THE GEOGRAPHY OF FIRE: A PALEO PERSEPCTIVE

by

JENNIFER R. MARLON

A DISSERTATION

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Confirmation of Approval and Acceptance of Dissertation prepared by:

Jennifer Marlon

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This dissertation has been accepted and approved in partial fulfillment of the requirements for the degree in the Department of Geography by:

Patrick Bartlein, Chairperson, Geography
Daniel Gavin, Member, Geography
W. Andrew Marcus, Member, Geography
Cathy Whitlock, Member, Geography
Ronald Mitchell, Outside Member, Political Science

and Richard Linton, Vice President for Research and Graduate Studies/Dean of the Graduate School for the University of Oregon.

September 5, 2009

Original approval signatures are on file with the Graduate School and the University of Oregon Libraries.

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Jennifer R. Marlon for the degree of Doctor of Philosophy

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Approved: ____
Dr. Patrick J. Bartlein

Fire is a fundamental, transformative, yet poorly understood process in the Earth system; it can radically reorganize ecosystems, alter regional carbon and energy balances, and change global climate. Short-term fire histories can be reconstructed from satellite (seasonal- to interannual-scales), historical (decadal scales), or dendrochronological records (for recent centuries), but only sedimentary charcoal records enable an analysis of the complex interactions between climate, vegetation and people that drive fire activity over longer temporal scales. This dissertation describes the compilation, synthesis and analysis of a global paleofire dataset and its application to understanding past, current, and future changes in fire activity.

Specifically, I co-led efforts to compile charcoal records around the world into a single database, and to conduct three meta-analyses to understand the controls on fire at multiple spatial and temporal scales. The first meta-analysis reconstructed global biomass burning since the Last Glacial Maximum (LGM) 21,000 years ago. Results

from this study demonstrated that global fire activity is low when conditions are cool and high when conditions are warm. This fundamental relationship between climate and fire is due in large part to associated changes in vegetation productivity. The second meta-analysis examined fire activity in North America during past abrupt climate changes and looked for evidence of continental-scale wildfires associated with a hypothesized comet impact ~13,000 years ago. This analysis found a correlation between increased fire activity and abrupt climate change, but provided no evidence for continental-scale wildfires. A final meta-analysis disentangled the climate and human influences on global biomass burning during the past 2000 years; it found a close relationship between climate change and biomass burning until ~1750 A.D., when human activities became a primary driver of global fire activity. Together, these three meta-analyses demonstrate that climate change is the primary control of global fire activity over long time scales. In general, global fire activity increases when the Earth's climate warms and decreases when climate cools. The paleofire data and analyses suggest that the rapid climate changes projected for coming decades will lead to widespread increases in fire frequency and biomass burning.

This dissertation includes previously published and unpublished co-authored material.

CURRICULUM VITAE

NAME OF AUTHOR: Jennifer Rachel Marlon

GRADUATE AND UNDERGRADUATE SCHOOLS ATTENDED:

University of Oregon
University at Albany, State University of New York

DEGREES AWARDED:

Doctor of Philosophy in Geography, 2009, University of Oregon
Master of Science in Geography, 2003, University of Oregon
Bachelor of Science, 1997, University at Albany, State University of New York

AREAS OF SPECIAL INTEREST:

Global Environmental Change
Fire History
Paleoecology

PROFESSIONAL EXPERIENCE:

Research Assistant, Department of Political Science, University of Oregon, 2005-2008

Instructor, Department of Geography, University of Oregon, 2006 and 2001-2002

Research Assistant, Department of Geography, University of Oregon, 2003-2004

Graduate Teaching Fellow, Department of Geography, University of Oregon, 2000-2001

GRANTS, AWARDS AND HONORS:

Best Ph.D. student paper presentation, Paleoenvironmental Change Specialty Group, Association of American Geographers, 2009

Doctoral Research Fellowship, World fire: a new long-term global synthesis and analysis, University of Oregon, 2008-2009

General University Scholarship, University of Oregon, 2008

Doctoral Dissertation Improvement Grant, Global fire since the Last Glacial Maximum, National Science Foundation, 2007-2008

National Science Foundation Travel Award, XVII INQUA Congress, Cairns, Australia, 2007

Summer Field Work Scholarship, Department of Geography, University of Oregon, 2007

General University Scholarship, University of Oregon, 2005

Best Ph.D. paper presentation (for Master's thesis), Biogeography Specialty Group, Association of American Geographers, 2004

General University Scholarship, University of Oregon, 2003

General University Scholarship, University of Oregon, 2002

PUBLICATIONS:

Marlon, J.R., Bartlein, P.J., Walsh, M.K., Harrison, S.P., Brown, K.J., Edwards, M.E., Higuera, P.E., Power, M.J., Anderson, R.S., Briles, C., Brunelle, A., Carcaillet, C., Daniels, M., Hu, F.S., Lavoie, M., Long, C., Minckley, T., Richard, P.J.H., Scott, A.C., Shafer, D.S., Tinner, W., Umbanhowar, C.E. Jr. Wildfire responses to abrupt climate change in North America (2009). Proceedings of the National Academy of Sciences 106(8), 2519-2524.

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This dissertation is dedicated to Lulu and Thea.

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CHAPTER I

INTRODUCTION

Fire is a fundamental, yet complex and poorly understood process in the Earth system. Fire can rapidly transform local landscapes and ecosystems, it can change carbon and energy balances at broad spatial scales, and it can impact global climate by altering atmospheric chemistry and physics. Today, concern is increasing about potential interactions between global warming and fire activity in particular (Bowman et al., 2009), but we have little knowledge of the basic processes that drive fire under rapidly changing climate conditions. Consequently, approaches for modeling fire in the Earth System or making broad-scale projections about the future are still in their initial stages. This research uses a synthetic approach to compile and analyze paleoenvironmental data to address the question "How and why does fire activity change on long time scales? Results from the research answer the question by identifying climate as the primary driving force behind changing patterns of broad-scale burning on decadal- to millennial-time scales. The findings of the dissertation improve our understanding of climate-fire-vegetation interactions under a range of climate conditions, thereby strengthening our ability to model fire in the Earth system. More accurate models should enable us to project potential climate-vegetation-fire interactions which can be used to improve decision-making about land management, conservation, climate change, and many other concerns.

Fire and Global Warming

Human-induced increases in CO_2 and the changes in temperature and moisture associated with the concomitant global climate change have already been linked to increasing fire activity in the United States (Westerling et al., 2006), Canada (Gillett et al., 2004) and Eurasia (Groisman et al., 2007; Balzter et al., 2007). Anthropogenic climate changes are also probably contributing to unusually extreme fires in Australia (Cary, 2002) southern Europe and South America (Cochrane and Barber, 2009). In the U.S. alone, extreme fires have become one of the most visible, costly and damaging impacts of climate change (Westerling et al., 2006). The average annual area burned has gone from ~3 million acres in 1985 to 1995, to ~5 million acres in 1995 to 2005 to over 8 million acres during the past five years (NIFC, Accessed 2009). Fire prevention and suppression costs have reached nearly $3 billion annually in recent years, with increases of 11% per year on average (Rasker, 2009), and costs are expected to increase (Flannigan et al., 2008; Gude et al., 2008). Climate change is not solely responsible for the recent changes in fire activity however. Patterns of burning result from complex interactions between climate, vegetation and people. Thus, understanding the causes of recent changes in fire activity is not straightforward. For example, there is ample evidence that Euro-American settlement, fuel build-up from fire suppression, and increased climatic variability may have all contributed to greater fire activity at particular places and times in the past (Swetnam and Betancourt, 1990; Veblen et al., 2000). The recent warming during the past few decades just adds more fuel to the fire, so to speak. Warmer climate conditions can lead to increasing fire activity in a variety of ways. Increases in spring and summer temperatures and earlier spring snowmelt, for example, have extended the length of the fire season and produced longer-lasting and more frequent fires (Westerling et al., 2006). Earlier spring snowmelt and longer fire seasons increase summer drought and give fuels more time to dry out. There is also concern that climate change may be increasing the occurrence of extreme fire weather events, thus contributing to the occurrence of unusually large, intense and severe burns (Moritz et al., 2004; Bowman et al., 2009; Moritz and Stephens, 2008). Finally, warmth-induced increases in vegetation productivity and thus fuel build-up can also promote fire. Not all regions or vegetation types are expected to experience more fire in the future, however. Areas with decreasing seasonal droughts and higher effective moisture may see reductions in fire activity (van der Werf et al., 2008c; Krawchuk et al., 2009). Shifts in vegetation that decrease connectivity, fine fuels, or flammability will reduce fire activity (Higuera et al., 2009). Increasing warmth in the absence of sufficient moisture for plant growth can also reduce fuels to the point where levels are insufficient to support fire spread.

Understanding how and why fire activity is changing requires data about past fires. Fire-history can be reconstructed from satellite (seasonal- to interannual-scales), historical (decadal scales), or dendrochronological records (for recent centuries), but only sedimentary charcoal records enable an analysis of the complex interactions between climate, vegetation and people that drive fire activity over longer temporal scales. This dissertation is a compilation, synthesis and analysis of the first global dataset of paleofire records. The results provide critical long-term context for current and future changes in fire activity.

Outline of the Dissertation

This thesis presents a new global charcoal-based fire-history dataset and three meta-analyses of the charcoal records. The data come from six continents and are aggregated at regional-to-global scales, providing information about the spatial and temporal distribution of fire around the world during the past 21,000 years.

Three meta-analyses examine different subsets of the large dataset and compare these subsets with other paleoenvironmental and paleoclimatic data to explain the variations in fire. The analyses illuminate the importance of climate in particular as a driving force of broad-scale patterns of fire activity. The effects of gradual climate changes on fire that occur over millennia are as apparent as those from abrupt climate changes. Comparisons of the fire data with trends in population and land-use data help explain more recent patterns of burning. These latter analyses allowed us to identify a surprising and dramatically rapid decrease in fire activity during the past century due to human activities including the expansion of agriculture, grazing, fire suppression and landscape fragmentation.

In addition to advancing our understanding of global biomass burning, the research process produced important insights into the challenges and benefits of creating and analyzing a large dataset. Because this was the first synthesis of paleofire data performed at the global scale, we spent a good deal of effort comparing and testing alternative methods for combining records from a wide variety of environments and based on diverse methodologies. Some of the findings from this process are discussed in the conclusions.

Below I provide a general background on the importance of fire in the Earth system. I then describe the range of techniques currently in use for reconstructing past fires on multiple temporal and spatial scales and for simulating future fire activity at global scales. The motivations for conducting fire-history research at the global scale are presented next, followed by a description of the three specific questions that guide the research. Chapters II, III, and IV – the main chapters – follow this introductory material. In Chapter V, I discuss future research needs and identify major lessons learned from the three studies, both in terms of scientific content and process.

Fire in the Earth System

Fire is a pervasive and persistent process in terrestrial ecosystems. As one of Earth's primary transformative process, it is intricately woven into complex relationships with both physical and biological systems. Following is a very brief overview of some of the primary and often opposite ways in which fire interacts with such systems – as a force of growth and destruction, as well as of species selection and composition, in ecosystems, as a cause and effect of climate changes, and as a tool and hazard to people. In this review, the term "fire activity" refers generally to all aspects of fire on the landscape, including biomass burned, fire frequency, fire intensity and area burned. "Fire regime" refers to the seasonality, frequency, intensity, and size of fires that are characteristic in a given area or ecosystem.

Fire as an Ecosystem Disturbance

Wildfire is viewed from a distance by most of us, if it is viewed at all. But its effects on the local environment – on people, vegetation, other organisms, soil, nutrient cycling, and hydrology – are far-reaching. The impacts of fire on ecosystem services, including soil fertility, grazing value, carbon sequestration, biodiversity and tourism, can be subtle but they are equally important (Lavorel et al., 2007). As the most widespread form of disturbance in terrestrial ecosystems, fire affects 200 to 500 million hectares annually worldwide– an area about the size of India (Lavorel et al., 2007). The relative importance of fire across these ecosystems varies greatly, as does the function or type of fires that tend to occur (Goldammer, 1990; Agee, 1993). Patterns of fire across time and space are determined primarily by climate, vegetation and land-use (Dwyer et al., 2000). Chuvieco et al. (2008) used metrics derived from satellite data to characterize the density, seasonal duration and inter-annual variability of active fires around the world. With just six years of recent data, they found that fires were a frequent occurrence on more than 30% of the land surface today.

The role of fire as a disturbance process has been relatively well-studied in particular geographic regions such as in the tropics (Goldammer, 1990; van der Werf et al., 2008c), in Mediterranean-type ecosystems (Moreno and Oechel, 1994), in forests of the Pacific Northwest (Agee, 1993), and in Australia (Bradstock et al., 2002). Other areas, such as riparian ecosystems, have received relatively little attention however (Dwire and Kauffman, 2003). Ecosystems are often associated with particular fire regimes, which can be categorized into three general types: 1) frequent, low-severity surface fires that are non-lethal to most plants; 2) infrequent, stand-replacing (high-severity) crown fires; and 3) mixed-severity regimes that experience both low and high-severity fires. The fire regime affects a wide variety of ecosystem characteristics including vegetation structure (Bond et al., 2005), community composition (Foster and Zebryk, 1993) and, by extension, habitat availability. Fire also affects many fundamental ecosystem processes such as nutrient cycling (Turner et al., 2007), vegetation succession pathways (Turner et al., 1997; Brown and Smith, 2000), other disturbance patterns such as pest outbreaks (McCullough et al., 1998), hydrology, and erosion (DeBano et al., 1998; Shlisky et al., 2007). Some of these processes are critical to species that inhabit the ecosystem. For example, fires maintain grasslands that would otherwise convert to shrub or woodlands, and thus provide habitat for grazers. Fires also provide nutrient flushes that stimulate plant growth and regeneration opportunities for shade-intolerant plants.

Over time, species have evolved a wide range of traits such as thick, protective bark or the ability to resprout that make them adapted to or even dependent on fire (Bond and van Wilgen, 1996). Lodgepole pine (*Pinus contorta*) is a common example of a fire-dependent species; its serotinous cones open only after fire melts the resin in the cones, which allows the scales to dry out and eventually curl open to release the seeds. The seeds in turn are black or speckled, which provides camouflage on the sooty ground after a fire (Peluso, 2007). Eucalyptus communities, which evolved in Australia with frequent fire are well-known as fire-dependent communities (Gill, 1981). Eucalyptus trees have many fire-related adaptations, including flammable oils in the foliage, highly flammable bark, deciduous bark streamers, lichen epiphytes, sprouting, and heat-resistant seed capsules. In California, where the species was introduced, Eucalyptus are sometimes referred to as “gasoline trees” by firefighters because of the high flammability of their crowns, which are open and have hanging branches that encourage strong updrafts (Brown and Smith, 2000). Many other plant and animal species depend on changes in vegetation, sunlight and nutrients that accompany fire. Kirtland's warbler (*Dendroica kirtlandii*), for example, is a rare (now endangered) bird in the Great Lakes region that prefers young, fire-dependent jack pine (*Pinus banksiana*) stands for nesting (Probst and Donnerwright, 2003). The black fire beetle (*Melanophila acuminata*) has special infrared antennae receptors that guide them to the heat generated during fires (Evans, 2005), and certain mushrooms (e.g. morels, *Morchella* species) fruit prolifically after fire (Pilz et al., 2004). Consequently, increases (or decreases) in fire activity can trigger community reorganizations and provide new habitat or regeneration opportunities for fire-dependent (or fire-sensitive) taxa (Brown and Smith, 2000).

Today, fire-regime changes may promote the invasion of non-native species, and add stress to already threatened or endangered species (McKenzie et al., 2004; Kodandapani et al., 2004). As a result, fire management is a critical yet complicated aspect of land and resource management. Expanding development at the wildland-urban interface coupled with increasing "megafires" – unusually large, intense and severe fires – are making fire management even more difficult (Arno and Fiedler, 2004).

Fire, the Carbon Cycle and Climate

Although fire is a local process, it has global consequences because emissions from burning vegetation (biomass burning) are carried aloft and mixed throughout the atmosphere. Emissions of gases and particulates affect climate through biophysical (e.g. albedo) (Randerson et al., 2006; McConnell et al., 2007) and biogeochemical (e.g. the carbon cycle) processes (Kasischke et al., 1995; Field and Raupach, 2004; Schultz et al., 2008) that alter the Earth's energy balance, and through impacts on the hydrologic cycle (Andreae et al., 2004). Globally, biomass burning contributes about 30% of carbon dioxide (CO_2), 32% of carbon monoxide (CO), 38% of tropospheric ozone (O_3), 7% of total particular matter and 39% of particulate organic carbon (Bowman et al., 2009). Smoke from fires in the Amazon increases the elevation at which clouds produce precipitation, and may affect global circulation patterns (Andreae et al., 2004). Annually, fires release about 2.5 petagrams (2.5 billion metric tons) of carbon per year (Pg C $year^{-1}$), but this amount can drastically increase during severe drought years. During the 1997-1998 El Nino, for example, burning contributed to an estimated 0.81 – 2.57 Pg C released in Indonesia alone – about 40% of all carbon emitted globally that year – and representing the largest increase in CO_2 since 1957 (Page et al., 2002). A trend towards increasing fires in this area and in the tropical Americas is largely caused by human-induced "deforestation fires" (Morton et al., 2008; Field et al., 2009), but climate-induced increases in droughts are expected to exacerbate the problem (van der Werf et al., 2008a; Cochrane and Barber, 2009). Because vegetation regrowth typically absorbs any carbon that was released during previous fires, however, periodic emissions from biomass burning are assumed to have no net effect on the carbon budget. However, increases or decreases in fire emissions that occur in conjunction with long-term changes in climate, vegegation and soil carbon, for example, may have longer-lasting effects on the carbon budget.

The emissions of aerosols and particulates from fire, including black carbon (soot) have direct and indirect effects on climate. Direct effects on radiative forcing occur through the scattering of solar radiation and the absorption/emission of terrestrial radiation. Indirect effects on radiative forcing occur primarily because aerosols affect cloud properties (Sherwood, 2002; Guyon et al., 2005). Because they are not mixed as thoroughly as gases in the atmosphere, the distribution of aerosols spatially and through the vertical profile of the atmosphere matters. As a result, the effects of aerosols on climate are difficult to model and thus remain one of the largest uncertainties in the simulation of global climate. Soot in particular plays an important role in the climate system because it can decrease the albedo (reflectivity) of snow, ice and land and thus increase surface temperatures (Hansen and Nazarenko, 2004). This impact is particularly of concern in the Arctic, where small changes to the amount and reflectivity of abundant snow and ice can have major impacts on surface temperatures due to positive feedbacks. Koch and Hansen (2005) used a global climate model to show that soot in the Arctic today comes from both industrial combustion and biomass burning, and that the primary source area for black carbon aerosols in the Northern Hemisphere is mainly from industrial pollution in Southeast Asia. Additional research with models indicates that soot has substantially contributed to rapid Arctic warming during the past three decades (Shindell and Faluvegi, 2009), and is responsible for 50 percent, or almost 1°C of the total 1.9°C increased Arctic warming from 1890 to 2007 because of its effects on snow and ice albedo at high latitudes (Ramanathan and Carmichael, 2008). Soot may also affect regional climate variability by absorbing radiation and heating the air, which can alter regional atmospheric stability and vertical motions, and thus affect the large-scale circulation and hydrologic cycle (Menon et al., 2002). Fossil-fuel combustion is the primary source of soot today, but biomass burning has been the primary contributor in the past, and depending on the location, magnitude and direction of future changes, may play an increasingly important role in the future.

Variations in fire emissions can be caused by changes in area burned, vegetation type burned (e.g. a shift from white spruce to black spruce can increase fires Lynch, 2003) or fire intensity. Thus, the changing geographic distribution of fire regimes will have important effects on global climate and atmospheric chemistry, yet the nature of these impacts and the processes by which they will occur are largely unknown. Recent research on the boreal forest suggests that increases in area burned will account for the vast majority of increases in CO_2 in Canada (Amiro et al., 2009), but community composition, vegetation density, and related factors are also clearly important. In addition, increased emissions that contribute to warming may be augmented by changes in surface albedo due to the deposition of black carbon and by post-fire vegetation changes in the boreal forest (Randerson et al., 2006).

