et al., 2010) as well as farther inland (Roos et al., 2010; Scharf, 2010). However, increased drought has also been invoked to account for high fire activity ca 2 ka in British Columbia (Hallett et al., 2003; Lepofsky et al., 2005). Marlon et al. (2012) found that variations in temperature and drought were sufficient to explain changes in biomass burning since 1.5 ka, including the maximum in burning during the MCA and minimum burning during the LIA.

Central North America shows a unique pattern of late-Holocene burning, with peak fire activity (ca 5–3 ka). A plausible explanation for this is that decreasing mid-continental aridity and increasing effective moisture, especially in summer, promoted grass productivity, resulting in higher fuel loads and more frequent/intense burns that inhibited woodland expansion (Umbanhowar et al., 2006). As the late-Holocene progressed, continued decreases in aridity had an opposite effect and reduced fire activity (Nelson et al., 2006).

In general, climate and vegetation changes explain North American biomass burning trends in the early Holocene. Mid- and late-Holocene changes in fire activity show broad similarities which suggest a climatic forcing, although paleoclimate data suggest a trend toward wetter conditions and stable or cooler temperatures after 6 ka which is not consistent with increasing biomass burning. It also seems unlikely that Native American populations were responsible for the maximum in fire at 2 ka as population maxima are often after this date, and the synchroneity of peak fire across the continent would imply a common fire practice and intensity of burning in four different regions. The adoption of agriculture cannot account for the increase in fire ca 2 ka either, as the timing differed among regions. In the northeast, for example, maize agriculture was only adopted about 1000 years ago (Doolittle, 2000). In the past few centuries, human impacts on fire in North America – particularly fire exclusion – are well-documented and widespread (Nowacki and Abrams, 2008; Marlon et al., 2012).

2.4. Fire–climate–people linkages at continental scales

The continental-scale trends in biomass burning in Europe and North America can be compared with similar continental-scale trends for Asia, the American tropics, Africa and Australasia. Paleofire records from these regions are much less abundant than in North America and Europe, and data on regional climate and human activities are also very limited. Although the paucity of data urges caution in interpretation, there are distinctly different trends in biomass burning at continental scales that appear largely explicable in terms of changes in climate (Fig. 6).

Biomass burning in North America and in Europe generally increases throughout the Holocene. In the monsoon areas of Asia and Australasia and also in Africa, burning is higher in the early and late Holocene than in the mid-Holocene, while the record from the American tropics shows lower levels in the early Holocene, highest levels in the mid-Holocene and somewhat lower levels again in the late-Holocene. However, shorter-term variability is superimposed on these long-term trends. For example, there is very low biomass burning in Asia ca 9–8 ka and very high biomass burning in Australasia (and Africa to a lesser extent) around the same time. Fire increases from 4 to 2 ka in Africa and Asia and declines from 2 ka onwards, but there is little change over these intervals in Australasia.

Humans were present on each of the six continents throughout the Holocene and population growth trends are similar in each sector (trending upward), although population levels varied widely. According to the HYDE data sets, there were about 140,000 people in North America by 8 ka, 900,000 in the American tropics, 500,000 in the Australasia monsoon area, 700,000 in Africa, 2 million in Europe, and 5 million in the Asian monsoon area. Populations increased in all six sectors between 8 and 4 ka but total numbers were highly variable. The last 4 ka has been marked by further large increases in population in most sectors according to the HYDE database.

Human fire-use is often assumed to have limited effects even in Europe and Asia during the early Holocene because population levels were low (but see Pinter et al., 2011). Declines in biomass burning in Asia and Africa during the early Holocene, for example, are inconsistent with population growth during this interval. Population increases during the interval from 8 to 4 ka were accompanied by increases in biomass burning in some regions (Fig. 6a, c, f) but not in others (Fig. 6b, e, d). Several large initial increases in estimated cultivated area (e.g., in the American tropics and Africa) are accompanied by short-lived increases in biomass burning that may reflect early intervals of land clearance (Fig. 6a, c–f). Burning in Europe, however, was relatively constant in the mid-Holocene despite large increases in cultivated area (Fig. 6b). Also, in Asia cultivated area and population increase after 8 ka, and there is local evidence for human impacts on fire (e.g., Shu et al., 2010), but biomass burning at the regional scale is highly variable from 8 to 4 ka (Fig. 6c). The abrupt increase in fire in the Australasian monsoon region between 6 and 5 ka, which has been associated with the intensification of human settlement and cultural development in the region (Haberle and David, 2004) is not supported by a direct comparison of fire records with continental-scale reconstructions of population and land-use intensity based on archaeological data, which suggests that the increase in fire pre-dates the human changes (Mooney et al., 2011). The attribution of long-term changes in fire regime at broad continental scales to human activities remains basically unproven.

Increased burning in the early Holocene in North America and Europe is associated with large, rapid climate changes in many sub-continental scale regions. High fire activity at regional scales is consistent with the widespread reorganization of ecosystems (Tinner and Lotter, 2001; Shuman et al., 2002b) that occurred in response to climate changes associated with deglaciation (Alley and Agustsdottir, 2005). There is also evidence from other time periods that fire tends may increase in response to abrupt climate changes regardless of the direction of change (Marlon et al., 2009, 2012).

Hydrologic changes inferred from $\delta^{18}O$ values in speleothem records imply that trends in effective moisture were similar in the Asian monsoon region and in Africa, showing wettest conditions ca 10–6 ka. Dry conditions in Asia before 10 ka (Fig. 6c) are consistent with the observed high levels of fire. Wang et al. (2005) attribute high fire activity at this time to cool, dry conditions that supported frequent natural fires. Reduced biomass burning after 10 ka is consistent with insolation-induced intensification of the Asian monsoon during the early- to mid-Holocene, and the late Holocene increase in biomass burning is also consistent with increased aridity after 6 ka (Fig. 6c) (Zhao et al., 2009). High early Holocene burning and lower levels of mid-Holocene fire are also consistent with a shift from arid to humid conditions in the early- to mid-Holocene in Sub-Saharan Africa (Kropelin et al., 2008) (Fig. 6). Nevertheless, shorter-term changes in effective humidity implied by the speleothem records are not always consistent with reconstructed changes in biomass burning. There is no decrease in effective moisture after 2 ka in the Sanbao Cave record paralleling the marked decline in biomass burning, for example, although the Dongge Cave record does show an inflection point ca 2000 years ago (Dykoski et al., 2005).

Hydrologic changes in the American tropics and in Australasia are different from those in Asia and Africa and also from each other, but both show drier conditions in the early than late Holocene. A pronounced excursion toward wetter conditions at ca 11 ka at Botuvera Cave (Fig. 6d) is paralleled by a marked reduction in biomass burning and the subsequent return to dry conditions is also paralleled by increasing fire. After 8 ka, however, there is little consistency between the two records, with maximum fire during the latter part of the Holocene when conditions were apparently wetter. Similarly, the changes in biomass burning in Australasia are not easily explained by the reconstructed moisture changes from the speleothem records from Borneo (Fig. 6d). The marked increase in fire between 6 and 5 ka, when identified in individual records, has been attributed to an increase in ENSO activity (see e.g., Haberle, 2005). One difficulty here is that the timing of the increase in ENSO is still a matter of some debate with the onset being placed at various times between 6.5 and 3 ka (see discussion in Mooney et al., 2011). Considerably better and more regionally representative information about climate changes during the Holocene are required in order to establish definitively whether the biomass burning records from the American tropics and Australasia are driven by climate.