Fire and People

People have used fire for millennia and for myriad reasons; from cooking, hunting and crop management to felling trees, collecting insects, warfare and communication (Williams, 2003; Pyne, 1995; Pyne, 2001). Fire has also provided many less-obvious yet crucial services by supporting the ecosystems that people depend on for resources (FAO, 2007). For instance, fire naturally clears and thins weak and dying vegetation and debris. This “cleaning” process is essential for the maintenance of healthy ecosystems that people depend on for fuel and food, tourism, recreation and other activities. Without periodic fire, vegetation will build up and can eventually lead to unusually severe and damaging fires. Fire also kills pests and diseases that are harmful to both plants and animals, and that kill far more trees each year than fire itself does (McCullough et al., 1998; Logan et al., 2003). By maintaining local habitat and species diversity, fire helps ensure that ecosystems are resilient to other disturbances like floods, severe weather, and insect outbreaks.

Paleoecological data suggest that aboriginal people have used fire as a tool to aid survival in the Australian bush (i.e., "fire-stick farming" Bliege Bird et al., 2008) for tens of thousands of years (Singh et al., 1981; Clark, 1983; Turney et al., 2001). There is ample paleoecological and historical evidence for intentional landscape burning by people in Europe (Tinner et al., 1999; Williams, 2006), Asia (Zong et al., 2007; Jiang et al., 2008; Chang Huang et al., 2006), the Americas (Clark and Royall, 1995b; Mann, 2005) and elsewhere during the Holocene (i.e. the past 11,000 years). Fire has been a tool, weapon and hazard to people in Africa for millennia as well, although documentation and data for such practices and processes are not abundant (but see Burney, 1987; Darbyshire et al., 2003).

The spatial extent to which human uses of fire have determined fire regimes and vegetation patterns in the past and at broad spatial scales is a matter of debate (Schule, 1990; Heusser, 1994; Boyd, 1999; Turney et al., 2001; Vale, 2002; Stewart et al., 2002; Wuerthner, 2006). As a U.S. Forest Service historical analyst recently noted, the intentional use of fire by native people "has been easy to document but difficult to substantiate" (Williams, 2003, p. 1). It is clear however, that humans have radically altered fire regimes *globally* in recent centuries (Pyne, 1995) and particularly since ~1750 A.D. (Marlon et al., 2008). The directions and pathways of these changes varied greatly from place to place, but many locations experienced a similar general pattern of fire activity due to European colonization (e.g. of Africa, Asia, the Americas and Australia). This pattern typically consisted of a radical shift from fire as an integral ecological process that evolved slowly on the landscape due to both natural and cultural forces, to a process that was heavily controlled and regulated or completely eliminated (Pyne, 1995). In some cases (e.g., America), the shift itself was accompanied by a relatively brief but massive increase in fire related to cultural clashes and warfare, logging, land-clearance, railroad building, and similar activities associated with colonialism (Pyne, 1982).

The widespread transformation of fire regimes by people in recent centuries has had dramatic consequences not just for vegetation communities (e.g. Cochrane, 2003), biodiversity (TNC, 2004), ecosystem services, and now the climate system, but also for human health and well-being. Some argue that climate change coupled with decades of active fire suppression in industrialized nations has led to the advent of "megafires" around the globe. Regardless of their cause, such conflagrations have immediate consequences on air quality and visibility, leading to respiratory problems and other public health concerns (Lohman et al., 2007).

Clearly, there is a critical need not only to better understand the physical, biological and cultural aspects of fire history, but to use and integrate this information into modern fire policy and management. Our challenge today is to find a proper balance between fighting and igniting fires (Pyne, 2004). In America, fire experts argue that our cultural aversion toward fire has led us to see it as something to be feared, controlled and suppressed (Wuerthner, 2006; Bowman et al., 2009), and that this attitude has prevented us from appreciating the great importance of fire in ecosystems and the critical services that it provides to people (Arno and Fiedler, 2004; Pyne, 2004).

Fire-history research increases our awareness and understanding of the biological importance of fire. It also provides essential long-term context for current changes in fire regimes by illuminating their nature, causes and consequences in the past. Critically, fire-history studies can provide concrete evidence of past fire dynamics that can be used to support increased public awareness and political support for restoring fire to ecosystems that need it, protecting environments where it is not needed or wanted, and helping us form a new positive relationship towards fire in an uncertain future.

Reconstructing Fire History

Fire history encompasses data and methodological approaches that vary across spatial scales from forest stands to the globe, and across temporal scales from days to millennia (Gavin et al., 2007). Each approach has its own strengths and weaknesses that tend to reflect trade-offs between spatial and temporal resolution, or between the quality or richness of fire-history information in the dataset and its spatial or temporal coverage.

Observations of Fire on Interannual to Century Time Scales

Much effort has been directed towards the observation of recent fires and their emissions, in particular. Such analyses may rely on satellite, field and/or experimental data. These modern datasets are primarily being used to 1) determine trace gas and aerosol emissions from fires (Guyon et al., 2005) including black carbon and soot (Koch and Hansen, 2005) and how such changes feed back to the climate system (Hansen and Nazarenko, 2004), and 2) to analyze the spatial and temporal distribution of fires to understand climate-fire-vegetation links at broad spatial scales (Le Page et al., 2008; Krawchuk et al., 2009).

Satellite datasets are unique because of their vast spatial coverage (Mota et al., 2006; van der Werf et al., 2006; van der Werf et al., 2008b). However, these datasets have very limited temporal coverage – they span several years or a few decades at most. In addition, there are often significant discrepancies in estimates of fire incidence and burnt area across products (Spessa et al., 2003). Nevertheless, studies based on satellite data have promoted major advances in fire science, and are critical for validating global models of fire activity (discussed in more detail below). Fire-history data from satellites have been used to examine the terrestrial carbon cycle, for example, including the uptake and release of carbon through vegetation burning and re-growth (van der Werf et al., 2006; Chuvieco et al., 2008), and the impacts of wildfires on atmospheric methane (CH_4) concentrations (van der Werf et al., 2004). The varied effects of climate on biomass burning, however, and of fire feedbacks to atmospheric chemistry and global climate, remain poorly understand and thus increase uncertainty in future projections of climate change.

Historical and dendrochronological records (i.e. fire-scar and stand-age data) are used to examine fire-related events and processes on time scales of days to centuries and on spatial scales from forest stands to the globe (Hardy et al., 2001; Heyerdahl et al., 2001; Mouillot and Field, 2005). Historical records generally span decades, whereas dendrochronological data can span seasons to centuries. The first global analysis of such data was compiled by Mouillot and Field (2005); it combined historical, dendrochronological and remotely-sensed data to estimate area burned during the 20th century (Mouillot and Field, 2005). The synthesis shows that from 1900 to 2000 burning generally increased in the tropics and decreased in the extra-tropics. At finer scales, historical and dendrochronological fire-history data have been used to guide resource management (e.g. Schoennagel et al., 2004)) and to understand climatic and human influences on fire regimes (Veblen et al., 1999; Veblen et al., 2000; Alaback, 2003). Substantial research efforts have been directed towards improving our knowledge of climatic controls on fire in particular (e.g. Romme and Despain, 1989; Grissino-Mayer and Swetnam, 2000; Girardin and Sauchyn, 2008), with emphasis on the climatic forcing of regionally synchronous fires on decadal- to centennial-time scales (Grissino-Mayer, 2004; Gedalof et al., 2005; Heyerdahl et al., 2008), and on periods with below-average fire activity (Morgan et al., 2008).

Paleofire Records

To go farther back in time, fire activity can be reconstructed from charcoal abundances in sediments from lakes, peat bogs, soils and other environments. When fires occur on the landscape, charcoal particles of various sizes, along with pollen grains and other organic and inorganic material are blown and washed into nearby lakes, bogs, etc.

that accumulate the material through time. Over time, older deposits are buried by younger layers, creating a sedimentary archive that reflects changing environmental conditions through millennia. If mixing and movement within the soil or sediment is limited, an annual- to decadal-scale sequence of changes in fire activity and vegetation can be reconstructed from charcoal and pollen abundances in the sediments. Field and laboratory methods for reconstructing paleofires from charcoal evolved from century-old techniques for reconstructing vegetation histories from pollen data. Several methods exist for quantifying and analyzing charcoal data that have been comprehensively detailed elsewhere (Whitlock and Larsen, 2002; Conedera et al., 2009) and so are not covered here.

Paleofire reconstructions generally span centuries to millennia. Because they integrate charcoal particles transported by wind and water, the source area for charcoal records is not well-defined (Clark and Patterson, 1997). Experimental and theoretical studies (Clark, 1988; Lynch et al., 2004a; Higuera et al., 2007) indicate that smaller particles travel longer distances than larger particles, so researchers tally different size classes of charcoal particles and use variations in macroscopic (generally >100μm in diameter) particle abundances to reconstruct local fire activity (within a few kilometers of the sites) and microscopic particles (<100μm in diameter) to reconstruct more regional fire activity (well beyond a few kilometers) (Clark and Royall, 1995a; Whitlock and Bartlein, 2004). Studies of macroscopic charcoal data typically employ a high-resolution sampling approach where a sediment core is sampled at continuous centimeter (or finer) intervals throughout the core. Microscopic charcoal is usually quantified on pollen slides, which are often sampled at ~10-centimeter intervals. As a result, microscopic charcoal records reflect regional burning trends, but not individual fire events. Macroscopic charcoal records can provide high-resolution datasets where individual charcoal peaks reflect individual fire events or clusters of events (fire episodes). Such records can be "calibrated" by comparing the date of recent peaks in charcoal abundances with dates of recent fires obtained from independent lines of evidence, such as historical documents, fire-scarred trees or stand ages (Clark, 1990; Tinner et al., 1998; Allen et al., 2008). In well-calibrated records, trends in fire frequency can be inferred by identifying peaks in charcoal above a certain threshold (Long et al., 1998). Changes in the level of biomass burning can also be obtained by smoothing the highly variable charcoal data to emphasize general trends (Marlon et al., 2006).

Paleofire research began a few decades ago and has been growing rapidly due to a greater awareness of the important and varied roles that fire plays in the Earth system and in human-environment interactions. Indeed, human impacts on fire regimes were often of primary interest in the earliest paleofire studies, which identified past fires from aboriginal burning (Clark, 1983; Turney et al., 2001), land clearance during island colonization (Burney, 1987), Native American burning (Clark and Royall, 1995b; Bush et al., 2008), cultural changes (Tinner et al., 1999) and European settlement (Clark et al., 1996). Climate-fire interactions were also an early interest (Swain, 1973; Swain, 1978; Clark, 1990) and have become even more so in recent years (Millspaugh et al., 2000; Black and Mooney, 2006; Figueroa-Rangel et al., 2008), as have vegetation-fire interactions (Lynch, 2003; Higuera et al., 2009) and the effects of local environmental characteristics on fire regimes (Gavin et al., 2003; Gavin et al., 2006).

Paleofires can also be reconstructed from sub-micron (aerosol) carbon particles (i.e., black carbon) in sediments. These records integrate particles from large source areas and may reflect environmental changes at broad or even hemispheric scales. Records of black carbon from marine sediments have been used to reconstruct broad regional changes in biomass burning (Bird and Cali, 1998), and black carbon in ice cores have been linked to hemispheric changes in industrial pollution and biomass burning (McConnell et al., 2007). Techniques for extracting such data are more complicated and problematic than for larger charcoal particles (Conedera et al., 2009), so far fewer records exist. Nevertheless, black carbon records hold significant potential for providing continuous records of fire history on orbital time scales. Wang et al. (2005), for example, used black carbon from sediments in the Loess Plateau region of China to reconstruct paleofires through the past glacial/interglacial cycle. They found that black carbon was more abundant during glacial periods and in the northern areas of the Plateau, which they attributed to drier climate conditions and more frequent burning of grasslands. Similar results for glacial versus interglacial paleofire activity were obtained from a black carbon record from the Banda Sea near Indonesia (van der Kaars et al., 2000). A long black carbon record from Africa, however, indicates that the highest levels of burning occurred during the transition from the glacial to interglacial period, suggesting that climate *variability* played an important role there (Bird and Cali, 2002). The variable source areas of black carbon, a lack of validation, and difficulties with data analysis all limit the use of individual black carbon records as proxies of broad-scale burning. When combined with networks of local charcoal-based records, however, such data may produce reliable and robust regional to global-scale reconstructions.

Earth's climate during the past few million years has been characterized by oscillations between cold glacial and relatively warm interglacial periods. Several paleofire records, particularly from marine sediments and loess deposits, have demonstrated that changes in fire have tracked many of these large climatic shifts (Daniau et al., 2007; Daniau et al., 2009). The proposed mechanisms for such a link have generally been through the direct effects of changing climate on fire activity (Beaufort et al., 2003; Thevenon et al., 2003) or via the influence of climate on vegetation and thus biomass accumulation (Wang et al., 2005; Daniau et al., 2007). The impacts of ice-sheet dynamics on atmospheric circulation and regional climates and hence fire have also been invoked (Verardo and Ruddiman, 1996) as have sea level increases that reduced the land area available to support vegetation and fires (Luo et al., 2001). Several long fire-history records have attributed changes in black carbon or charcoal abundances to human colonization of southeast Asia and Australia (Wang et al., 1999; Kershaw et al., 2002; Beaufort et al., 2003; Thevenon et al., 2003), but climate, vegetation and other environmental changes have not been ruled out in these cases.

In the more distant past, fossil charcoal buried in ancient sediments has provided evidence that fires have burned on Earth for millions of years (Scott and Glasspool, 2006). The oldest charcoal layers date to the late Devonian, about 350 million years ago (mya). Although lightning has always occurred, adequate amounts of terrestrial vegetation were probably not present until at least the mid-late Devonian. Sufficient levels of atmospheric oxygen were also a prerequisite for combustion and thus fires to commonly occur (Scott and Glasspool, 2006). Fluctuations in fire activity levels through following geologic epochs were primarily a function of variations in atmospheric oxygen, vegetation characteristics, and climate (Scott and Glasspool, 2006).

Despite the complicated mix of factors that control fire regimes on multiple temporal and spatial scales, aggregations of individual paleofire records across very large areas (i.e., hundreds or even thousands of kilometers) have demonstrated that very distant locations can share certain fire-history characteristics on centennial and greater time scales (Haberle and Ledru, 2001; Carcaillet et al., 2002; Whitlock et al., 2007; Marlon et al., 2008; Power et al., 2008; Whitlock et al., 2008; Marlon et al., 2009). Such similarities implicate the importance of changing climatic boundary conditions, such as incoming solar radiation (insolation) (which changes with variations in the Earth's orbital parameters and solar output), atmospheric composition (in particular the concentrations of long-lived greenhouse gases), aerosols (from volcanic eruptions and terrestrial dust sources), and the impact these changing boundary conditions have on ice sheets and sea-surface temperatures that in turn govern regional paleoclimatic variations. This is not to say that all records will show the same variations, or that nearby records will have similar features or trends. Patterns of biomass burning at local scales and at annual- to centennial-time scales are naturally heterogeneous and variable. Indeed, the paleoecological literature is largely composed of studies that detail the importance to fire regimes of local controls such as people (Burney and Burney, 2003; Parshall et al., 2003; Dull, 2007), topography (Gavin et al., 2003), geology (Briles et al., 2008), hydrology (Campbell and Campbell, 2000) and vegetation (Colombaroli et al., 2008; Higuera et al., 2008), or some combination of these factors (Lynch et al., 2004b; Gavin et al., 2006; Umbanhowar Jr et al., 2006). This may be a result (in part) of the lack of local climate data available for comparison with paleoecological data rather than a reflection of the relative unimportance of weather and climate on fire regimes. Nevertheless, increasing attention is being directed towards patterns of burning at a range of spatial scales (Lynch et al., 2007; Whitlock et al., 2007; Power et al., 2008), and the focus of this thesis is generally on processes operating at regional to global scales.

Projecting Future Fire Activity

Computer modeling has been used extensively to simulate fire behavior at the landscape scale (e.g. Bachelet et al., 2000; Keane et al., 2004; Andrews et al., 2008), but less research has focused on regional to global-scales due to the heterogeneous and complex nature of the interactions between fire, fuels, people and climate at those scales. Nevertheless, modelers recognize the important impacts that fire has on ecosystem composition, the carbon cycle, and climate and are working towards simulating these effects (Fosberg et al., 1999).

In some cases, process-based fire models are embedded within dynamic global vegetation models, which are in turn driven by climate data (Lenihan et al., 1998; Bachelet et al., 2001; Thonicke et al., 2001; Williams et al., 2001; Venevsky et al., 2002; Bond et al., 2005). More often, however, fire *risk* is simulated based on statistical models and climate data (Flannigan et al., 2000; Flannigan et al., 2005; Giannakopoulos et al., 2005; Scholze et al., 2006). Alternatively, modelers interested primarily in atmospheric chemistry may simulate only fire emissions and their associated impacts on greenhouse gas concentrations and radiative forcing (Randerson et al., 2006; Schultz et al., 2008).

Three noteworthy studies recently used models to answer "What if?" questions that demonstrate the key role that fire plays in ecosystem dynamics, and that highlight the magnitude of possible impacts that broad-scale changes in fire regimes could initiate. In the first study, Bond et al. (2005) use a dynamic global vegetation model to simulate a world without fire, and thereby provide evidence that fire rather than climate is the key determinant of the distribution of grasslands and savannas (Bond et al., 2005). The authors argue that in the absence of the frequent burning they experience today, closed forests would double in area given current climate conditions. The second study by Scholze et al. (2006) examined how fire risk, defined in terms of specific changes in particular climatic variables, would change with a doubling or quadrupling of atmospheric CO_2. This study found substantial increases in fire risk in widespread parts of the world, including Amazonia, the far north, and many semiarid regions (Scholze et al., 2006). In a third study, Krawchuk et al. (2009) combined statistical models with output from a General Circulation Model to simulate the global distribution of future probabilities of fire occurrence. The results suggest that future warming may produce a net balance between regions with increased versus decreased fire over the next century. The authors note that such a balance does not imply insignificant ecological or social impacts, however. Also, the model does not include potential impacts from vegetation feedbacks to the climate system, or consider potential impacts from increased CO_2 (Balshi et al., 2007) or mortality (Mantgem et al., 2009) on vegetation productivity (dieback), which may have important effects on future fire activity.

The creation of remotely-sensed global burned-area datasets have spurred model development because such products can be compared with simulations of recent fire activity (past ~20 years) to serve as a first-order test of model accuracy. The magnitude and speed of projected climate changes over the next century, however, are likely to drive shifts in climate and vegetation well outside the range of variability observed during the past few decades. Thus, simulations of paleofires and comparisons with empirical paleofire data for time periods when climate was very different from today would serve as a strong test of model validity. Any realistic simulation of regional or global fire activity that aims to make projections about ecosystems or the carbon cycle must include robust data on vegetation (fuel) characteristics and human activity (ignitions), both of which substantially influence fire regimes (Veblen et al., 2000; Archibald et al., 2009), as well as climate.

Motivation for a Global Synthesis

The vast majority of paleofire studies (and paleoecological studies in general) are, by necessity, performed at the local scale. This is the case because data collection and analysis for a single site is highly labor intensive. Consequently, studies are designed to reconstruct detailed environmental histories of particular places that are deep in time and limited in space. Methodological differences between paleofire studies, which can be substantial, can further complicate comparative analyses. Thus, despite the development of hundreds of charcoal-based paleofire records during the past few decades, meta-analyses have not been widely attempted.

Syntheses of pollen data are more common, due in part to the longer history of palynology. As recently described by Gajewski (2008) in a review of research employing the Global Pollen Database, pollen syntheses have been used to 1) generate broad-scale maps; 2) test hypotheses; and 3) explore (mine) data. Meta-analyses on paleofire data could be categorized in a similar fashion, and are described in detail later. Although pollen data has been collected for over a hundred years now, the Global Pollen Database was formally released only in 2006 (although much of these data were available since the 1990s), and pollen-based meta-analyses are relatively uncommon (but see Williams et al., 2004). Several reasons may account for this, including 1) the low taxon resolution of pollen data, which complicates interpretation; 2) the differential pollen productivity of species, also a complication for interpretations; and 3) the challenges of organizing large-scale collaborative efforts, in terms of both content (i.e. data) and process (i.e. people). Thus, while large syntheses of pollen data do exist, such studies are often driven by the need to test models (Farrera et al., 1999; Prentice et al., 2000; Bigelow et al., 2003; Pickett et al., 2004) rather than to understand past vegetation changes directly. This latter type tends to focus on sub-continental scales (e.g., in Europe (Huntley, 1990; Bradshaw and Hannon, 2004), parts of South America (e.g. (Heusser, 2004) and North America (Williams, 2002; Williams et al., 2004; Shuman et al., 2002b; Shuman et al., 2002a)). Thus far, paleoclimatologists and paleoecologists have reconstructed *global* vegetation for only a few carefully-selected time slices (e.g. the Last Glacial Maximum [LGM] and 6000 years ago) (Prentice and Webb, 1998).