The dependence of the long-term variations in charcoal influx on climate can be illustrated for the three Northern Hemisphere regions, which have continuous regional paleoclimatic records (Fig. 7). The construction of statistical relationships between paleo time series are complicated by 1) smoothing of the charcoal influx data from multiple records; 2) autocorrelation in the charcoal and paleoclimate data; 3) differences in the geographic representativeness of each time series, 4) the limited availability of paleoclimate data at regional scales, and 5) the inherent nonlinear relationships between biomass burning and climate (Daniau et al., 2012). To address these issues, we used Generalized Additive Models (GAMs) (Hastie and Tibshirani, 1990; Wood, 2006) to link regional charcoal influx and climate records; for North America and Europe, we used pollen-based reconstructions of mean annual temperature (MAT, Viau et al., 2006; Davis et al., 2003, respectively), while for monsoonal Asia, we used the $\delta^{18}O$ record of Wang et al. (2008), an index of moisture availability.

To reduce the impact of pseudo replication, we sampled the 500-yr smoothed charcoal influx records and associated climate records at 200-yr intervals. The individual models were all significant ($p < 0.001$) with R^2 values of 0.882, 0.603 and 0.330 for North America, Europe and Asia, respectively. Partial residual plots for each model (Fig. 7D–F) show relatively simple curvilinear relationships between charcoal influx and climate, with those for MAT for North America and Europe concave upward, and that for $\delta^{18}O$ (moisture) for Asia convex upward, consistent with the response surfaces described by Daniau et al. (2012) for both paleo charcoal-influx data and modern satellite remote-sensing burned area data, which show that biomass burning increases with increasing temperature, and is highest at intermediate levels of effective moisture.

The residuals from each model were serially correlated, but were each represented well by a low-order autoregressive, integrated, moving-average (ARIMA) model (Box and Jenkins, 1976). For Europe and Asia, an ARMA(1,1) model (with single autoregressive and moving average terms) fit the residuals well, while for North America, an IMA(1,1) model fit well. (This model is equivalent to an ARMA(1,1) with a first-order autoregressive term close to unity). The two classes of models (GAM and ARIMA) can be iteratively fit to the data, but the initial combination of models produced residuals that were not significantly correlated (p-values greater than 0.15, 0.90 and 0.07 for North America, Europe and Asia, respectively, for the Ljung and Box (1980) test for serial autocorrelation of residuals). No signs of model inadequacy were discernible with the usual diagnostic checks. The R^2 values for the joint (GAM plus ARIMA) models were 0.951, 0.889 and 0.878 for North America, Europe and Asia, respectively.

The correlations between the observed charcoal influx values and predicted values based on regional and continental-scale temperature and moisture data, and in a way that accounts for autocorrelation in the charcoal time series, is consistent with our conclusions that millennial-scale variability in biomass burned can partly be attributed to variations in temperature and effective moisture during the Holocene, at least in the Northern Hemisphere. Specifically, biomass burned increases with increasing temperature, and in Eastern Asia, biomass burned increases with decreasing moisture availability. The development of dynamic global vegetation and fire models that can be driven by transient climate simulations will greatly improve our ability to identify the specific mechanisms through which climate changes influence fire at regional to global scales.