The Global Charcoal Database was released in 2008 by the Global Palaeofire Working Group (GPWG), an international research team co-led by Sandy Harrison and Mitchell Power at the University of Bristol, and by Patrick J. Bartlein and myself at the University of Oregon. I designed and built the framework for the GCD after producing a series of proof-of-concept maps that demonstrated that 1) there were enough charcoal records to warrant the development of a truly global-scale charcoal database, and 2) coherent patterns were evident in the data based solely on changes in the level of "background" charcoal (Marlon et al., 2006). The GPWG emerged during this process and after discussions at the International Geosphere Biosphere Programme (IGBP) Fast-Track Initiative on Fire. The goal of the GPWG was to develop a complete global paleofire dataset for research purposes (rather than archival purposes) and with the primary aim of using the dataset to test global fire models and to better understand past changes in fire regimes. The dataset built upon the collection of publicly-available charcoal records in the International Multiproxy Paleofire Database (IMPD) – an archival database containing tree-ring and charcoal-based fire-history records from the Americas. The IMPD had been created in 2002 by a group of researchers from the University of Arizona's Laboratory of Tree-Ring Research, Rocky Mountain Tree-Ring Research, Inc., the University of Oregon's Environmental Change Research Group, and several other institutions to serve as a public archive of different kinds of fire-history data on the National Oceanic and Atmospheric Administration (NOAA) paleoclimate website (URL: http://www.ncdc.noaa.gov/paleo/impd/paleofire.html).

The demand for large regional- to global-scale paleoenvironmental datasets and analyses is unlikely to diminish in the near future. Such data form baseline information about global-scale climate and ecosystem dynamics, they provide validation tools for Earth system models, and they are a rich source of data that can be explored to address current and future environmental challenges (Gajewski, 2008). Insights from paleoenvironmental datasets can greatly enhance our ability to understand and thus mitigate against and adapt to climate change and its impacts (Whitlock et al., 2003; Willis and Birks, 2006; Gavin et al., 2007).

Research Questions and Dissertation Organization

The geography of fire is broadly determined by the interactions of climate, vegetation and people. This dissertation explores this geography and its controls on long time scales and broad spatial scales.

The research presented here was guided by the following questions:

1) How do variations in climate control broad-scale patterns of fire activity?
2) How has global biomass burning varied since the Last Glacial Maximum 21,000 years ago?
3) How did fire regimes in North America respond to the abrupt climate changes associated with the Younger Dryas climate reversal about 13,000 years ago, and what evidence is there for continental-scale burning associated with a hypothesized comet impact at that time?
4) How has biomass burning varied during the past 2000 years, and what factors best explain this variation?

Chapters II, III and IV address these questions. Chapter II introduces a new global dataset based on sedimentary charcoal records compiled from a combination of peer-reviewed literature, non-peer-reviewed reports and documents, and from unpublished data obtained directly from scientists. The data are synthesized to produce trends in biomass burning at regional to global scales as well as global time-slice maps for selected intervals. This chapter also sets the broad context for the syntheses and analyses presented in Chapters III and IV, which examine subsets of data from the global dataset during specific time intervals.

Chapter II was published in the journal *Climate Dynamics* (Power, Marlon et al., 2008) as a co-authored article with M.J. Power, N. Ortiz, P.J. Bartlein, and S.P. Harrison (who all contributed to the research design, data preparation, analysis and manuscript writing), F.E. Mayle, A. Ballouche, R. Bradshaw, C. Carcaillet, C. Cordova, S. Mooney, P. Moreno, I.C. Prentice, K. Thonicke, W. Tinner, C. Whitlock, Y. Zhang, and Y. Zhao (who attended a workshop that focused on data cleaning and interpretation), and 65 others who contributed data and edited the manuscript). My ideas and efforts were central to the project from start to finish. Specifically, I contributed 1) the design and implementation of the pilot study for the project, which showed the feasibility of collaboratively synthesizing a large body of heterogeneous information; 2) the design and construction of the Global Charcoal Database (GCD) that houses the data; 3) the collection of half of the data for the project; and 4) other elements of the overall research design, data analysis and writing.

Chapter III uses database records from North America to examine fire activity during a period of large and sometimes rapid climate changes between 15,000 and 10,000 years ago. This interval is referred to as the "Last Glacial-Interglacial Transition" or LGIT. Chapter III was co-authored with P. J. Bartlein, M. K. Walsh, S. P. Harrison, K. J. Brown, M. E. Edwards, P. E. Higuera, M. J. Power, R. S. Anderson, C. Briles, A. Brunelle, C. Carcaillet, M. Daniels, F. S. Hu, M. Lavoie, C. Long, T. Minckley, P. J. H. Richard, A.C. Scott, D. S. Shafer, W. Tinner, C. E. Jr. Umbanhowar, and C. Whitlock and was published in the *Proceedings of the National Academy of Sciences* (Marlon et al., 2009). With support from P. J. Bartlein, I designed the research, collected additional (pre-existing) charcoal and pollen data, analyzed the data and wrote the manuscript. M.K. Walsh, S. P. Harrison, K. J. Brown, M. E. Edwards, P. E. Higuera and M. J. Power provided assistance with analysis, interpretation and editing, and the remaining authors provided editorial assistance and contributed charcoal and pollen data.

Chapter IV takes a closer look at the past 2000 years, when populations, agriculture, and civilizations were rapidly expanding, and when relatively large variations in climate occurred. The chapter was published with P.J. Bartlein, C. Carcaillet, D.G. Gavin, S.P. Harrison, P.E. Higuera, F. Joos, M.J. Power, and I.C. Prentice in *Nature Geosciences*. All aspects of this work were collaborative, but I contributed the majority of time to data collection and organization and led the writing and preparation of the manuscript. I also made substantial contributions to the research design and analysis.

There are three main findings from this dissertation. First, climate is a major determinant of fire patterns at regional- to global-scales. When the climate cools, burning decreases; when the climate warms, burning increases. Furthermore, rapid warming apparently leads to rapid increases in biomass burning and fire frequency. Second, human activities during the past century (e.g., agriculture, grazing, and fire suppression) have led to a dramatic and rapid decrease in fire activity. Third, the combination of conclusions one and two above (i.e., a strong fire-climate link and a massive reduction in burning during the 20th century that allowed fuels to build up in many areas) in conjunction with current environmental changes such as increasing tree mortality (Mantgem et al., 2009; Adams et al., 2009), increasing pest outbreaks, and increasing biomass (Mickler et al., 2002), appear to be creating a "perfect storm" for unusually high fire activity in the coming decades in some areas. Additional details on these conclusions and a discussion of areas in need of further research are presented in Chapter V.

CHAPTER II

CHANGES IN FIRE REGIMES SINCE THE LAST GLACIAL MAXIMUM: AN ASSESSMENT BASED ON A GLOBAL SYNTHESIS AND ANALYSIS OF CHARCOAL DATA

This chapter has been published as a co-authored manuscript in the journal *Climate Dynamics* (Power et al. 2008).

1.0. Introduction

Fire has direct and important effects on the global carbon cycle, atmospheric chemistry, and in regulating terrestrial ecosystems and biodiversity. Fire has direct and important effects on the global carbon cycle, atmospheric chemistry, and in regulating terrestrial ecosystems and biodiversity (Gill et al. 1995; Cofer et al. 1997; van der Werf et al. 2004). Uncertainty over the effects of future climate change upon the incidence of fire, and the importance of vegetation-climate-atmosphere feedbacks, has fostered an increasing effort to develop coupled models of vegetation and fire (Prentice et al. 2007) to understand these future changes.

The fire regime of a given location is generally described in terms of the frequency, intensity, seasonality, extent and type of fires (Gill 1977; Box 1 in Bond and Keeley, 2005), all of which affect the area and amount of biomass burned. Changes in late-Quaternary fire regimes inferred from sedimentary charcoal records provide insights into the coupling and feedbacks between fire and major changes in climate and its boundary conditions or controls (orbital forcing, greenhouse gas concentrations), vegetation type, fuel amount, and human activity. Two pioneering charcoal-data syntheses (Haberle and Ledru 2001; Carcaillet et al. 2002) have established that there are inter-hemispheric linkages in palaeo-fire activity on millennial timescales. There has been no attempt, however, to synthesize the palaeo-charcoal records in order to examine the spatial patterns of fire activity at a global scale.

The aim of this paper is therefore to map global patterns of charcoal abundance, (which we take to be an index of the overall fire frequency, intensity and extent of the regional fire regime), at 3000-year intervals since the Last Glacial Maximum (LGM, conventionally centred at 21,000 cal yr BP). We offer some insights into the relationship between changes in fire regimes and potential large-scale controls on fire, but we do not attempt to explain the observed changes in regional fire regimes through time. To do this in a coherent way would require carefully-designed experiments using a vegetation-fire model to differentiate between competing explanations and evaluation of the derived patterns through comparison with observations. The global charcoal database used here to map changing fire regimes through time constitutes a first step in providing the necessary empirical data for testing the validity of fire models under markedly different biological and physical conditions from present (Marlon et al. in prep). This data will ultimately allow us to test hypotheses about the changing controls of fire regimes on glacial-interglacial timescales.

1.1. Controls on Fire and Charcoal Abundance

The incidence of fire over space and time is influenced by complex interactions between climate (over many scales of variability), fuels (type, amount, and arrangement), and ignition (whether anthropogenic or lightning). For example, at the scale of biomes (i.e. major vegetation types) the dominant role of climate on fire is demonstrated by the marked difference in fire frequency between highly humid climates (e.g. northern Europe or western Amazonia) and climates having a prolonged and severe dry season (e.g. the Mediterranean region or the subtropical savanna regions of Africa and South America). Within biomes, fire frequency and the amount of biomass burned vary temporally with changing climate, fuel, and ignition (Pyne et al. 1996).

The fire regime in a particular region is registered in sedimentary charcoal records through total charcoal abundance (per unit of sediment), which is proportional to the total biomass burned in a given depositional environment, (Marlon et al. 2006; Thevenon et al. 2004), and as peaks in charcoal accumulation, which mark individual fires in sedimentary records of sufficiently high sampling resolution (e.g. Power et al. 2006). Two examples illustrate the utility of charcoal abundance as an indicator of fire occurrence. In the northwestern USA, a region where fires have remained relatively frequent through time, palaeoecological data show that variations in charcoal abundance are closely associated with changes in the relative abundance of forest (as opposed to non-arboreal environments such as tundra and grassland), demonstrating a strong positive relationship between fire as sensed by charcoal abundance and biomass (i.e. fuel load) in this region (Marlon et al. 2006). In the rainforest-savanna ecotone regions, which have experienced a decrease in fire as forests have expanded since the early Holocene (e.g., Burbridge et al. 2004), fossil charcoal data reveal a negative relationship between fire (i.e. charcoal abundance) and biomass (i.e. fuel load). The contrasting relationship between fire occurrence and available fuel reflects the marked differences in the specific influence of climate and vegetation on fire between these two biomes. In both cases, however, total biomass burned is reflected by the overall charcoal abundance (Higuera et al., 2007). These examples show the magnitude of biomass burned, and implicitly aspects of the fire regime that influence the amount of material burnt is not solely determined by the nature of the vegetation. Thus, it is necessary to reconstruct changes in fire regimes in response to climate changes through time explicitly and independently of vegetation, which we do here by focusing on overall charcoal abundance.

A number of issues could influence the fidelity of overall charcoal abundance as an indicator of fire regime. Charcoal taphonomy and basin morphometry can have important influences upon charcoal deposition within lake or mire basins (Whitlock and Millspaugh 1996; Marlon et al. 2006). Several studies have suggested that macroscopic charcoal (i.e. large charcoal particles that can be removed from sediments by sieving) reflects local fires while microscopic charcoal (i.e. charcoal particles of a size commonly recorded from pollen slides) reflects fires on a more regional scale (Clark 1998; Long et al. 1998; Tinner et al. 1998; Carcaillet et al. 2001). Peaks in abundance of macroscopic charcoal could reflect higher energy sediment inputs to a basin (e.g., sudden inwash of coarse, clastic material rather than any change in fire regime: see Thevenon et al. 2003), or rare instances of long-distance transport (Tinner et al. 2006). Comparisons of the age of charcoal peaks with those of known fires and the overall charcoal input with estimates of area burned suggests that despite these issues charcoal deposition in lakes or bogs provides a useful index of overall fire regime (Tinner et al. 1998; Gardiner and Whitlock 2001; Whitlock and Bartlein 2004). For alluvial charcoal records, which are much less common, summed probability distributions for a large (n = ~50-100) sample of radiocarbon dates on fire-related deposits indicate relative changes in fire, and the charcoal content, thickness, and depositional processes of these deposits allow inferences on general fire severity (e.g., Meyer et al. 1995; Pierce et al. 2004). Although these various controls on charcoal deposition cannot be ignored when making inter-site comparisons at local or regional scales (e.g., Carcaillet et al. 2002), the dominance of variations in large-scale climate and biome type in controlling the fire regime at a regional- to continental-scale should lead to coherency in observed changes in charcoal abundance on millennial timescales among locations with similar climate, vegetation and human impact.

2.0. The Global Charcoal Database

The Global Charcoal Database (GCD) contains information about palaeofire regimes in the form of sedimentary charcoal records from sites across the globe since the LGM. Published and unpublished charcoal data were acquired from a network of sites between 70° N and 70° S (http://www.bridge.bris.ac.uk/projects/QUEST_IGBP_Global_Palaeofire_WG).

Many methods were used for recording changes in charcoal abundance in a sedimentary context, and the database therefore contains a variety of different types of records (e.g., both macroscopic and microscopic charcoal, the latter mostly from pollen-slides) from a variety of site types (e.g., lake, mire, and alluvial-fan sediment records), and with varying temporal resolution and dating control. The database therefore includes a large amount of descriptive data (metadata) about both the sites and the charcoal samples. It also contains detailed information on site chronology: the radiocarbon dating technique (AMS or conventional), the sample size, standard deviations, and calibrated ages.

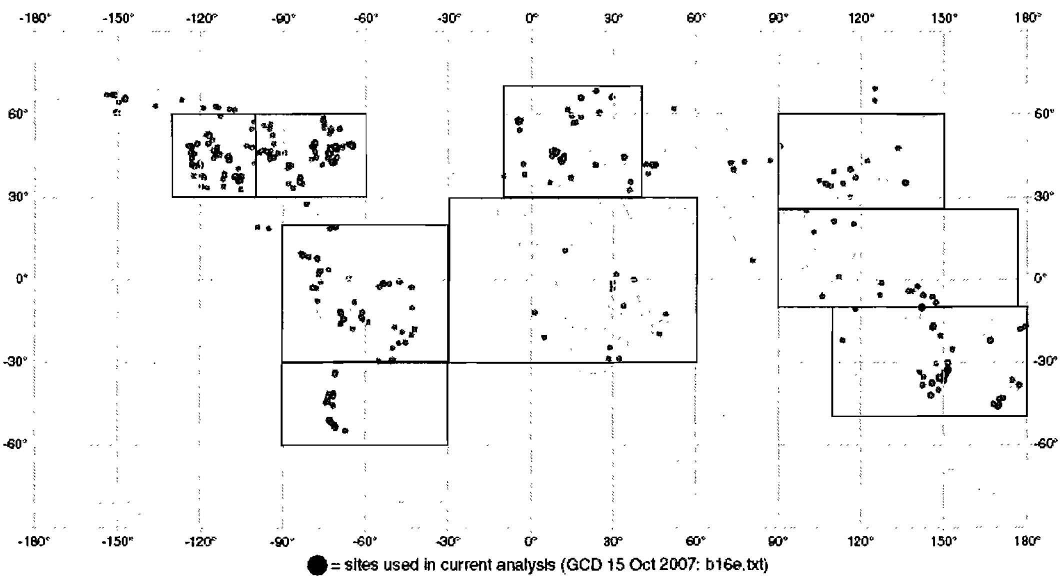


Figure 1. Inventory of global charcoal records currently used in the analysis (*gray dots*) and regional delineations (*black boxes*) where sites were averaged together for regional summaries of fire regimes (see Fig. 5a).

The database currently contains charcoal records from 405 sites (Figure 1), 355 of which are used in this analysis and 33 of which have records back to the LGM (here defined as 21,000 ± 500 calendar yr BP). Records of the mean charcoal value at a site over a 1000-year long interval were extracted for seven time slices (i.e. 3000, 6000, 9000, 12,000, 15,000, 18,000, and 21,000 cal yr BP); thus the fire regime is represented by mean charcoal accumulation over the mapped intervals 3000±500, 6000±500, 9000±500, 12,000±500, 15,000±500, 18,000±500, and 21,000±500 cal yr BP. These values were compared with an estimate of the modern (pre-industrial) fire regime, based on the mean charcoal value for the period 100-1000 cal yr BP. The choice of a 1000-year window to characterize the charcoal record for each time slice reflects the fact that charcoal deposition and accumulation in sediments is intrinsically highly variable (Carcaillet and Richard, 2000) and it was necessary to select a period that would avoid single, anomalous fire events in order to elucidate longer-term (in this case millennial-scale) trends. The 1000-year window allows us to make use of sites with low sampling resolution and few radiocarbon dates. Other palaeoenvironmental data syntheses, and in particular the BIOME 6000 reconstructions of changing vegetation patterns (e.g. Prentice et al., 2000; Bigelow et al., 2003; Williams et al., 2004), have made use of a 1000-year sampling window. Using a similar sampling approach facilitates comparison with these data sets, and will allow us to investigate the relationships between e.g. changes in vegetation and fire regimes at a later stage.

2.1. Data Acquisition and Age Models

Charcoal data were provided by the original author or came from the individual charcoal analyst as unpublished data, or from the published literature. Data were extracted from the literature by digitizing the original published figures and using the plotted values to produce tables of charcoal values by depth or by age.

As a consequence of using both published and unpublished data, some of which were produced more than a decade ago, age models based on a consistent calibration had to be developed for each site. Over 4000 radiocarbon dates, calibrated and uncalibrated, from 405 sites were entered into the charcoal database. All radiocarbon dates that were uncalibrated were converted to calibrated years BP using the Fairbanks et al. (2005) calibration curve and program (http://radiocarbon.ldeo.columbia.edu/research/radcarbcal.htm). The mean calibrated age was selected for each radiocarbon date. In many cases, individual charcoal samples were expressed by depth or radiocarbon years and required new calibrated age models. Age calibration and the creation of age models were performed only for records with at least two radiocarbon dates. When the surface samples from sediment cores or soil profiles was established as modern, an age of -50 cal yr BP (2000 AD) was assigned. Considering the multi-centennial resolution used to analyze these records, assigning ages to the surface samples for creating age models had little impact on the final result. In cases where the date of core collection could be established, that date was assigned to the uppermost sample, for example, a core collected in 2003 AD was assigned a core top age of -53 cal yr BP. Age models were constructed using all available calibrated ages, including dated tephra layers, and pollen stratigraphic ages, and were based on four possible age model styles; (1) linear interpolation, (2) a polynomial constrained to pass through zero, (3) an unconstrained polynomial fit, and (4) a cubic smoothing spline (Ripley and Maechler, 2006). The "best fit" age model was selected for each record, based on goodness-of-fit statistics and the appearance of the resulting curve.

2.2. Standardizing Charcoal Data and Calculating Anomalies

Charcoal values (e.g., influx, concentration, charcoal/pollen ratios, gravimetrics, image analysis) can vary by over 10 orders of magnitude among and within sites (Figure 2) because of the broad range of record types, site characteristics, and methodological or analytical techniques. It was therefore necessary to standardize the records to facilitate comparisons between sites and through time. The standardization procedure involves three calculations applied to each site record (see Figure 3): (1) rescaling values using a minimax transformation, (2) homogenisation of variance using the Box-Cox transformation, and (3) rescaling values once more to Z-scores. The minimax transformation rescales charcoal values from a given site record to range between 0 and 1 by subtracting the minimum charcoal value found during the record from each charcoal value, and dividing by the range of values:

$$c_i' = (c_i - c_{\min}) / (c_{\max} - c_{\min})$$

where c_i' is the minimax-transformed value of the *i*th sample in a particular record, c_i, and $c_{\max}$ and $c_{\min}$ are the maximum and minimum values of the c_i's. The inherently skewed distribution of charcoal values, with a long upper tail, would result in a preponderance of negative anomalies without further transformation. We therefore transformed the rescaled values using the Box-Cox transformation:

$$c_i^* = \begin{cases} ((c_i' + \alpha)^{\lambda} - 1)/\lambda & \lambda \neq 0 \\ \log(c_i' + \alpha) & \lambda = 0 \end{cases}$$

where is the transformed value, λ is the Box-Cox transformation parameter and is a small positive constant (here, 0.01) added to avoid problems when c_i' and λ are both zero. The transformation parameter λ is estimated by maximum likelihood using the procedure described by Venables and Ripley (2002). The transformed data were rescaled once more, as Z-scores, so all sites have a common mean and variance (Figure 3e)

$$z_i = (c_i^* - \bar{c}^*_{(4\text{ka})}) / s^*_{c(4\text{ka})}$$

where, $\bar{c}^*_{(4\text{ka})}$ is the mean minimax-rescaled and Box-Cox transformed charcoal value over the interval 4000 to 100 cal yr BP, and $s^*_{c(4\text{ka})}$ is the standard deviation over the same interval.

There is considerable variation in the length of the charcoal records in the database, so a common base period (100-4000 cal yr BP) was used to calculate the mean and standard deviation for each site. The choice of the last 4000 years as a base period represents a compromise between a period long enough to not be dominated by sample-to-sample variability within an individual record and one short enough to not exclude a large number of database records from the subsequent analyses. Most of the records in the database (95%) are at least 4000-years long. The base period does not include the last

100 years (i.e. the post-industrial period) because of the intensification of most modern human activities during this part of the fire record. Mean charcoal values, expressed as average Z-scores for a 1000-year window, were calculated for each site at 500-year intervals. Changes in charcoal values through time are expressed as Z-score anomalies (i.e. the difference between the mean Z-score for each 1000-year window and the mean Z-score for the "modern", where "modern" is defined as the interval 1000 to 100 cal yr BP).

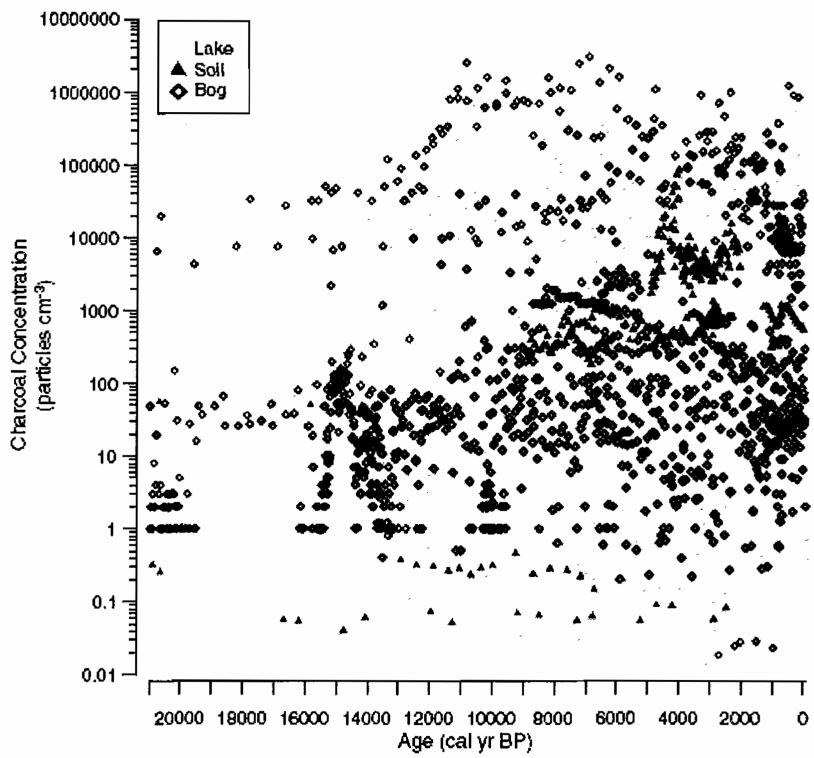


Figure 2. Scatter diagram for all charcoal concentration (particles/cm^{-3}) values contained within the database. Three types of depositional environments are represented; gray circle represent lake sediment records, open diamonds are from bog sediment records, and dark triangles are from soil charcoal records. All data are plotted by original charcoal units, illustrating the heterogeneity of analytical methods and laboratory techniques used to show charcoal-abundance variations through time. The >10 orders of magnitude represented by charcoal concentration values illustrates the need for standardization of these data.

2.3. Creation of 3000-Year Steps of Time-Slice Maps (1000-Year Window)

Time-slice anomaly maps of modern (1000 to 100 cal yr BP) mean Z-score minus palaeo mean Z-score were created at 3000-year intervals (Figure 4a-c, Figure 5a-c, Figure 6a, 6b) from 21,000 cal yr BP to present. The anomalies were classified into 5 roughly equal-frequency groups ranging from those >+1.15 (strong positive anomalies, dark red) +1.15 to +.365 (positive anomalies, red), +0.375 to -0.375 (weak positive or negative anomalies, gray), -0.375 to -1.15 (negative anomalies, blue) to those <-1.15 (strong negative anomalies, dark blue).

3.0. Results: Changes in Fire Regime Between LGM and Present

Charcoal records of changes in fire regime during part or all of the past 21,000 years are available from 405 sites (Figure 1). There are relatively few charcoal records available for the LGM and the early phase of the deglaciation. Although there are >200 sites with records for the past 10,000 calendar years BP (i.e., most of the Holocene, that is about the last 11,500 calendar years), some regions (e.g., the boreal forest zone of Russia and most arid regions) are only represented by a few sites even in the Holocene epoch. As a result, the interpretation of regional patterns presented here should be considered preliminary, particularly those for the deglacial period (i.e. ca. 15,000 to 12,000 cal yr BP). Nevertheless, the maps show broad-scale changes in charcoal abundance through time that can be interpreted as reflecting changes in fire regime, and our analyses provide hypotheses that can be tested as more data becomes available.

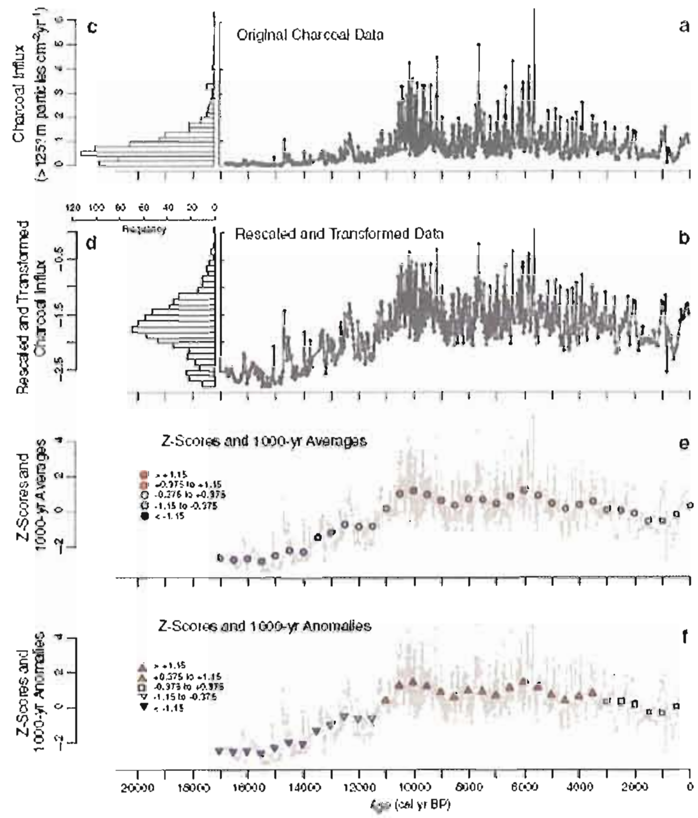


Figure 3. An example of charcoal standardization performed on all charcoal records contributing to the global paleofire database. The original charcoal abundance (from Millspaugh et al. 2000) values (a) for each record were rescaled to range from 0.0 to 1.0 over the whole of the record, the rescaled values were then transformed (b) using the Box-Cox power transformation to approach normality where possible (compare histograms in c and d), with the transformation parameter estimated using maximum likelihood. The transformed values were then standardized or converted to Z scores (e) using the mean and standard deviation for each record over the interval (base period) from 4,000 to 100 cal yr BP. Anomaly Z scores, or differences in charcoal values between the "modern", defined as between 1,000 to 100 cal yr BP, and the base period (f) were then calculated for each record.

We discuss the spatial patterns of charcoal anomalies at each time step relative to present (here defined as the interval 1000-100 cal yr BP: Figure 4a) in terms of a spatial hierarchy from global through regional to local or landscape-scale patterns. We have compared our "modern" basemap with observed global-scale patterns of fires (e.g.

Carmona-Moreno et al., 2005). There is remarkable consistency in the regional patterning, after allowing for regions where human-set fires are known to be important – an impact which we have minimized in our baseline map by excluding the records of the past 100 years.

Interpretations of fire regimes at the LGM are constrained by the small number of sites (n=33), although South America, southeast Australia, Europe, and Indonesia are represented. Globally, 66% of these charcoal records show less-than-present fire. Sites from Europe, Africa, and the mid and high latitudes of South America record lower–than-present fire during the LGM, but greater-than-present fire is recorded in the southern latitudes of Australia and tropical latitudes of Southeast Asia (Figure 4b). Within these broad regions, however, there is spatial heterogeneity in the change in fire regimes. For example, sites in southern Australia record greater-than-present fire whereas adjacent sites to the north reveal less-than-present fire. Charcoal records (n = 4) from Indonesia and Papua New Guinea, show both similar-to-modern and strong negative anomalies, or less-than-present fire.

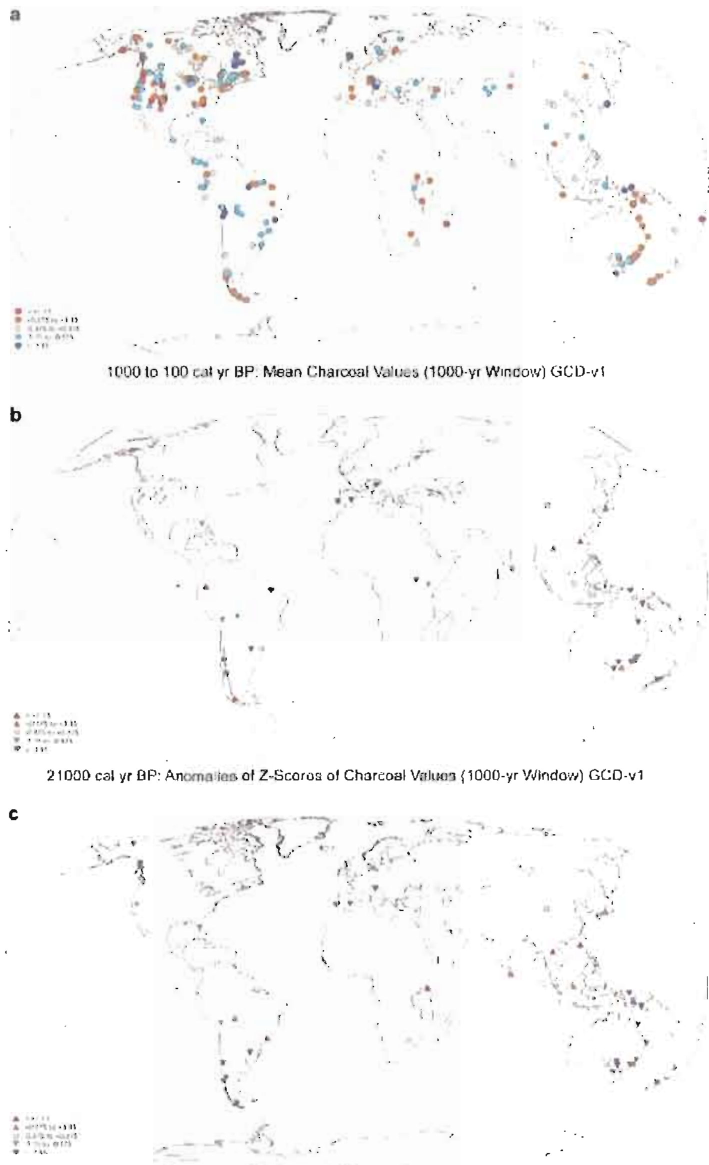


Figure 4. (a) Global map of mean Z scores of charcoal values for 1,000 to 100 cal yr BP (present). Triangles indicate sites with higher (red) or lower (blue) average charcoal values during the last millennia when compared to the last four millennia. The continental palaeogeographies are mapped for 21,000 and 6,000 cal yr BP. The anomaly maps reveal both the spatial heterogeneity as well as regional coherencies of global charcoal. (b) Global anomaly map for 21,000 cal yr BP. (c) Global anomaly map for 18,000 cal yr BP.

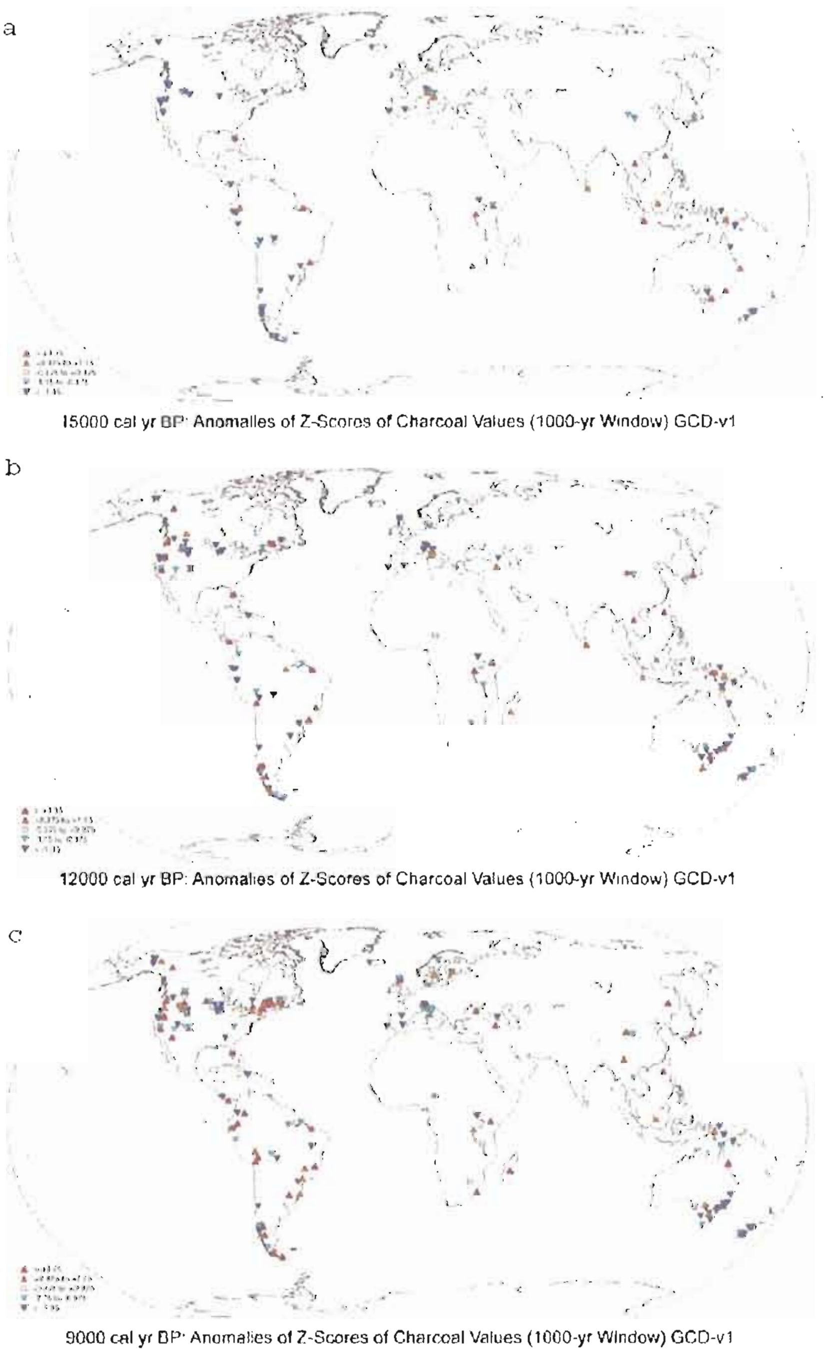
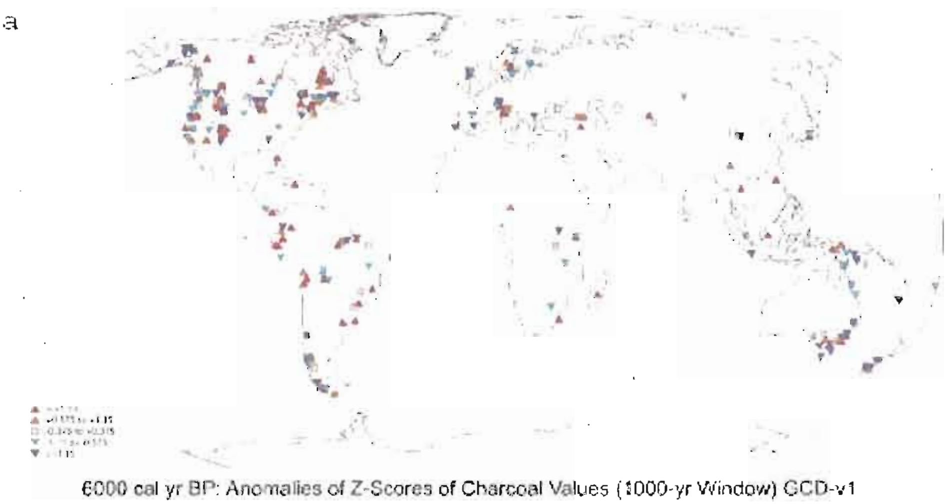


Figure 5. Anomalies of Z-scores of charcoal values, 1,000 yr window. (a). 15,000 cal yr BP. (b) 12,000 cal yr BP. (c) 9,000 cal yr BP.

a



b

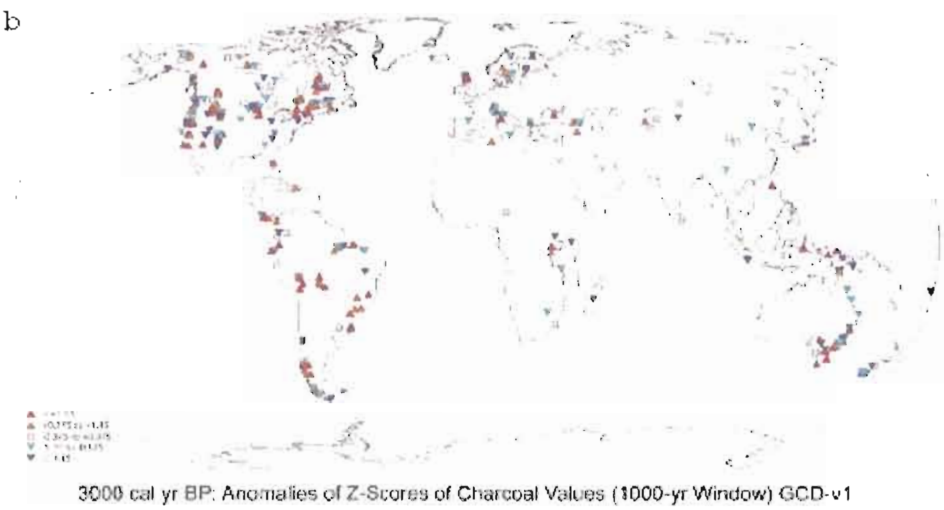


Figure 6. Anomalies of Z-scores of charcoal values, 1,000 yr window. (a). 6,000 cal yr BP. (b) 3,000 cal yr BP.