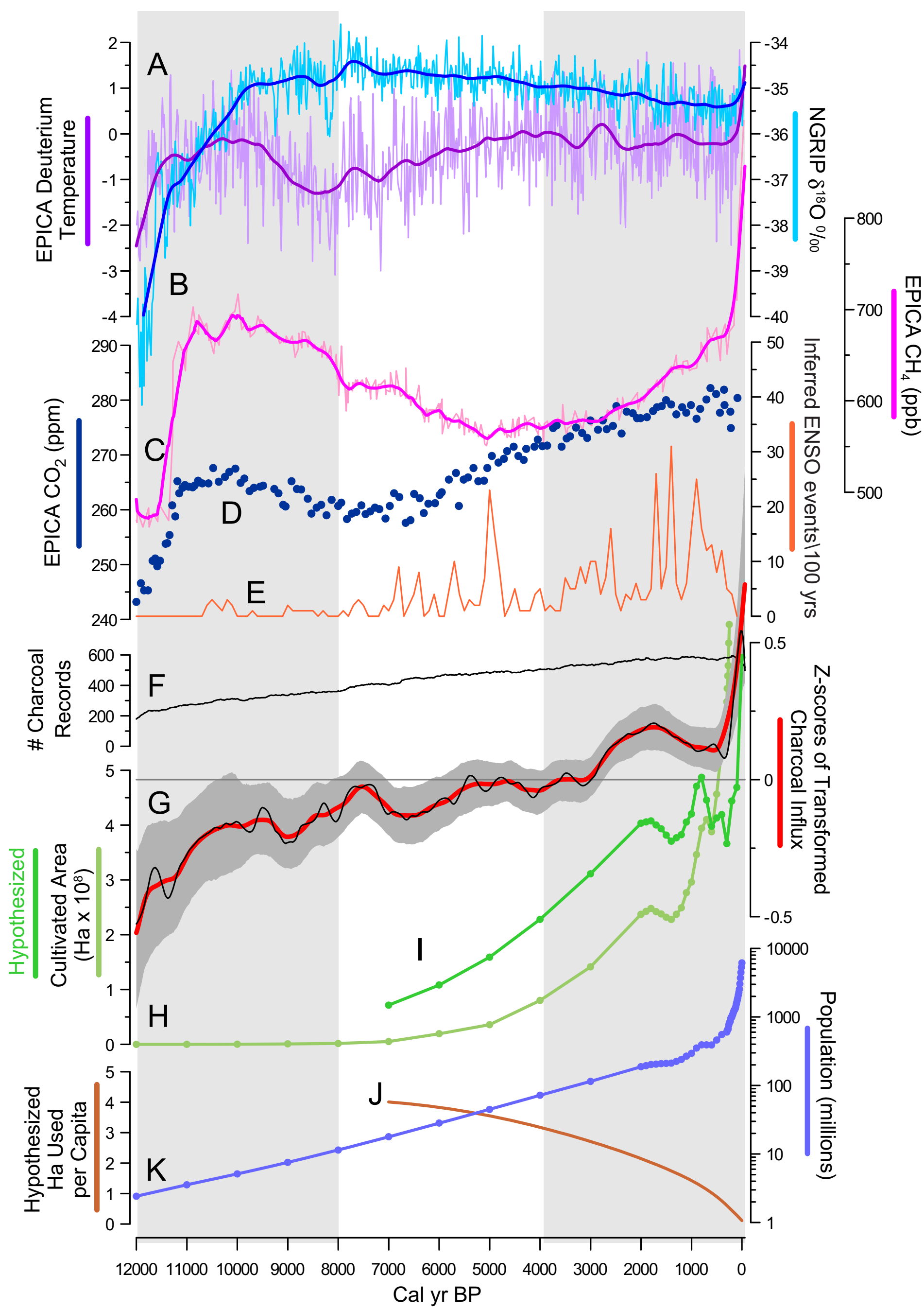

Fig. 8. Paleoclimate records including A) EPICA deuterium temperature record (Jouzel et al., 2007), B) NGRIP $\delta^{18}O$ record (Andersen et al., 2004), C) EPICA CO_2 record (Lüthi et al., 2008), D) EPICA CH_4 record (Loulergue et al., 2008), compared with E) ENSO activity from Laguna Pallcacocha in Ecuador (Moy et al., 2002); F) the number of charcoal records contributing to G) global biomass burning from sedimentary charcoal records smoothed with a 500-year (red) and 250-year (black) window with 95% bootstrap confidence intervals (gray shading), H) global total cultivated area based on HYDE (Klein Goldewijk et al., 2010) and I) Ruddiman and Ellis (2009), J) hypothesized changes in hectares of land used per person over time (after Ruddiman and Ellis, 2009), K) global population (Goldewijk et al., 2011).

2.5. Global summary

It has been suggested that forest clearance and changes in land-use, facilitated through extensive fire use, affected pre-industrial Holocene climate by altering atmospheric composition through the production and/or sequestration of greenhouse gases (Ruddiman, 2003; Ruddiman et al., 2008; Dull et al., 2010; Nevle et al., 2011). Demonstrating that fire use was integral to such changes requires that 1) humans produced large changes in biomass burning, and 2) those changes were large enough to affect atmospheric chemistry. We can examine whether the data are consistent with (1) by comparing trends in global biomass burning with those of climate, population and cultivated area during the Holocene.

Global biomass burning increases rapidly in the early Holocene and varies modestly in the mid-Holocene. In the late Holocene, global biomass burning is stable from 4 to 3 ka, increases from 3 to 2 ka, and then declines from 2 ka to the industrial era. In the industrial era fire increases sharply and then begins to decline after ca 1900 AD.

Global population increased exponentially in the Holocene as did cultivated area since 7 ka (Fig. 8i,k). Local maxima in population and cultivated area occurred ca 2 ka. The highest population concentrations and growth at this time were in Central America, Europe, China and southern Asia (Fig. 9). Population growth slowed

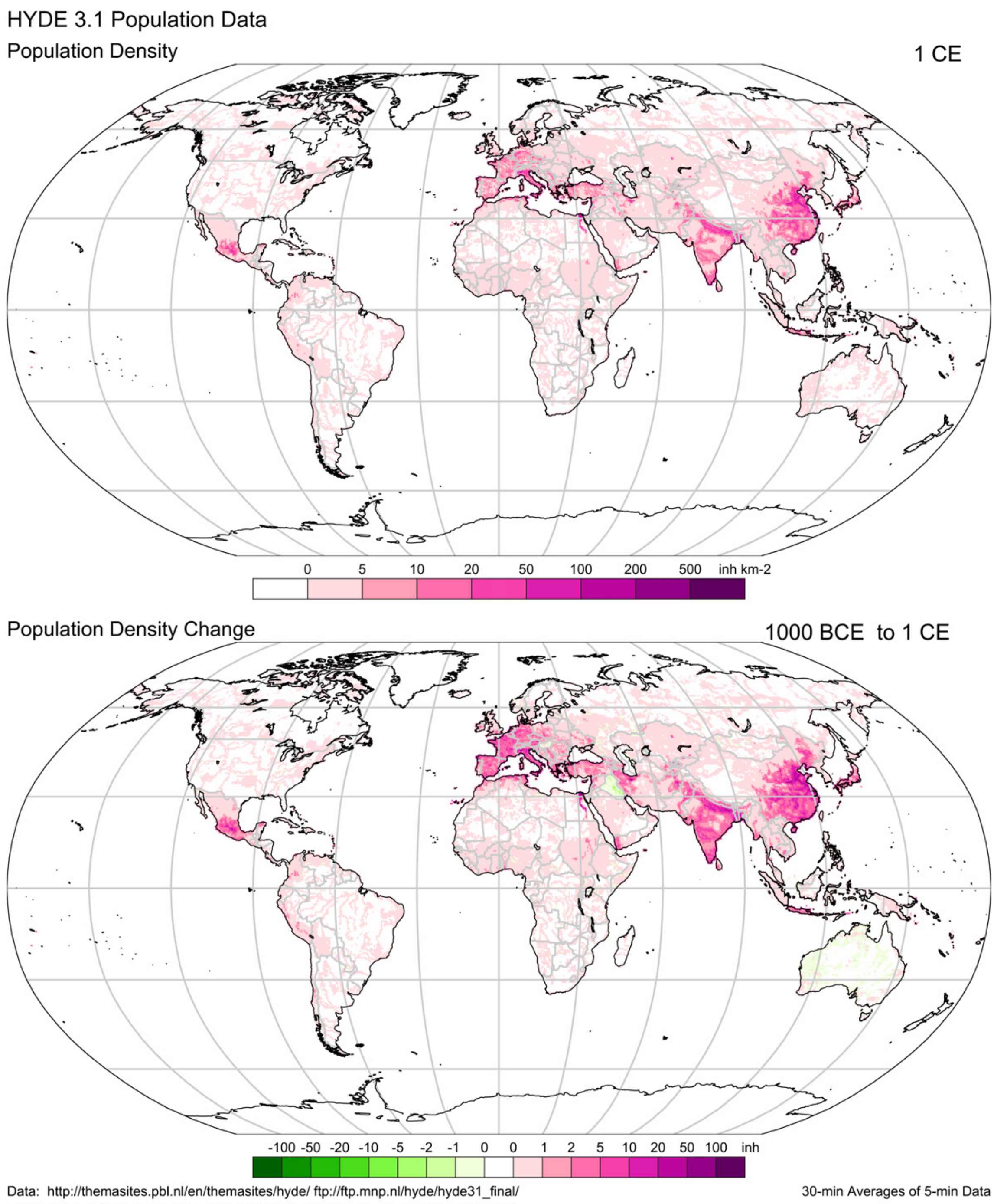

Fig. 9. Population estimates from the HYDE dataset for A) 2000 cal yr BP, and B) for the change from 3000 to 2000 cal yr BP (Klein Goldewijk et al., 2010).

for several centuries after 2 ka and declined at 800 years BP due to plagues in Eurasia, but then increased rapidly toward present, exceeding 6 billion in 2000 AD. Alternative estimates of cultivated area (Fig. 8j) based on the idea that more land was needed per person during the early stages of agriculture increase more rapidly than the original estimates from 7 to 2 ka (Fig. 8h); subsequent variations are also more pronounced and show local minima ca 1.5 and 0.5 ka and local maxima ca 1 ka. Both estimates, however, share several key features with global biomass burning, including the strong increase from 3 to 2 ka, a short-lived decline ca 0.5 ka and a dramatic increase in the industrial era.

Holocene temperature changes inferred from ice-core data (Fig. 8a,b) show rapid warming from 12 to 10 ka followed by local minima ca 8.3 ka and local maxima ca 7.7 ka (Andersen et al., 2004; Jouzel et al., 2007). Subsequently, temperatures steadily decline in the NGRIP record whereas they increase in the EPICA record during the mid-Holocene and vary modestly in the late Holocene. Changes in global biomass burning generally parallel changes in the NGRIP $\delta^{18}O$ and EPICA deuterium temperature records prior to