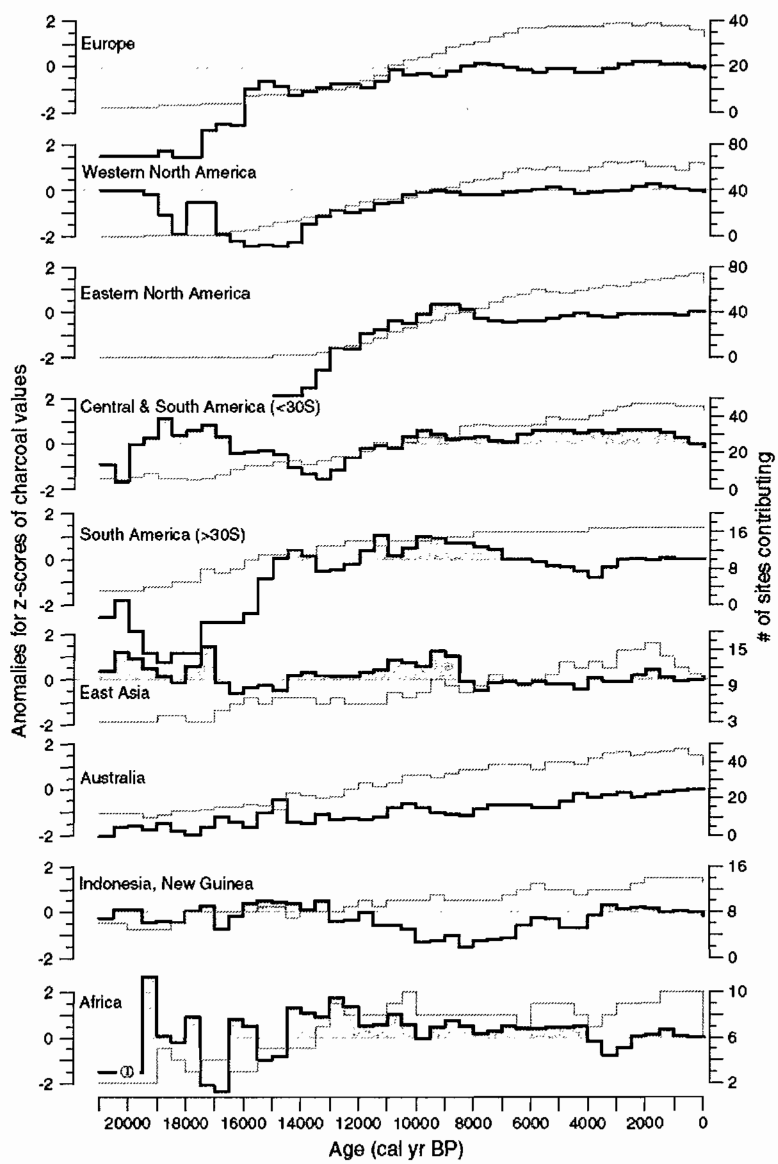


Figure 7. Global and regional summaries of average anomaly Z scores of charcoal values. The number of sites (gray line) contributing to each regional summary (see Fig. 1) are compared to the regional average anomaly Z scores (black line), revealing the potential influence of increasing sites for each regionally averaged time series. Periods within the time slices of positive charcoal anomalies relative to present are shaded gray. Large charcoal anomalies that extend beyond the +2 or -2 are indicated by circled arrows.

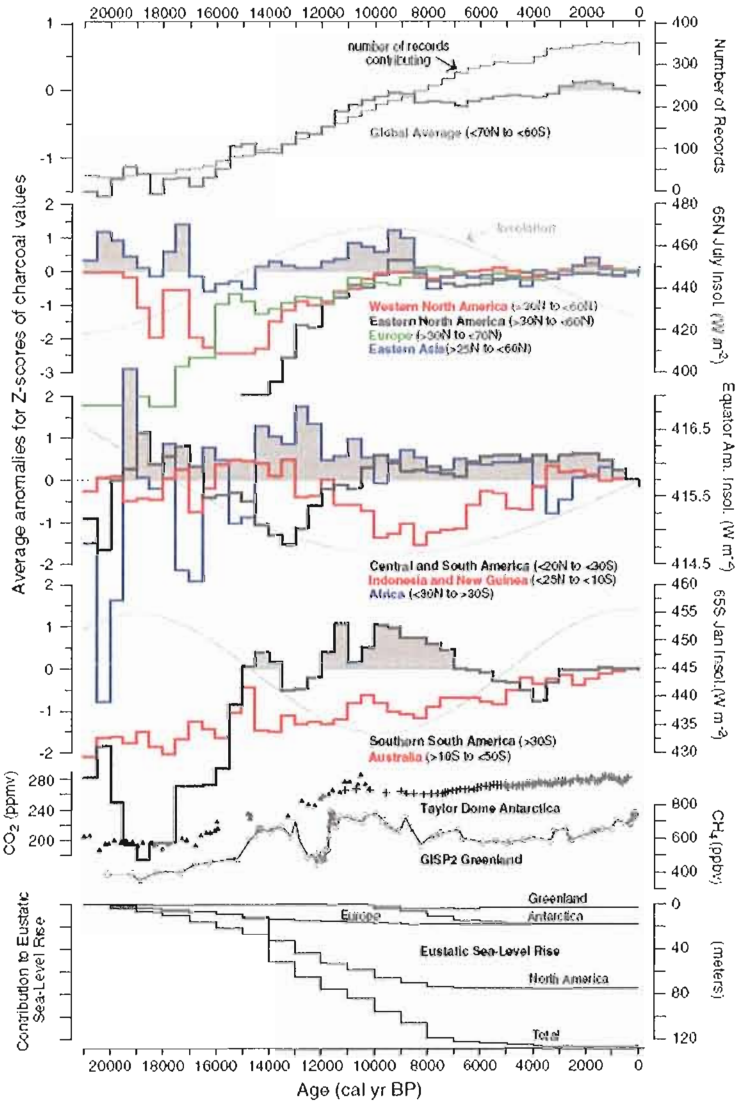


Figure 8. Global average anomalies for Z scores of charcoal values for all records (top). Regional summaries are grouped by similar latitudes and compared with summer insolation (gray line) at 65°N and S for the high latitude sites and average annual insolation for the circum-equatorial sites (Berger and Loutre 1991). Atmospheric carbon dioxide concentrations are shown from Taylor Dome, Antarctica ice core records (Indermühle et al. 1999, 2000; Monnin et al. 2001) and a methane record from GISP2 (Brooks et al. 1996). Contributions to eustatic sea level rise (Peltier 2004, Ice-5G) are also shown for comparisons with global charcoal values.

There was little change in spatial patterns of fire between 21,000 and 18,000 cal yr BP, with 58% of the records showing less than present fire (n = 40, global average Z-score anomaly = -1.3) at 18,000 cal yr BP. Charcoal records from Australia and New Guinea indicate that fire was generally less than present (Figure 4c). Records from southeast Asia and Indonesia show greater-than-present fire. In South America fire was less than present in the high and mid latitudes and greater than present at lower latitudes. The relatively few sites in North America, Europe and Africa suggest less-than present fire.

At 15,000 cal yr BP global fire was lower-than-present with 63% of charcoal records (n = 80, global average Z-score anomaly = -0.78) showing less-than-present fire (Figure 5a). The northern mid-latitudes (30-60° N) of North America (n = 15) show a consistent pattern of less-than-present fire while sites in Europe and Asia show spatial heterogeneity. Sites in Central and South America (n = 25) indicate lower-than present fire with the exception of those near the equator and in eastern Brazil, which show greater-than-present fire. In Africa, Australia, Indonesia, and Southeast Asia most records (n = 31) indicate similar-to or greater-than-present fire. In contrast, sites from New Zealand and two sites from New Guinea show less-than-present fire.

There is a significant increase in the number of charcoal records between 15,000 and 12,000 cal yr BP (n = 143 sites). A global pattern of less-than-present fire continues at 12,000 cal yr BP for 55% of the records. There are changes in the fire regime between these time intervals in several regions, including e.g., a reduction in fire in southeastern Australia (to levels less than present) and an increase in fire in extra-tropical latitudes of South America (to levels greater than present). In general, the mapped patterns show more spatial heterogeneity than during earlier periods (Figure 5b), but the globally averaged Z-score anomaly remains lower than present (-0.52). In the low latitudes of South America, areas of eastern Brazil continue to indicate greater-than-present fire whereas charcoal records west of the Amazon basin suggest less-than-present fire (with the exception of the Lake Titicaca record: Paduano et al., 2003). Several sites, not yet included in the global charcoal database, suggest charcoal has been largely absent from within the Amazon basin since the LGM (Colinvaux et al., 1996; Bush et al., 2004a; Bush et al., 2004b; Urrego et al. 2005). Positive charcoal anomalies from southeast Asia and Indonesia (and eastern Brazil) suggest fire was greater than present at 12,000 cal yr BP. Tropical charcoal records from both sides of the Pacific, however, including records from eastern New Guinea, northeastern Australia and from western Colombia, Ecuador and Peru, show less-than-present fire. In southeastern Australia, a shift to less-than-present fire occurred at most mainland sites, although Tasmanian records indicate greater-than-present fire and sites in New Zealand continue to show less-than-present fire. Fire in the mid-latitudes of Europe and North America, although increased relative to 15,000 cal yr BP, remained less than present (n = 55, -0.9 average Z-score anomaly). At regional scales, fire increased from glacial times at sites in northeastern and in western North America.

The interval between 12,000 and 9000 cal yr BP is marked by a significant change in fire regimes. Broadly considered, the northern extra-tropics and western tropics show increased fire and the southern extra-tropics and eastern tropics show a reduction in fire between these two intervals. By 9000 cal yr BP, however, the charcoal records (n = 209, -0.05 average Z-score anomaly) indicate spatial heterogeneity in fire regimes with 44% of all records showing less-than-present fire. Regional summaries (Figure 7) show greater-than-present fire throughout South America, eastern North America, eastern Asia, and Africa. Mapped patterns show spatial heterogeneity in fire regimes in western North America and Europe at the sub-continental scale, but at regional scales, the records are spatially coherent (Figure 5c). For example, predominantly greater-than-present fire occurred in northeastern North America, while less-than or similar-to present fire occurred at sites in southwestern Europe and in central North America. Similarly, greater-than-present fire occurred in southern Brazil and less-than-present (or complete absence) fire occurred at sites within the Amazon basin. Coherent patterns of less-than-present fire also can be seen at larger, sub-continental, scales in Australia and New

Zealand at 9000 cal yr BP. Sites in the northern mid-latitudes of Europe show heterogeneity in fire regimes at 9000 cal yr BP.

At 6000 cal yr BP, the charcoal records (n = 282) show continued spatial heterogeneity in fire regimes, with 43% of all records showing less-than-present fire, but regionally coherent patterns emerge (Figure 6a). In the Northern Hemisphere, regional summaries show greater-than-present fire in Central and South America, less-than-present fire in eastern North America, and heterogeneous conditions similar to modern across Europe (Figure 8). At regional-to-landscape spatial scales, however, positive and negative anomalies in fire regimes relative to present appear more spatially coherent. For example, regional patterns of fire across low latitudes (<30°S) of South and Central America indicate greater-than-present fire (n = 35, 0.59 average Z-score anomaly) in the neotropics. In the mid and high latitudes of South America (>30°S), fire was less than present along the Pacific coast but greater than present east of the Andes. Throughout the eastern tropics, including the low latitudes of Indonesia and eastern New Guinea, the fire regime was similar to or less than present at 6000 cal yr BP. In southeastern Australia and New Zealand, fire was mostly less than present (n = 25, -0.77 average Z-score anomaly), similar to conditions at 9000 cal yr BP.

By 3000 cal yr BP, global fire regimes were heterogeneous with 37% of all records showing less-than-present fire. In general, fire was greater than present in the neotropics, central and eastern Europe and less than present in northeastern Australia. The mid-latitudes of the Northern Hemisphere suggest a highly heterogeneous fire regime for 3000 cal yr BP but lower latitudes of Central and South America (~10-50°S) generally show greater-than-present fire. A more heterogeneous pattern of fire emerges for 3000 cal yr BP in the mid and high latitudes of Patagonia. Less-than-present fire characterizes the Eastern Hemisphere, including eastern New Guinea, New Zealand and eastern Australia. Sites in Indonesia and western Asia show increasing heterogeneity by 3000 cal yr BP. Sites throughout the mid- and high-latitudes of North and South America generally indicate heterogeneous fire regimes, although fewer sites show significant anomalies compared to modern. Examination of the individual charcoal time series (Figure 3) suggests that the patterns of change in fire regimes are not related to the analytical methods but rather reflect the slowly varying changes in the underlying climatic controls on vegetation and fire (see below).

4.0. Discussion: The Climatic Control of Observed Changes in Fire Regimes

Changes in regional climate have direct effects on fire regimes, through controlling the incidence of ignitions and the likelihood that fires will spread, and indirect effects through changing vegetation type and productivity, and hence available fuel load (Pyne et al. 1996). The major factors governing regional climate changes since the LGM are changes in the seasonal and latitudinal distribution of insolation, the disappearance of the Northern-Hemisphere ice sheets (and concomitant changes in land-sea geography), changes in the southern-hemisphere ice caps, changes in sea-surface temperature patterns and variability, and changes in atmospheric composition. All of these factors directly and/or indirectly influence regional-scale atmospheric circulation patterns. The LGM, ca. 21,000 cal yr BP, represents the global (though not regional) maximum of the extent of the ice sheets (Peltier 2004) and a time when sea level was ca. 120m lower than at present and tropical land areas were more extensive than today. Greenhouse gas concentrations (compared to pre-industrial) were low (Raynaud et al. 2003) and atmospheric aerosol loadings were high (Kohfeld and Harrison 2001). Ocean temperatures were, in general, lower than today with the largest changes occurring in high northern latitudes (Schäfer-Neth and Paul 2003). The transition from glacial to interglacial conditions was marked by asynchronous warming in the two hemispheres (Schaefer et al. 2006; Smith et al. 2005), with the Southern Hemisphere leading the Northern Hemisphere by up to two millennia (Labeyrie et al. 2003). Insolation changes became the major driver by the early Holocene, with regional climates responding to the increased seasonal contrast in insolation in the Northern Hemisphere and the correspondingly decreased seasonal contrast in insolation in the Southern Hemisphere (Berger 1978; Liu et al. 2004). These insolation anomalies changed towards reduced seasonal contrasts in the Northern Hemisphere and stronger seasonal contrasts in the Southern Hemisphere in the last 6000 years. These broad-scale changes in climate forcing can be used to explain much of the observed change in regional fire regimes at orbital timescales (the 21,000 yr precession cycle). Superimposed on these orbital-time scale changes were millennial- and shorter time-scale climate changes that typically were associated with smaller (spatial) scale anomaly patterns.

Time-slice maps at 500-yr intervals (not shown) that supplement those in Figure 4 and the individual charcoal Z-score anomaly time series (not shown) were used to divide the records into continental and regional groups of records with similar histories (Figure 7). Comparison of these grouped time series with time series of large-scale climate controls (Figure 8) suggest that the global charcoal record since 21,000 cal yrs ago can be divided into four relatively distinct intervals: 1) a glacial interval (typified by patterns at 21,000 through 16,000 cal yr BP) when global temperatures were low, it was generally drier than present, and terrestrial biomass was relatively low; 2) a late-glacial interval (15,000 through 12,000 years ago) when global (and particularly Northern Hemisphere) temperature increased, pronounced millennial climate variations were registered, and vegetation exhibited dramatic changes on a global scale (Williams et al. 2004); 3) an early- Holocene interval (from 11,000 through 7000 years ago), when monsoonal regions in both hemispheres were wetter than at present and regions under the influence of the subtropical high pressure system were drier, and 4) a mid-to-late Holocene interval when global climate approached that of the present, and ENSO and human influences on fire regimes became important.

4.1. Interpretation of Fire Regimes During the Glacial Interval (21,000 - 16,000 cal yrs BP)

Although there are relatively few charcoal records for the LGM and subsequent millennia, they show a consistent pattern of low fire (Figure 4b). Indeed the glacial interval is the period of lowest fire in the last 21,000 years. This is consistent with the fact that the global climate was generally (but not exclusively) colder and drier than present (Braconnot et al., 2007), leading to an overall reduction in terrestrial biomass (Francois et al. 2000) and thus a decrease in fuel availability.

At a regional scale, less-than-present fire in Patagonia at 21,000 cal yr BP is consistent with reconstructions of regional climates cooler than present (Markgraf 1993; Markgraf et al. 1992). Pollen and lake-level data from the Amazon basin suggests cooler climates during the LGM with average temperatures roughly 4.5-5°C less than present (see Liu and Colinvaux 1985; Bush et al. 1990; Bush and Silman 2004). Increased ice volume, lowered sea levels, cooler sea-surface temperatures, and decreased atmospheric carbon dioxide, combined with weakened subtropical-high pressure and intensified westerlies, would have contributed to widespread aridity in middle latitudes. In contrast, evidence from the high latitudes of Patagonia (Moreno et al. 1999) suggests the intensification of westerlies resulted in greater-than-present humidity during the LGM. Therefore, cold and wet conditions may have reduced fire in the middle and high latitudes of South America.

The charcoal-abundance records from southeastern Australia generally show less-than-present fire anomalies at 21,000 cal yr BP (Figure 4b), with sites further south showing greater-than-present fire. At the landscape scale, greater-than-present fire could reflect human activity in those regions (Kershaw and Nanson 1993). However, a north-south gradient in fire could also reflect changing latitudinal gradients in the seasonal cycle of insolation. Average January (austral summer) insolation values at 65°S were similar to present (455 watts m-2) during the LGM (Figure 8), which promoted greater seasonality in the Southern Hemisphere than during early-Holocene times. High summer insolation may have contributed to relatively dry and warm conditions across the middle latitudes of the Southern Hemisphere. Alternatively, high summer insolation at 65°S (Figure 8) may have resulted in aridity limiting fuel load and thus contributed to reduced fire.

The globally cooler- and drier-than-present climates at 21,000 cal yr BP limited fires across the middle and high latitudes of both hemispheres until after 16,000 cal yr BP. The main changes in fire regime after the LGM occurred in the low latitudes of South America and in southeastern Australia. In South America, fire increased at sites in southern and eastern Brazil while remaining low in western and southern South America. In southeastern Australia, lake level and pollen data suggest enhanced fluvial activity after the LGM (Nanson et al. 2003) and this may help to explain the further decrease in fire observed there.

4.2. Interpretation of Fire Patterns During the Deglacial Period (ca. 15,000 cal yr BP - 12,000 cal yr BP)

By 15,000 cal yr BP an east-west gradient of charcoal anomalies developed across South America, and there was an increase in fire throughout Australia and Indonesia (although not Papua New Guinea). In South America, evidence from Lake Titicaca (16-20°S) (Paduano et al. 2003) suggests a rapid climate shift in tropical climates after 17,700 cal yr BP, as fire first appeared but fuels remained limited around the Titicaca basin. The precise timing of tropical climate change and subsequent deglaciation of the central Andes is unclear (Seltzer 2001), and, so the regional controls of fire from 21,000 to 11,500 cal yr BP in tropical South America are difficult to identify (Smith et al. 2005). Regional controls of fire regimes at 15,000 cal yr BP in South America may be related to their proximity to the oceans and the role of the Andes in reducing moisture advection from the tropical Atlantic (Cook and Vizy 2006). In the mid- and high latitudes, late-glacial patterns in fire have been attributed to shifts in the position of the westerlies and millennial-scale climate variability (Whitlock et al. 2007; Huber et al. 2003; Moreno 2000). Cool and dry climates in the mid-latitudes likely reduced biomass production resulting in less-than-present fire. Lower-than-present sea surface temperatures in the southern Pacific (Lamy et al. 2004) as well as lowered sea level and expanded continental shelves throughout Australasia may have increased continentality and contributed to increased aridity and decreased annual average temperatures.

Regional scale controls of fire regimes in southeast Asia, Indonesia, and Australia at 15,000 cal yr BP may be related to lower sea levels (Peltier 1994; 2004) (Figure 8). Exposed continental shelves were colonized by tropical lowland forest and palynological evidence (Kershaw et al, 2001) suggests greater aridity than during the LGM in the western part of Indonesia than near New Guinea (Hope et al. 2004). Glaciers were likely still present on the highest mountains of New Guinea (Peterson et al. 2002), but increased moisture availability or decreased human activity may explain reduced fire at this time. Haberle and Ledru (2001) suggest that lower land temperatures and the increasing influence of the summer monsoon (Huang et al. 1997) may have contributed to reduced fire. Greater-than-present biomass burning 15,000 cal yr BP in southeastern Australia contrasts with the lower-than-present burning in most of South America despite being at similar latitudes. Treeless vegetation was promoted by drier and windier conditions across southeastern Australia following the LGM (Hope et al. 2004). Increased fire in Australia 15,000 cal yr BP relative to earlier may have been related to both climate controls and human activity (Black and Mooney 2006; Haberle and David 2004). Fire slightly increased in Europe between ca. 14,500 and 12,000 cal yr BP, but remained less than present. Cooler climates and the presence of continental ice sheets in the high latitudes of the Northern Hemisphere may have limited fire in northern Europe, but the increasing terrestrial biomass and the possible role of anthropogenic fire for forest clearing may have contributed to increased fire at sites in southern Europe.