about 7 ka, including a rapid increase in fire and temperatures from 12 to 10 ka, slightly varying levels between 10 and 8 ka, and local maxima ca 7.8 ka. However, it is important to note that the ice-core records strongly reflect high-latitude or polar conditions, and are not necessarily representative of mid-latitude trends (Daniau et al., 2012). The large increases in fire and temperatures during this interval are consistent with results from Daniau et al. (2012) indicating that temperature was the primary driver of biomass burning through the glacial–interglacial transition. The modest changes in fire during the mid-Holocene are more consistent with opposing northern and southern hemisphere trajectories in temperatures than with progressively increasing population and cultivated areas. After 3 ka, however, temperatures inferred from the ice-core records are stable, whereas biomass burning and cultivated area increase sharply to 2 ka, suggesting that human fire-use linked to the expansion of agriculture may have led to increased biomass burning in multiple regions at this time. The subsequent decline in fire globally from 2 to 0.5 ka, however, does not support the idea that the pre-industrial expansion of agriculture consistently led to widespread increases in biomass burning (Fig. 8f).

Regardless of what controlled changes in fire regimes during the Holocene, the trends in biomass burning and greenhouse gases (Fig. 8a,b,f) during the Holocene are not similar. An increase in biomass burning from 11 to 7 ka corresponds to a decrease in CO_2 and CH_4. The large shifts in biomass burning after 3 ka are not matched by similar shifts in greenhouse gases either. The lack of consistency between global biomass burning and trends in CO_2 and CH_4 are consistent with model experiments suggesting that factors other than fire, such as changes in peatlands or shallow water sedimentation of $CaCO_3$, are a more significant driver of centennial and millennial-scale variations in atmospheric composition (Kleinen et al., 2010; Pongratz et al., 2011; Stocker et al., 2011).