4.3. Interpretation of Fire Patterns During the Early Holocene Interval (ca. 11,000 cal yr BP - 7000 cal yr BP)

Dominant influences on global fire regimes leading into the early-Holocene interval include the rapidly changing boundary conditions (e.g., Kutzbach et al. 1998) of decreasing ice-sheet size, rising sea-surface temperature and sea level (Peltier 2004), and vegetation changes (Williams et al. 2004; Huntley and Birks 1983), including reforestation of regions formerly covered by glacial ice. Greater-than-present summer insolation resulted in warmer and drier summers in regions of the Northern Hemisphere influenced by stronger-than-present subtropical high pressure. Regional summaries of fire regimes suggest increased spatial heterogeneity during this interval with marked shifts in all regions toward either stronger positive or negative anomalies in fire (Figure 7 and 8). Records from North America, Europe and South America show shifts toward increased fire while records from Australia show lower-than-present fire with shifts toward decreasing fire culminating around 8000 cal yr BP. In southern South America and western North America, these patterns have been attributed to the regional changes caused by increased annual and summer insolation and to ice-sheet dynamics (in the North Hemisphere; Carcaillet and Richard, 2000) and increased annual and winter insolation (in the Southern Hemisphere) in the early Holocene (Whitlock et al.,2007; Whitlock and Bartlein, 2004). These large-scale changes in the climate systems would have affected regional circulation patterns, including the strength and position of the westerlies, the strength of the monsoons and subtropical highs, and ultimately the duration of the fire season.

In southern Europe, evidence from Lago Piccolo di Avigliana and Lago di Origlio suggests increased fire starting from ~10,500 cal yr BP (Finsinger et al. 2006). A non-linear response of vegetation to higher drought stress and fire resulted in the expansion of Corylus (hazel), which re-sprouts after fire events (Delarze et al. 1992; Tinner et al.

1999) and is more drought-tolerant than other deciduous trees (Huntley 1993; Finsinger et al. 2006).

Increasing fire in the Northern Hemisphere and South America can also be compared with records of atmospheric carbon dioxide from Antarctica (Indermühle et al. 1999) (Figure 8). Increased CO2 after 12,000 cal yr BP may be related to increasing fire in the tropics and high latitudes of South America as well as temperate and boreal forests of the Northern Hemisphere. Recent evidence suggests global fires may also contribute to atmospheric methane (van Aardenne et al. 2001; Andreae and Merlet 2001; van der Werf et al. 2007). Methane records from the GISP2 ice core reveal increased variability around 12,000 cal yr BP (Figure 8), and have been partly explained by high latitude summer insolation forcing (Brook et al. 1996) and developing boreal peatlands (MacDonald et al. 2006) but there may also be linkages to increased fire. In addition to these climate explanations, human populations were increasing in the Americas and may have locally contributed to the changing incidence of fire (Cooke 1998).

Globally, fire decreased after 9000 cal yr BP, with significant decreases in eastern Asia, Indonesia, eastern North America, and Africa by ~8000 cal yr BP. A record of decreasing atmospheric carbon dioxide from Taylor Dome, Antarctica (Monnin et al. 2001; Indermühle et al. 1999) during the early Holocene may be linked to the reduction in fire throughout the Americas, Africa, Indonesia and Australia. Low incidence of fire during the early Holocene in Indonesia and Papua New Guinea (Haberle and Ledru 2001) and eastern Australia (Black and Mooney 2006) has previously been attributed to a relatively stable climate at that time. Whereas the first agricultural activities, beginning around 10,000 cal yr BP, in the Near East (Gupta 2004) and slightly later in China (Zong et al. 2007) may have influenced records of fire within those regions.

4.4. Interpretation of Fire Patterns from 6000 cal yr BP to Present

The middle to late Holocene was a period of changing large-scale controls of fire as summer insolation decreased in the Northern Hemisphere (but increased in the Southern Hemisphere) most glacial ice had disappeared, and sea levels were approaching near-modern position (Figure 8), but seasonal insolation anomalies were still large enough to evoke large regional climate anomalies relative to present. Over the interval the climate system was responding to the transition to modern boundary conditions with a consequent shift in the predominant controls of fire regimes. In addition, increasing human populations may have had a localized role in modifying the fire regimes in certain locations.

Combined climatic and human controls may have shifted vegetation types (and thus fuel type) and disturbance regimes by 6000 cal yr BP. For example, in Australia, a period of maximum precipitation between 7000 to 5000 cal yr BP (Harrison and Dodson 1993) may have been responsible for reduced fire at 6000 cal yr BP. Throughout Indonesia and New Guinea, fire remained less than present, but with increasing spatial heterogeneity. Haberle and Ledru (2001) suggested areas of increased fire were related either to increased variability in El Nino/Southern Oscillation (ENSO) and the related Walker circulation or to the increased role of agricultural activities after 6000 cal yr BP. Black and Mooney (2006) related similar increases to modern ENSO phenomena. Elsewhere, Tinner et al. (1999) report increased fire in the European Alps for the period 7000-5000 cal yr BP that resulted from combined effects of intensified land-use activities and centennial-scale shifts to warmer and drier climatic conditions.

By 3000 cal yr BP, dominant controls of fire regimes were similar to modern. Despite similar-to-present climate, however, fire was greater than present in the mid-latitudes of Eurasia and summer-wet regions of the western United States (Whitlock and Bartlein 2004; Marlon et al. 2006). Progressively decreasing summer insolation in the Northern Hemisphere through the late Holocene led to reduced fire at 3000 cal yr BP compared to 6000 cal yr BP in many regions, but in other northern hemisphere regions weakening of early-mid Holocene monsoons led to greater-than-present fire (Whitlock and Bartlein 2004). Greater heterogeneity in fire patterns in the mid-to-high latitudes of South America at 3000 cal yr BP has been attributed to the onset or strengthening of ENSO and increased human populations (Whitlock et al. 2007). Heavily populated regions of eastern New Guinea, eastern Australia and New Zealand show less-than-present fire, possibly a result of ENSO's greater influence in recent millennia. In contrast, greater-than-present fire across western Eurasia, where Bronze and Iron Age populations used fire as a tool for deforestation, may explain greater-than-present fire during the late Holocene. Many sites in central Asia and central North America show near-modern fire regimes around 3000 cal yr BP.

5.0. Conclusions

Time-slice anomaly maps of fire from the LGM to present illustrate the changing importance of fire as a global phenomenon. These records can be interpreted in terms of changes in biomass burning and imply that climatically-determined changes in fire regimes could have had significant impacts on the global carbon budget through time. The two most important signals shown by the charcoal records, when considered globally, are (a) the monotonic increase in biomass burning between the LGM and present, and (b) the shift from low to high spatial heterogeneity in fire activity ca. 12,000 cal yr BP.

The relatively few charcoal records for the LGM show a consistent pattern of low fire (Figure 4), characterizing the glacial interval from 21,000 through 16,000 cal yr BP. It is possible that the coherency of the records is more apparent than real, and that more spatial heterogeneity will be revealed as more and more highly-resolved charcoal records become available. Nevertheless, the fact that most of the available records show low fire is not surprising given that the climate was globally colder and drier than at present (Braconnot et al., 2007). The cold, dry climate, in combination with lower-than-present

CO_2 levels, would result in an overall reduction in terrestrial biomass (Francois et al. 2000) and thus a decrease in fuel availability. Furthermore, when the troposphere is colder and drier than present there would be less convection, a reduction in lightning activity and thus fewer ignitions. With the waning of the Northern Hemisphere ice sheets, the increase in CO_2 concentrations, and the expansion of the terrestrial vegetation, our charcoal-based reconstructions show that the incidence of fire generally increased towards the present (Figures 4-8).

The charcoal records show an apparent increase in the spatial heterogeneity of the charcoal deposition from the LGM towards the present. Again, this could be an artifact of an increase in the number of records over time. However, examination of the patterns in regions with comparable densities of sites at 15,000 and 12,000 cal yr BP (i.e. northwestern North America, southern South America) suggests that the spatial patterns of charcoal anomalies were indeed more homogeneous in late-glacial times than later. The increased spatial heterogeneity may also reflect the transition away from the glacial state: during the glacial, the overall reduction in biomass was a severe constraint on fire regimes but during the later part of the deglaciation, as temperatures rose, regional responses to climate and climate-induced changes in vegetation cover overwhelmed the global signal and spatial heterogeneity increased. Despite the considerable spatial heterogeneity in fire regimes during the period since ca. 12,000 cal yr BP (at the continental and global scales), there is nevertheless regional coherency at sub-continental and regional scales that appear to be explained by direct climate controls and the indirect effects of climate changes on vegetation cover and fuel loading. The dominant controls on fire regimes are temporally variable and have been changing on millennial timescales since the LGM. For example, we have argued that widespread cool, dry climatic conditions coupled with reduced biomass were important controls regulating fire in the LGM. In contrast, with respect to the Northern Hemisphere, increased seasonality and biomass regulated early Holocene fire regimes whereas decreased seasonality, coupled with increased human activity, were important regulators of fire in the late Holocene.

Our interpretations of the spatial and temporal patterns of change in fire regimes can be regarded as hypotheses. Though plausible, they require rigorous testing at global and regional scales. More data is required to do this in an objective and statistically robust way. This implies a need for continued synthesis of existing data – there are still many records that have not been included in this compilation – but it also requires the collection and analysis of new charcoal records in regions where the nature of the change in climate (and/or vegetation) might have led to different fire behaviour. We anticipate that the maps (Figures 4-6) and regional time series (Figures 7 and 8) presented here will motivate new data collection initiatives.

We have focused predominately on the role of climate rather than human intervention in modulating past fire activity, although studies of individual regions suggest that humans may have played a role, especially during the latter part of the Holocene (e.g., Zong et al., 2007; Clark et al., 1989). There is a general positive relationship between human population and fire incidence during the mid-to-late Holocene (Mouillot and Field 2005, and references therein). For example, frequent fires in parts of Scotland during the middle Holocene have been attributed to human activity (Innes and Blackford, 2003) as well as to the expansion of fire-prone blanket mire vegetation (Froyd 2006). In southern Scandinavia, microscopic charcoal accumulation rates (Berglund et al. 1991) and macroscopic charcoal under and within clearing cairns (Lagerås 2000) were related to forest clearings by humans from 6000 cal yr BP, but especially from 3000 cal yr BP. In the European Alps, fires were intentionally set to disrupt forests and gain open areas for arable and pastoral farming (Tinner et al. 2005). After disruption of forests by fire, controlled burning was used to maintain open areas for agricultural purposes. Similarly, in Central America, late Holocene fire activity has been closely tied to human activity (Horn 2007). An analysis of the role of human activities, in causing and in suppressing fire during recent millennia, requires a better understanding of changes in fire regime and cultural development than is currently available for most regions of the world.

The palaeofire reconstructions presented here offer a unique opportunity to validate models of the coupled behaviour of vegetation and fire (Marlon et al., in prep). Successful simulation of past changes in fire regimes is an integral part of assessing whether we can predict future changes in biomass burning in a realistic way. This, in turn, has implications for maintaining biodiversity, addressing issues of climate change, and assisting governmental agencies in developing appropriate fire management policies. Model-validation exercises necessarily depend on the quality and quantity of palaeodata available (Kohfeld and Harrison, 2000). While extensive, the current version of the charcoal database has marked spatial heterogeneity in sample site distribution. Some regions such as North America contain a relatively high number of sites whereas many Old World regions are generally less well represented. Additional sampling in regions inadequately represented is necessary to ensure that the spatio-temporal coverage of the current charcoal database is sufficient for meaningful data-model comparison.

CHAPTER III

WILDFIRE RESPONSES TO ABRUPT CLIMATE CHANGE IN NORTH AMERICA

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Introduction

It is generally asserted that anthropogenic climate change will lead to widespread and more frequent fires (1, 2). Data from western North America in recent decades are consistent with this; they show that increases in the frequency of wildfire and the duration of the fire season are linked to increased spring and summer temperatures and earlier spring snowmelt (3). Changes in the pattern of precipitation are likewise affecting fire activity (4), as is the development of high fuel loads associated with long-term fire suppression (5). The effects of climate variability on fuels and fire regimes on multiple time scales have received much attention (6-8), and some research has linked shifts in fire regimes at individual sites to rapid climate changes (9). However, the broad-scale response of wildfires to large, abrupt climate changes in the past has received little attention (10, 11). One period of particular interest is the last glacial-interglacial transition (LGIT, 15 to 10 ka), when large and sometimes abrupt (i.e. decades to centuries) changes in climate and biota occurred in many parts of North America. In some regions, environmental changes at the beginning and end of the Younger Dryas chronozone (YDC: 12.9 to 11.7 ka) (12) in particular, were larger than those at any subsequent time (13). Such changes are similar in terms of the magnitude and rate of change to those projected for the future (14-16), and thus provide an opportunity to examine the response of fire regimes to rapidly changing environmental conditions in a variety of settings.

Investigating wildfire activity during the LGIT also allows us to test the recent proposal that a catastrophic extraterrestrial impact event at ~12.9 ka had "continent-wide effects, especially biomass burning" (17). Firestone *et al.* (2007) proposed that a comet exploded over the Laurentide ice sheet, producing a shock wave that would have traveled across North America at hundreds of kilometers per hour, and if multiple large airbursts occurred, could have ignited many thousands of square kilometers. Firestone *et al.* (2007) also hypothesized that the event triggered global cooling, and that extreme wildfires destroyed forests and grasslands and produced charcoal, soot, toxic fumes and ash. These impacts, in turn, ostensibly limited the food supplies of herbivores, contributing to the extinction of North American megafauna and forcing major adaptations of PaleoAmericans (17), although this latter point has been disputed (18).

Even without invoking catastrophic events such as a comet impact, there are still reasons to expect a broad-scale response of fire activity in North America to the abrupt climate changes during the LGIT (19-21). At the beginning of the YDC (12.9 ka), North Atlantic meridional overturning slowed or shut down (21, 22). This led to abrupt cooling in the circum-North Atlantic region and general changes in atmospheric circulation around North America (23-25). Because atmospheric circulation affects temperature, precipitation and the position of storm tracks (26, 27), the particularly abrupt onset of the YDC was registered across the continent. A large, rapid climate reversal occurred in regions adjacent to the North Atlantic, while more distant regions registered changes in the progress of the LGIT (19, 28, 29). Other abrupt climate transitions focused on the North Atlantic, such as the onset of the Bølling-Allerød interval (14.7 ka), or short climatic oscillations, such as the intra-Allerød cold period (IACP, ~ 13.2 ka), may also have had continent-wide impacts on climate.

Large-amplitude, rapid climate change affects fire regimes directly by altering the patterns of ignition and fire weather (30) and indirectly through vegetation composition (19, 31, 32), a major determinant of landscape flammability (33). The nature of the changes in ignition, fire weather and vegetation composition will not be homogenous at a regional scale, but any rapid climate change, whatever its direction, imposes stress on an ecosystem and can trigger some change in the fire regime. Stress would result in increased mortality of the woody vegetation and a build up of fuel, for example, via pest outbreaks or physiological intolerance of new climate extremes. The rate at which such factors affect the fire regime varies, so a broad-scale change in fire activity would not necessarily exhibit absolute synchroneity but some change should still be evident at most sites.

Charcoal and pollen from 35 lake-sediment records across North America (see Supplementary Information [SI] in Appendix B, Fig. 15 and Table 1) were used to assess changes in fire activity (defined here as biomass burned and fire frequency) and woody biomass during the LGIT. Variations in charcoal abundance or influx (particles/cm^2/yr) provide a record of past trends in biomass burning (34-37). Fifteen high-resolution macroscopic charcoal records (i.e., <50 years per sample and particles >100 μm) were further analyzed to reconstruct past fire episodes (defined as one or more fires occurring during the time spanned by a charcoal peak) (36, 38) and charcoal peak magnitude, an assumed metric of fire size, severity, or proximity (39) (SI Methods). The proportion of arboreal pollen (AP) in the lake sediments, which reflects the abundance of tree and shrub taxa on the landscape, was used to estimate the levels of woody biomass in the vegetation at the sites. AP can overestimate tree cover and mask shifts in trees and shrubs (40), so we consider it only a general indicator of available woody fuel levels. Records of charcoal influx, peak frequency, and AP were used to document trends in biomass burning (35, 36), fire-episode frequency (hereafter fire frequency) and woody fuel levels. These trends were compared with ice-core records of CO_2 (41) and $\delta^{18}O$ (21), the latter clearly illustrating abrupt climate changes, in order to explain the broad-scale changes in fire activity.

Results & Discussion

Trends in Fire Regimes and Woody Fuels

The general trend of charcoal influx across all sites (as represented by a three-segment linear regression, Fig. 9c) indicates a significant ($p < 0.01$) increase in biomass burning until the beginning of the YDC, no overall change during the YDC, and then a further increase in biomass burning thereafter ($p < 0.01$). A local regression curve, which does not assume a specific form for the trend, displays a similar pattern. The bootstrap confidence intervals around charcoal influx indicate these trends are not induced by any particular record. Inspection of the records (Fig. 10; SI Fig. 16), however, shows that there can be different responses at individual sites reflecting modulation of the regional-scale response by local factors. For example, while sites 4-9 in southern British Columbia (BC) all show increasing biomass burning from 15 to 10 ka, spatial patterns are complex in the Pacific Northwest, Sierra Nevada and Northern U.S. Rocky Mountains (NRM). The three sites in Alaska (AK) show increasing burning during the Bølling-Allerød and stable levels during the YDC, but trends are variable after the YDC. Almost no spatial coherence is evident in the Southwest, Midwest, and East, although these regions have limited data. Thus, while the composite record strongly indicates broad-scale trends in biomass burning, heterogeneity is expected and apparent at local to regional scales.

The overall trend in fire frequency increases during the Bølling-Allerød (Fig. 9d, SI Fig. 17) and has no discernable trend thereafter. Some regions show coherent patterns in fire frequency, including AK (sites 1 and 2), the Pacific Northwest (sites 11, 13, and 14), and the NRM (sites 21-23, and 25) (SI Fig. 17), although the nature of the changes naturally differ between regions. Fire frequency is most variable after 11.7 ka; only sites 21 and 29 show little or no change after that time. In general, peaks in fire frequency tend to match local maxima in biomass burning (e.g. at 13.9, 13.1, 12.3, and 11.7 ka).

There are no empirical studies that link the absolute size of charcoal peaks to a specific fire characteristic, such as area burned or severity, so the peak magnitudes must be interpreted with caution (SI Fig. 17). However, in previous research, unusually large peaks have been linked to extreme fire years in the historical record when large areas burned at the regional scale (42, 43). For example, fires in 1910 that burned over 400,000 ha in the NRM comprised the largest peak of the last 120 years at site 20 (42). Consequently, peak-magnitude data suggest that many large fire episodes occurred between 15 and 10 ka, and large or severe fire episodes were more likely after the end of the YDC than before it, as for example in the Pacific Northwest (sites 11-13), the NRM (sites 20, 23-25), and the Southwest (site 27) (Fig. 10). Fire frequency was also high at most of these sites after the YDC.

The woody biomass trend increases during the Bølling-Allerød, is stable during the YDC, and decreases thereafter (Fig. 9e). Trends at individual sites again vary regionally and with elevation (Fig. 10, SI Fig. 16). Woody biomass declines at most sites in BC and increases in the Sierra Nevada, Southwest and Northeast. Other regions show mixed patterns. Fire-fuel relationships among sites also show regional similarities. For example, trends in charcoal influx and AP are similar at mid- to high-elevation sites in the Pacific Northwest and NRM (sites 13, 15, 23, 24, 25), where biomass burning and woody fuel levels generally increased together as open forests became more closed or alpine vegetation was replaced by parkland and then forest during the LGIT (8). In BC (i.e. at sites 5, 6, 7, 8, and 10), an inverse relationship in fire and fuels is apparent because biomass burning increased as closed mixed conifer forests were replaced by more open forests (44). Charcoal influx is often opposite to AP in the Midwest as well, where grass abundance (low woody biomass) is a good predictor of biomass burning (45). Important changes in woody fuel levels in AK are obscured in the AP trends, because AP does not show changes in the relative importance of shrubs versus trees. AP declines at site 3 at 11.0 ka, for example, despite a large increase in *Populus* at that time. Overall, the spatiotemporal variability in woody fuel levels and biomass burning makes it difficult to

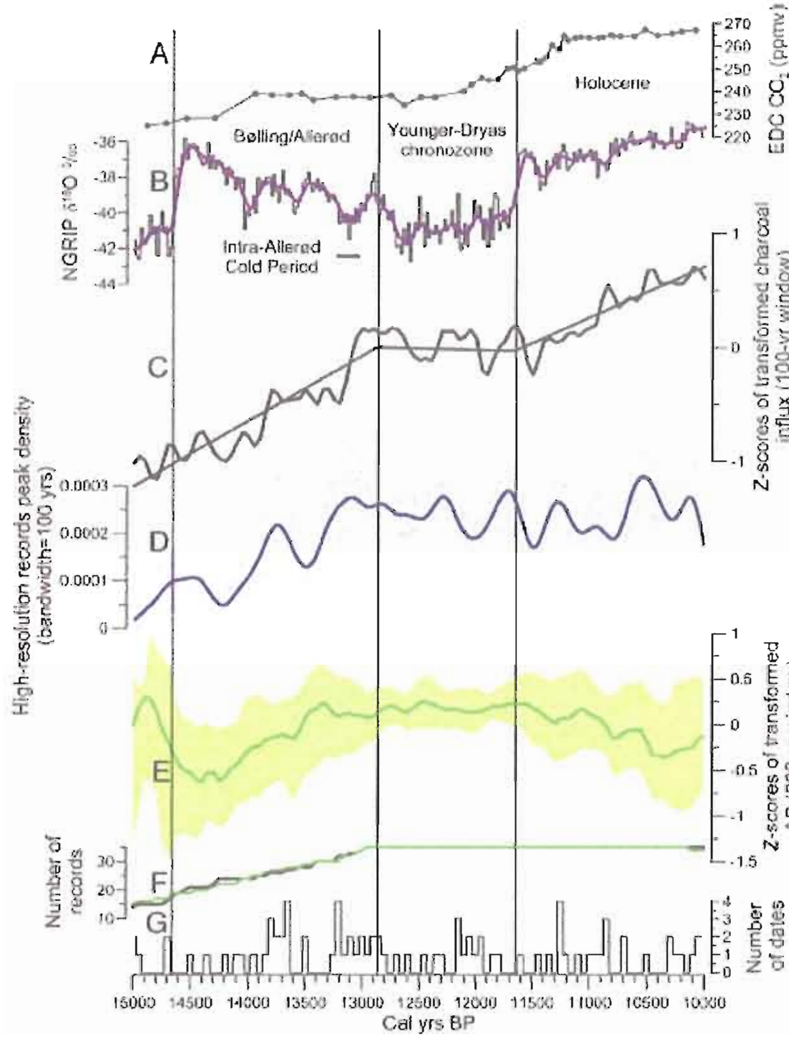

Figure 9. Reconstructions of biomass burned, fire frequency, and woody biomass levels in North America. (*A*) The CO2 ice-core record from Antarctica (41). (*B*) The NGRIP $\delta^{18}O$ record, a proxy for North Atlantic temperatures (21). (*C*) Reconstruction of biomass burned based on 35 records; the straight lines are segmented regression curves, and the smooth curves are local-regression fitted values. (*D*) Reconstruction of fire frequency based on 15 high-resolution records, expressed as the density of peaks per site-year. (*E*) Trends in woody biomass based on 35 records. (*F*) Number of records contributing to the biomass burning (black) and woody biomass (green) trends. (*G*) Number of dates per 50-year interval in the 35 paleo records. Confidence intervals (95%) are based on bootstrap resampling of sites. Vertical lines mark the beginning (~12.9 ka) and ending (~11.7 ka) of the YDC.

generalize about fire-climate-vegetation linkages at the continental scale, but the role of climate in determining both woody fuel levels and fire activity underpins the regional coherence in charcoal-AP relationships. The AP data do indicate that availability of woody fuels was not a limiting factor in determining levels of biomass burning at the beginning or end of the YDC.

Evidence for Continent-Wide Wildfires at 12.9 ka

Firestone et al. (17) hypothesized that a comet impact at 12.9 ka ±50 yrs triggered continental-scale wildfires across NA. One specific example has been proposed by Kennett et al. (2008) (46). However, the well-documented rapid climate changes of this time alone may have triggered increased fire at a regional scale. To separate these effects, we compared the response of fire during intervals of rapid climate changes at the beginning and at the end of the YDC. Fire-episode events that occurred during the transitions into and out of the YDC were identified in both the high- and low-resolution records (see Methods) to determine whether fire episodes, regardless of magnitude, were more likely to occur (within ±50 yrs) at 12.9 ka than at 11.7 ka (Figs. 9a, and 10). Due to high uncertainties in radiocarbon during the YDC, both 100-yr and 500-yr window widths were used to identify fire episodes (Fig. 10). Using a 100-yr window, 14 sites across the continent (Fig. 10) showed a peak (or increasing charcoal if no sample was within the window) at 12.9 ka. The peak was large (i.e., above the 90th percentile based on quantile regression) in the nine low-resolution records, but it was not present in any of the five high-resolution records (SI Fig. 17), suggesting that the relatively high magnitude of fires at 12.9 in the low-resolution sites may be an artifact of the small number of samples in these records. The data also indicate that fires were twice as likely to occur at 11.7 ka, the abrupt *end* of the YDC, than at 12.9 ka (Fig. 10; SI Figs. 15 and 16). Using a large 500-yr window-width greatly increased the number of sites recording fires around 12.9 ka; however it also increased the number of fire episodes recorded at 11.7 ka (Fig.

10). It could be argued that poor dating control on some of the records prevented identification of fire episodes at 12.9 ka, however, when we limited our analysis to the 14 records with dates within ±300 years of 12.9 ka (Fig. 10), the results did not change. Peaks in charcoal influx were registered throughout the LGIT, particularly associated with abrupt climate changes, but there was no evidence of continent-wide wildfires at the beginning of the YDC.

Potential Controls on Fire Regimes and Woody Fuel Levels During the LGIT

The broad-scale trends in biomass burning, fire frequency and magnitude, and woody fuels during deglaciation are consistent with climate changes documented by ice cores, marine and lake sediments, speleothem and other records from North America (21, 28, 47, 48). During the Bølling-Allerød, woody biomass, biomass burning and fire frequency all increased (Fig. 9e), a likely consequence of warming and increased tree cover (40). A stepped increase in biomass burning is evident at 13.9 ka, coincident with a short period of warming and is matched by a peak in fire frequency.

A particularly steep increase in charcoal influx occurred at 13.2 ka (Fig. 9c); this is the largest and most rapid change in biomass burning during deglaciation. Burning was widespread but not continent wide (see site details in Supp). Furthermore, the change in fire regime is not unique: several sites show similar peaks prior to the onset of the YDC and many show an even larger peak at the end of the YDC. The widespread increase in fire activity (i.e. charcoal influx and peak frequency) at 13.2 ka appears about 300 years before the hypothesized comet impact (17). Of the sites that do show fire activity at 13.2 ka, many are from regions distant from the proposed locus of the impact area over the Laurentide ice sheet, as well as from the proximal influence of the ice sheet on regional climates (e.g., in AK, the Southwest, Pacific Northwest, and the NRM). The timing and distribution of fire activity at 13.2 ka is consistent with the IACP – an abrupt short-term climate reversal recorded in the GISP $\delta^{18}O$ ice-core data (Fig. 9b). The IACP is associated with a rapid oscillation in North Atlantic temperatures that may have affected atmospheric circulation patterns across the continent (21, 23, 49) and increased the likelihood of drought as well as severe frost damage on some tree species (50). Any increase in vegetation mortality associated with such events would have added to the available fuels and facilitated an increase in fire.

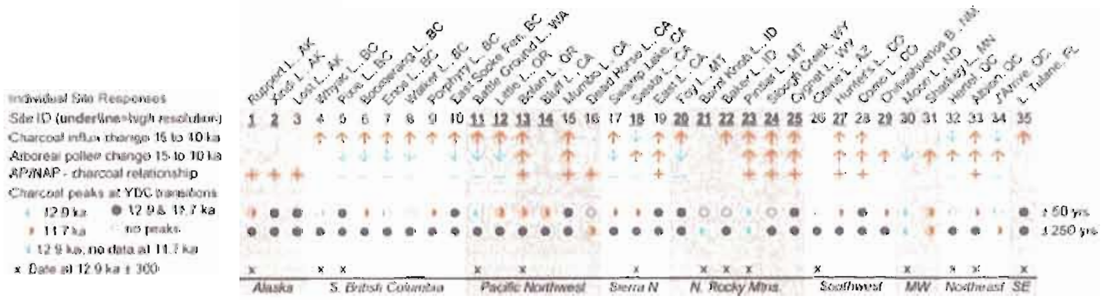

Figure 10. Site summaries of changes in charcoal influx, arboreal pollen (AP), and charcoal–pollen relationships during the LGIT and of charcoal peaks at the beginning and end of the YDC. High-resolution site numbers are in bold type and underlined. Regions are identified by alternate shading. A ↑ (↓) indicates a general upward (downward) trend in charcoal influx. A + (–) indicates a positive (negative) relationship between charcoal and AP. Records that had a radiocarbon or tephra date within 300 years of 12.9 ka are marked by an x in the bottom row.

During the YDC, ice-core $\delta^{18}O$ data indicate cool and variable temperatures in the North Atlantic region. Cooling is also evident in parts of western North America based on pollen and speleothem records (25, 28, 49), but climate patterns likely varied across the continent (27). The composite records (Fig. 9) show that biomass burning was higher but more variable than pre-13.2 ka. Fire frequency and biomass burning had local maxima at ~12.3 ka and at the end of the YDC (11.7 ka). Although there are fundamental and widespread changes in vegetation at the beginning (and end) of the YDC (19), the woody biomass trend shows little change during the YDC. This lack of change does not preclude change in specific regions e.g. Alaska (48) or at individual sites.

Biomass burning and fire frequency both decline at 11.7 ka but increase thereafter. Woody biomass, however, decreases from 11.7 to 10.0 ka. This contrast in behavior marks a shift in the relationship between fire and vegetation: prior to 11.7 ka woody biomass and fire activity generally change in parallel, post 11.7 ka they change in opposite directions. Early-Holocene warming and enhanced seasonality facilitated the emergence of new vegetation communities and disturbance patterns (19, 32, 51). Low-elevation sites in the western US show the biggest changes, with declining woody biomass as forests became more open (44, 52) and more likely to burn (SI Fig. 16 and Fig. 10). High-elevation sites in the Pacific Northwest and NRM also show increasing fire activity, but in association with increasing rather than decreasing woody fuel levels. New fire-fuel patterns also evolved in the Northeast after the YDC, with declines in biomass burning associated with increases in woody biomass.

Factors other than climate may have contributed to observed changes in fire regimes during the LGIT, including changes in atmospheric CO_2, the arrival of Clovis people between ~13.4 and 12.8 ka (53), and the extinction of herbivorous megafauna (54). Changes in CO_2 affect vegetation productivity (55) and potentially fuel load. Atmospheric CO_2 increased in step-wise fashion from the Last Glacial Maximum to the beginning of the Holocene (56) (Fig. 9a). The changes in woody biomass, fire frequency and biomass burning are not coincident with changes in CO_2, although increasing CO_2 may have contributed to woody biomass production during the early part of the Bølling-Allerød. Clovis people appeared in North America between 13.4 – 12.8 ka, broadly coincident with the sharp increase in biomass burning at 13.2 ka, and then rapidly spread out across the continent (18). Paleoindians may have increased fire activity directly by setting more fires (57) or indirectly by reducing megafaunal populations. The decline in megafaunal populations, in turn, could have increased fuel loads and changed soil moisture regimes, both of which could have promoted fire (58, 59).

The 13.2 ka fire peak is registered at sites widely dispersed across the continent; it is not consistent with the progressive colonization of North America by Paleoindians. It also seems unlikely that people (or megafauna) would have caused an increase in burning across the full range of elevations represented by the sites, and particularly at high-elevation sites (the fire peak is evident at over five sites above 2000 m; see SI text and Table 1). Furthermore, most fire records show a discrete peak rather than a permanent regime change as might be expected if humans or megafauna exerted a major control on fire regimes. It is possible, however, that the arrival of people and/or the extirpation of megafauna (18, 53, 54) played a role in permanently altering fire regimes at the sites that show a fundamental fire-regime shift at 13.2 ka. After 13.2 ka, fire-regime changes are not coincident with periods of increase in human populations. Thus, the spatial and temporal distribution of the fire signal point towards climate as the primary cause of increased fire activity at 13.2 ka.

In summary, fire records from North America show stepped increases in biomass burning during the LGIT. Abrupt climate changes are generally marked by a shift in the level of burning as well as an increase in the incidence of fires. No continent-wide fire response is observed at the beginning of the Younger Dryas chronozone, the time of the hypothesized comet impact. The results provide no evidence of synchronous continent-wide biomass burning at any time during the LGIT. The data indicate variability in the direction of changes in fire regimes among paleofire records, which may be due in part to noise and local variability (60) but is more likely a result of spatially complex climate controls and/or vegetation changes. Although there is broad congruence between changes in climate, fire and human populations at the beginning of the YDC, we find no convincing evidence that the observed changes in fire activity were caused solely by changes in human or herbivorous megafauna populations.

Methods

We used 30 lake-sediment records in North America from the Global Charcoal Database (GCD v. 1[1,2]) and five records from authors that 1) were recording fire activity

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[1] http://www.ncdc.noaa.gov/paleo/impd/gcd.html

before, during and after the YDC; 2) had at least five data points and one date (radiocarbon or tephra) from 10 to 15 ka; and 3) had pollen data from the same site. We did not include charcoal data from records that only sampled the beginning of the YDC because there is no baseline for analyzing changes in the fire regime with such data (46). We also excluded marine charcoal data (61) because there is no evidence that charcoal influx and peaks in influx in such records reflect recent fire activity from a consistent source region. Pollen data were obtained from authors or from the North American Pollen Database[3] (SI Table 1). We examined the chronologies for each record to ensure that the age-depth relationships were generally consistent throughout the LGIT and that no age reversals occurred during that interval. Under such conditions, age controls in lake-sediment records are sufficient to describe centennial-scale variations (see SI text).

For all analyses, charcoal concentration data (particles cm^{-3}) were converted to influx values (particles $cm^{-2}\ yr^{-1}$) (see SI text). For the low-resolution records, millennial-scale (background) trends were identified by smoothing the data using quantile regression (62). Any increase in charcoal influx above background within a defined interval (i.e., either ±50 or ±250 years) was considered a peak. High-resolution records were smoothed using a decomposition technique (63) that separates peaks from background charcoal and allows the reconstruction of peak magnitude and fire frequency. Arboreal pollen proportions were obtained by dividing the sum of arboreal and shrub pollen percentages (AP) by the sum of the total terrestrial pollen percentages (AP/(AP+NAP)).

To display the general trends in the charcoal influx, the data were transformed to stabilize the variance and standardized to facilitate comparisons across a range of charcoal influx levels (37). To assess the significance in the trend, we fit a segmented linear regression model to these data, with breakpoints at the beginning and end of the

[2] http://www.bridge.bris.ac.uk/projects/QUEST_IGBP_Global_Palaeofire_WG/

[3] http://www.ncdc.noaa.gov/paleo/napd.html

YDC (see SI). We also summarized the data by using "lowess" or local regression curves. Confidence intervals for the local regression curves were generated by a bootstrap approach in which individual records (not samples) were sampled with replacement over 1000 replications. The approach reveals the sensitivity of the trends to the particular selection of charcoal and pollen records used here. Pollen data were also transformed (64) and summarized using local regression curves. The peak frequency trends in the high-resolution records were summarized by a local-density (kernel smoothing) procedure.

CHAPTER IV

CLIMATE AND HUMAN INFLUENCES ON GLOBAL BIOMASS BURNING OVER THE PAST TWO MILLENNIA

This chapter has been published as a co-authored manuscript in the journal *Nature Geoscience* (Marlon et al. 2008).

Introduction

Fire is a key Earth System process affecting ecosystems, land-surface properties, the carbon cycle, atmospheric chemistry, aerosols, and human activities. Humans manage fire intensively today, so it is easy to forget that fire is a natural process that has dominated the ecology of many terrestrial ecosystems throughout their history[1-3]. Empirical data on long-term changes in fire activity, particularly at broad spatial scales, however, is limited. Historical records[4], remotely-sensed data[5], and tree-ring data from the last few centuries[6-8] provide the vast majority of information about the interactions of fire, climate, vegetation and people. Climate-change projections indicate that we will be moving quickly out of the range of the natural variability of the last few centuries. Charcoal records from lake sediments allow us to infer the impacts of climate changes and human activities on global biomass burning during periods when both have changed substantially. Although hundreds of such records have been developed during paleoecological analyses[9, 10], until now no attempt has been made to analyze them for large-scale patterns and trends over the past 2000 years.

We present global and regional reconstructions of biomass burned over the past 2000 years (Figs. 11 and 12) based on a global sedimentary charcoal dataset (Fig. 13). We interpret temporal patterns in biomass burned, as indicated by changes in the input of charcoal to sediments, by comparison with independent reconstructions of human

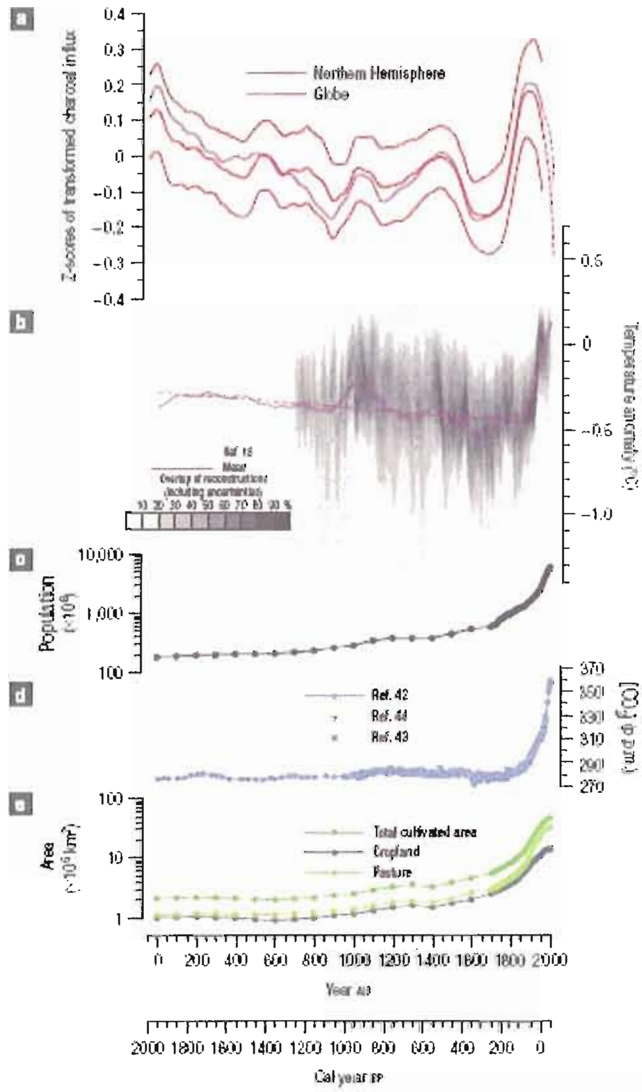

Fig. 11. Reconstructions of biomass burning, climate, population and land cover. **(a)** Reconstruction of global (red line) and Northern Hemisphere (NH) (purple) biomass burning with confidence intervals based on bootstrap resampling by site. A dashed line is used to represent increased uncertainty in late-20th century changes in biomass burning. **(b)** Reconstructions of NH climate[18] with mean values (purple line) of available reconstructions, trend line (dotted) for first part of record, and overlap of uncertainty ranges of ten NH temperature reconstructions after 700 A.D. (gray shading). **(c)** World population from the HYDE 3.0 database[27]. **(d)** Atmospheric CO_2 concentration[41-43]. **(e)** Global agricultural land cover[27].

population and temperature changes, and with climate simulations that mimic the broad features of reconstructed temperature changes.

Charcoal Records of Biomass Burning

Charcoal accumulation in sediments has been shown[11, 12] to reflect biomass burning within tens of kilometres of the sampling site. We developed regional and global composite stratigraphies based on 406 charcoal records from lake sediments and peats. Although there are geographic gaps, there is good coverage of climatic zones and of all the major biomes except grassland/dry shrubland (Fig. 13; Fig. 18), where low woody biomass limits charcoal production. Composite records were standardized and transformed (see Appendix C) to represent centennial-scale trends in charcoal accumulation rates or “influx” (in units of quantity area^{-1} yr^{-1}) for the globe or a given region and to reveal the relative changes in biomass burned through time[11]. Interannual to decadal variations are not resolved so these data do not record changes that might be attributable to higher-frequency climate variability or land management changes over the past few decades[5, 6]. The strength of these data lies in their ability to provide a long-term observational record with site coverage that does not degrade substantially with increasing time from the present.