2.6. *From charcoal to fire history to emissions*

Patterns in charcoal records from different environments and integrated over various scales are often consistent with independent fire-history records (Whitlock and Millspaugh, 1996; Clark and Patterson, 1997; Tinner et al., 1998; Duffin et al., 2008; Higuera et al., 2011; Marlon et al., 2012), but factors relating to charcoal production, transportation, and deposition complicate reconstructions (Whitlock and Bartlein, 2004). It is unclear, for example, how changes in charcoal accumulation rates relate to burned area, fire intensity, and severity across a range of environments and vegetation types (Conedera and Tinner, 2010). As a result, calculating emissions from such data is particularly challenging. Current fire emissions estimates are based largely on satellite data and are calculated as the product of burned area and/or active fires, fuel loads (available biomass), the fraction of biomass combusted (combustion completeness), and emissions factors; these data are integrated over a given time and space (van der Werf et al., 2006). Charcoal data have been used to infer past emissions by reconstructing changes in vegetation and biomass burned over time by using a linear model to relate past estimates of biomass burning to carbon emissions from modern fires in different ecozones (Bremond et al., 2010). In some cases, changes in fire frequency at regional scales have been reconstructed and compared with biomass burning estimates to capture multiple aspects of fire-regime changes (Marlon et al., 2009) that may eventually support assessments of past fire emissions. Only mechanistic fire models evaluated by paleofire data, however, can elucidate the processes underlying the relationships between charcoal records and fire emissions (Bowman et al., 2009).

3. Conclusions

Globally fire was low at the beginning of the Holocene and increased during the Holocene. This trajectory is consistent with the global increase in temperature through the glacial–interglacial transition. However, there were many regions where the early Holocene was characterized by high fire, presumably because local conditions were conducive to this. Records near glaciated regions, including parts of Alaska, Canada, and northeastern Europe show this pattern, as do records from Patagonia, the Mediterranean, and monsoon Asia and Africa. Such patterns may reflect the influence of physical features (e.g., mountain ranges) and/or a sensitivity to atmospheric circulation patterns during the early Holocene (e.g., shifting jet streams (Carcaillet and Richard, 2000)), storm tracks (Whitlock et al., 2007), or monsoon-related shifts that have a strong effect on vegetation communities and fire.

It is clear that neither changes in human populations or changes in land use are sufficient to explain the biomass burning record at regional and continental scales during the Holocene. Our analysis does not exclude an important role for human activities on biomass burning at local scales or during specific intervals, but does provide evidence for a pervasive influence of climate across multiple spatial and temporal scales. Our results do not support the idea that human fire-use increased atmospheric CH_4 or CO_2 concentrations since the mid-Holocene. Early and mid-Holocene trends in fire are very different from those in CH_4 and CO_2, and late-Holocene variations in biomass burning are not matched by similar variations in CH_4 and CO_2.

There was a widespread increase in fire from 3 to 2 ka, everywhere except Australasia. There is evidence for the expansion of cultivated areas in Europe, Asia and the American tropics around ca 3–2 ka, and there is also evidence for increased warmth and drought in Europe, Asia and North America around the same time. However, the general increase in fire between 3 and 2 ka is puzzling. There is little evidence for a widespread climatic reorganization ca 3–2 ka that would lead to an increase in fire everywhere. Although an increase in e.g., ENSO frequency could potentially increase fire in both wet (via increasing drought) and dry (via increasing fuels) tropical ecosystems, a change in ENSO would not be expected to increase fire simultaneously in the tropics and extratropics. Likewise, it seems unlikely that human fire-use became globally synchronized 3000 years ago given the known socio-cultural diversity and human–environment interactions of that time. Improved understanding of this marked change in fire regimes across so much of the world will not be possible until we have a more comprehensive set of climate reconstructions as well as more reliable and detailed estimates of human population and land-use changes.

Finally, although fire exclusion is pervasive in the past century (Marlon et al., 2008), some regions (e.g., Europe) show a strong recent increase in fire that stands in sharp contrast to the widespread decline. Such spatial heterogeneity is expected given the complex combination of climate changes and human activities that affect fire regimes today.

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