The global sedimentary charcoal record (Fig. 11a) shows a long-term decline in biomass burning from 1 to ca 1750 A.D., followed by a marked increase. A maximum, corresponding to the highest biomass burning rate in the past two millennia, occurs around 1870 A.D. This maximum is followed by a sharp downturn in the composite record. At centennial scales, there is a local maximum in biomass burning at 1 A.D. that does not correspond to warmer temperatures, but the temperature reconstruction consists of only two records during this period. Biomass burning minima occur ca 400 and 900 A.D., and at 1700 A.D., during the “Little Ice Age” (ca 1400-1800 A.D.). Our analyses (see Supplementary Information) show that these patterns in the global record are robust and unaffected by differences in record type or sampling resolution, changes in sedimentation rates and age model construction, or by the choice of statistical techniques used to construct the composite records.

The long-term decline in biomass burning prior to 1750 A.D. is most strongly expressed in the northern extratropics (Fig. 12), particularly in western North America and Asia (Figs. 28 and 31). The long-term decline is also characteristic of records from Central and tropical South America (Fig. 14), although the decrease in biomass burning there occurs earlier (~1300 A.D.) and persists longer than in other regions. In contrast, the tropics and southern extratropics show variable levels of biomass burning to ~1850 A.D., after which the records show a widespread decline in biomass burning (Figs. 25 and 26).

Increased burning in the tropics and western U.S. during the last three decades has been widely reported[5, 6, 13, 14]. The global charcoal record does not reflect this for a variety of reasons, including data coverage, methodological issues and chronological constraints in core-top sediments. For example, many of the charcoal records from the tropics do not span the most recent few decades. In addition, the geographic coverage of lakes and bogs in the tropics does not overlap well with the areas of most severe burning (Fig. 13). However, individual paleofire records near areas of recent severe burning do show increasing charcoal levels towards present. For example, the only site in Kalimantan[15], Indonesia, where extensive peat fires occurred, indicates high biomass burning in the last few decades. This is true in the western U.S. as well[16, 17]. Additional data and regional syntheses are needed to resolve these patterns.

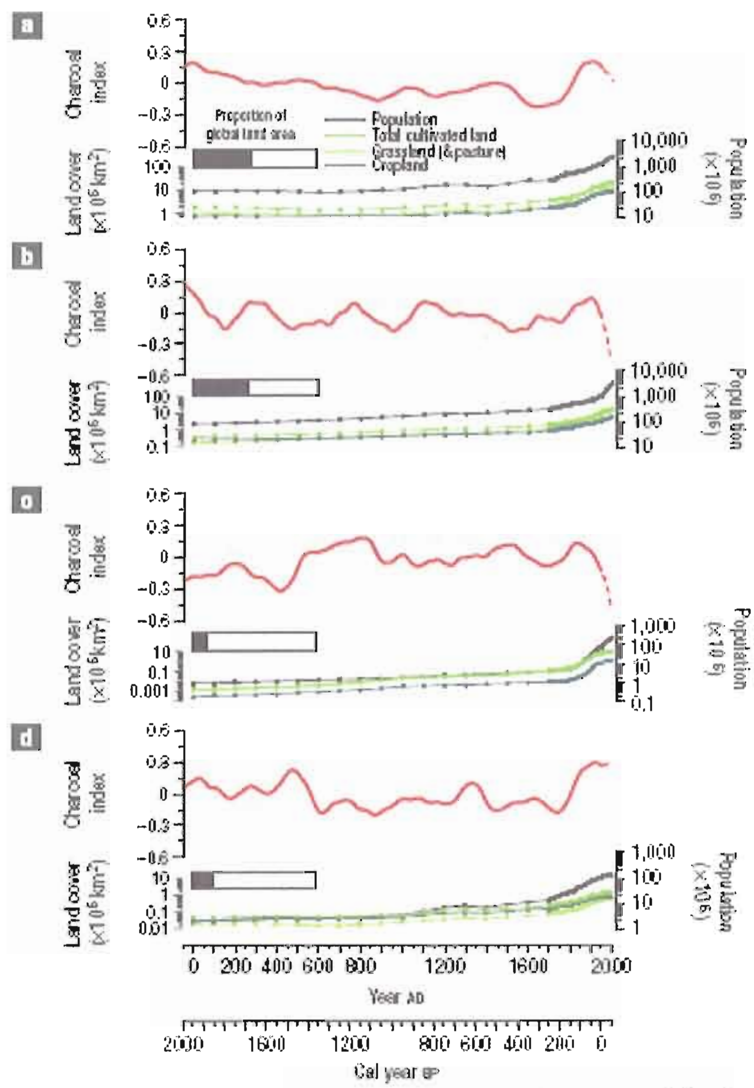

Fig. 12. Zonal changes in biomass burning, population and land cover. Changes in biomass burning (as in Fig. 1) with land-cover change[27], and proportion of global ice-free land area, for **(a)** northern extratropics (> 30°N), **(b)** tropics (30°N to 20°S), **(c)** southern extra-tropics (> 20°S), and **(d)** the northern high latitudes (> 55°N). The composite records are based on at least 10 sites per region and thus should reveal the dominant patterns that reflect processes operating at large spatial scales (see Methods).

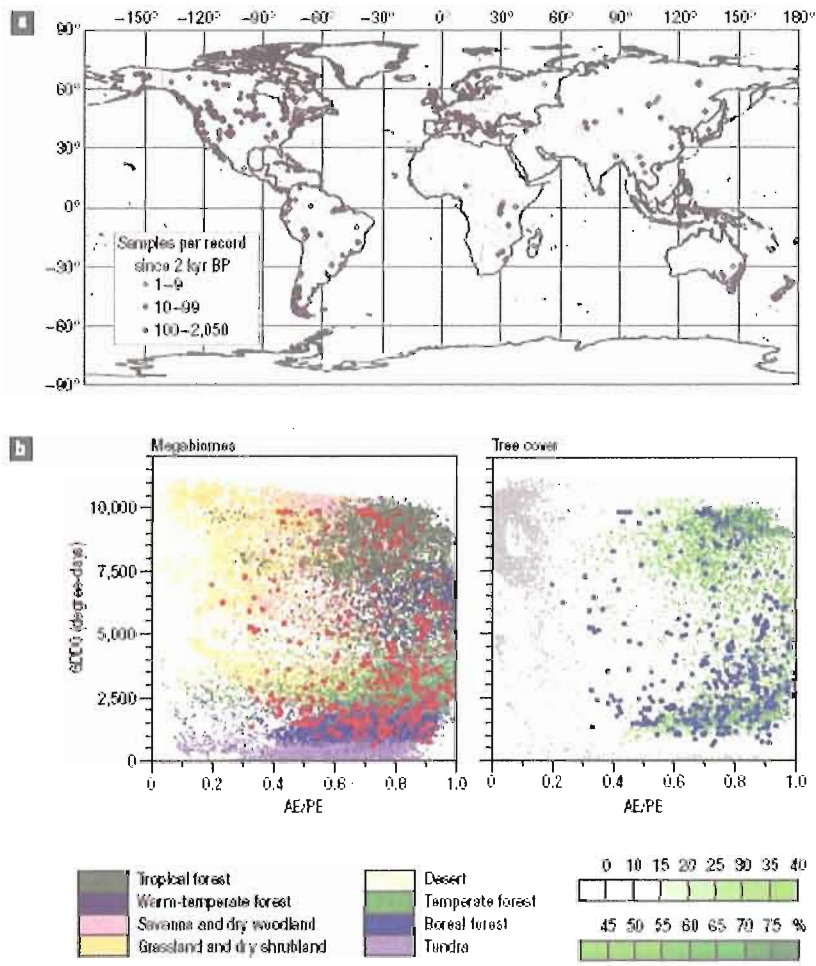

Fig. 13. Distribution of sites in geographic, climate and vegetation space. (a) Locations of charcoal records and number of samples over the past 2000 years (see Methods). Mean sampling density is one sample per 70 years. (b) Distribution of records in bioclimate space showing locations plotted against growing degree days above 0°C (an index of effective warmth during the growing season) and the ratio of actual to equilibrium evapotranspiration (an index of effective moisture)44, 45, with remotely-sensed tree cover46 and modeled biome47 shown for comparison.

Climate, Population, and Biomass Burning

Comparisons of global biomass burning with reconstructed and simulated climate conditions during the past two millennia show strong similarities until the Industrial

Revolution, ~1750 A.D. (Fig. 11b, Fig. 22). Reconstructions of Northern Hemisphere (NH) mean annual temperature from paleoclimatic evidence[18] show a gradual cooling trend from 1 A.D. to 1750 A.D. Simulated temperatures show a similar decline[19] (Fig. 22). Global and NH reconstructed temperatures show local maxima around 1000 and 1400 A.D., corresponding to the local maxima in charcoal influx, and a pronounced decline from 1400 to 1700 A.D. (Fig. 11b).

Over the interval from 1 to 1750 A.D. world population and land-cover conversion to agriculture generally increased (Fig 11c) with a very slight century-scale decrease related to the Black Death in Europe (ca. 1300 to 1400 A.D.). Population growth and accompanying land-cover change therefore do not account for the apparent decrease in global biomass burning during the first phase of the composite record. Conversely, the correspondence between declining biomass burning and long-term cooling on hemispheric or global scales can be explained by the positive temperature-dependence of effective moisture and vegetation productivity (and hence fuel availability) in cooler climates.

Biomass burning sharply increased around 1750 A.D.; temperatures (Fig. 11b, Fig. 22), greenhouse gas concentrations (Fig. 11d) and the rates of land-cover conversion and population growth (Figs. 11c and 11e) also began to increase. Increasing temperatures, rapidly increasing population and land-cover conversion, and rising CO_2 (promoting an increase in biomass through CO_2 fertilization) could in principle all have contributed to the biomass burning increase. However, after ~1870 A.D., a sharp downturn in biomass burning occurs despite accelerated increases in temperature and CO_2, strongly suggesting the involvement of human activity. The decades following 1870 A.D. coincide with the period of maximum expansion of population and agriculture (Figs. 11c, 11e, and 12), so the biomass burning decrease beginning then certainly cannot be explained by reduced human activity. Indeed, this has been the period of most rapid land-use change, characterized by large-scale conversion of native vegetation to croplands and the widespread introduction of domestic grazing animals such as cows and sheep[20-22]. We therefore suggest that the downturn can plausibly be explained as an effect of land-use change, resulting in landscape fragmentation and a generally less flammable landscape in many regions[23]. Active fire suppression since the early 20th century has also presumably reduced total biomass burning in recent decades.

Records from the northern high-latitude regions (poleward of 55°N) show the general climate-induced decline and the post-1750 A.D. increase in biomass burning characteristic of the global signal but do not show the recent downturn in biomass burning (Fig. 12d, Fig. 27). This region has been much less affected by agricultural expansion than regions farther south (Fig. 12), and is influenced by high-latitude amplification of the global warming signal leading to increased temperatures, dryness, and greater fire activity[24, 25]. The biomass burning record from Europe (Fig. 30), which has been subject to agricultural exploitation throughout the past two millennia, also fails to show the recent downturn, consistent with our explanation. The regions in which the downturn is most strongly expressed, including western North America (Fig. 28), the tropics (Fig. 12b and Fig. 25), and Asia (Fig. 31), are characterized by strong intensification of land management since the mid-19th century. Although initial colonization may have been marked by an increased use of fire for land clearance, the expansion of intensive agriculture and forest management activities in these regions was associated with a reduced incidence of fire[23].

The record of global biomass burning over the past 2000 years can be divided into four distinct intervals: 1) 1 to 1750 A.D., when climate drove the long-term downward trend in biomass burning; 2) 1750 to late 19th century, when population-driven land-cover changes along with increases in global temperatures produced the sharp increase in biomass burning, 3) late 19th to mid-to-late 20th century which includes the striking decrease in biomass burning that accompanies the decrease in the rate of land-cover change; and 4) the last several decades, which is beyond the resolving power of the global charcoal record, when increased biomass burning is reported in many tropical regions.

Fire and Atmospheric Chemistry

Ferretti *et al.* (2005)[26] showed an "unexpected" trend towards more negative $\delta^{13}CH_4$ values from 1 to 1700 A.D. in the Law Dome ice-core record, and invoked changes in biomass burning to explain this trend. They noted the similarity of their $\delta^{13}CH_4$ record both to the Law Dome record of atmospheric CO concentration and to proxy-based reconstructions of NH temperatures. Because fire is a major source of CO and the main natural source of relatively ^{13}C-enriched CH_4, Ferretti *et al.* postulated a declining trend in biomass burning, and attributed it to long-term cooling of the land surface. Houweling *et al.* (2008)[27] expressed skepticism about this possibility and proposed an alternative scenario involving changing wetland emissions, plant emissions, and early rice cultivation to explain the $\delta^{13}CH_4$ trend, avoiding Ferritti et al.'s implication of high biomass burning levels in pre-industrial time. Our results however provide strong empirical support for the decline in biomass burning since 1 A.D. proposed by Ferretti *et al.*[26].

Ferretti *et al.* also noted an especially steep fall in $\delta^{13}CH_4$ from 1500 to 1700 A.D., and invoked population decline in the Americas as an additional factor to account for it. The global charcoal record also shows a sharp decrease from 1500 to 1700 A.D., but global population did not fall during this period (Fig. 11c). South American population declined after 1500 A.D.[28, 29] (Fig. 14), but a comparison of biomass burning, population estimates for the Americas[27], and climate data from Andean ice cores[30] and Cariaco Basin titanium concentrations[31], which is a proxy for Atlantic Intertropical Convergence Zone latitude and hence tropical precipitation trends, suggests that the largest decrease in biomass burning there preceded the population decline and followed regional climate trends (Fig. 14). The similarities between the global charcoal record and climate proxies suggest a continuing climatic control through at least 1750 A.D.

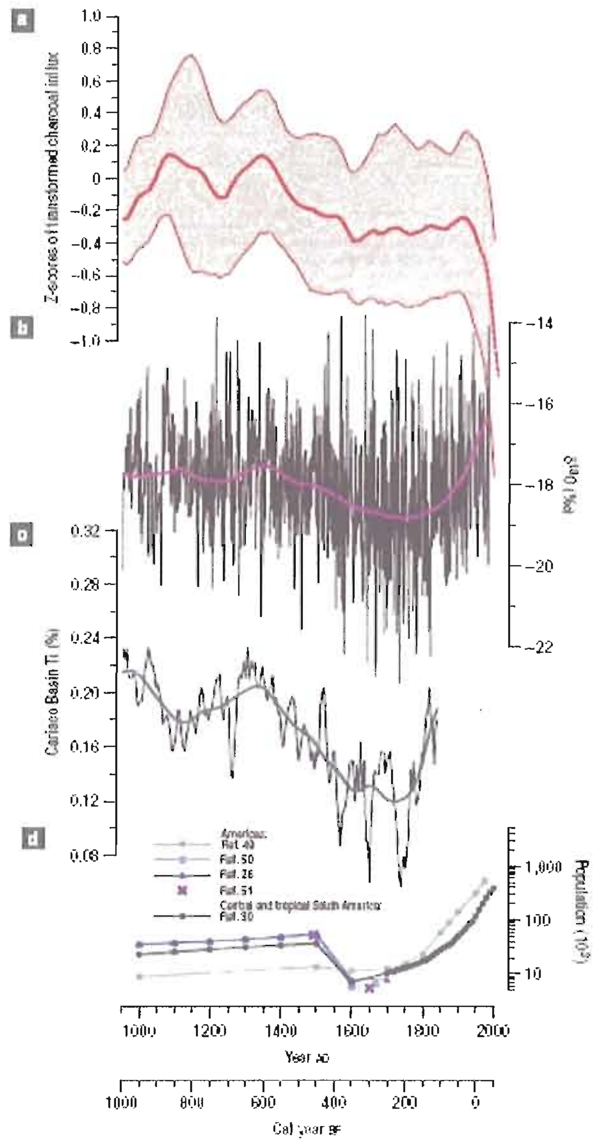

Fig. 14. Comparison of biomass burning reconstruction for Central and tropical South America with climate and population data. Biomass burning reconstruction for Central and tropical South America (20°S to 30° N) for the past 1000 years **(a)**, with oxygen isotope data from the Quelccaya ice core[30] **(b)**, and titanium concentration from a Cariaco Basin marine core[31] **(c)**, both smoothed as the charcoal records. **(d)** Population estimates for the Americas derived from archaeological records[26, 27, 48, 49, 51]. The two crosses are estimates of the population of the Americas before and after European contact from Denevan (1992)[50].

The peak in biomass burning at ~1870 A.D. corresponds to a peak in black carbon (BC) around 1900 A.D. observed in Greenland ice, however, McConnell *et al.*[32]attribute this to a peak in industrial emissions. The separation of biomass and fossil components of the BC record, however, rests on the interpretation of the accompanying vanillic acid (VA) record as a proxy for biomass burning[33] and, in part, on the choice of scaling factors between VA and BC.

Our results strongly suggest that climate change has been the main driver of global biomass burning for the past two millennia. The decline in biomass burning after 1870 A.D. is opposite to the expected effect of rising CO_2 and rapid warming, but contemporaneous with an unprecedentedly high rate of population increase. This suggests that, during the industrial era, a major impact of human activities has been to reduce biomass burning through the large-scale expansion of intensive grazing and cropping with associated landscape fragmentation, and also active fire management. However, in regions less dominated by intensive land management, wildland fire has remained at high levels or increased as the climate has warmed. In the future, it is plausible that the main impact of human activity on burning will be through anthropogenic climate change[34]. To assess this possibility, it will be necessary to improve our understanding of paleoecological records of fire and to develop reliable models of the fire regime. Such models should include human as well as natural ignitions, account for the effects of land use as well as climate, and be tested using paleoecological and ice-core data[35].

Methods

Analyses are based on 406 sedimentary charcoal records from the Global Charcoal Database (GCD version 1)[10], plus supplementary data from members of the Global Palaeofire Working Group that will be incorporated into GCD version 2. Only records from lakes, bogs and small hollows were included in our analyses (see the Supplementary Information).

We examined the distribution of data in terms of geographic, climatic and vegetation space to ensure that the dataset could be considered a reasonable representation of global biomass burning. Given the unequal distribution of land between the northern and southern hemispheres, the charcoal data set is reasonably representative of the globe. Sixty-seven percent of the sites are in the northern hemisphere, which represents ~74% of the ice-free land area. Nevertheless, some geographic regions are less well sampled than others (Fig. 13). To assess the global coverage of the GCD in terms of climate and vegetation, we plotted individual charcoal sites on top of climate, simulated biomes and tree cover data (Fig. 13). The charcoal sites provide a reasonably representative sampling of all the major climatic zones, and of forest biomes (Fig. 18). Grassland/dry shrubland, savannas and deserts are under-represented (Fig. 13). There are few natural fires in desert regions, thus this under-sampling is not important. The paucity of charcoal records from grassland and dry shrubland, and from savanna reflects the dominance of low-intensity ground fires in these regions and the smaller number of lakes present in dry regions. Such low-intensity fires do not consistently leave a charcoal record (although see ref. [36]), and the total biomass burned is low compared to forests[37]. Thus, the under-sampling of these two vegetation types is unlikely to affect the composite global biomass-burning curve significantly.

The broad range of data types and charcoal quantification methods, as well as the skewed distribution of most charcoal records, motivated the transformation and standardization of the data prior to generating composite records. After testing the impact of several standardization and normalization techniques, we chose to use a modified version (see the Supplementary Information) of the normalization and standardization procedure described in Power et al. (2007), which resulted in highly robust charcoal summarizations.

Charcoal data were normalized to stabilize the variance and standardized to make them comparable across a broad range of data types, sampling and processing methods. Records were smoothed by fitting lowess curves[38] to the data in a two-stage process that prevented records with the highest resolution from dominating the global signal, and that avoided interpolating data in the low-resolution records. This approach also minimizes the impact of the extreme values in sedimentary charcoal data, which allowed us to focus instead on the long-term trends in the data. Various sensitivity analyses were performed to ensure that the composite records were robust to the use of influx versus concentration values, to the effects of changing sedimentation rates, and to the differences in coverage among regions (see the Supplementary Information).

We compared the reconstructed changes in fire regimes with climate proxy records of Northern Hemisphere (NH) temperature variations during the past 2000 years[18], and confirmed that these data series are statistically correlated during the interval from 1 to 1750 A.D. (see Supplementary Information). We used ten NH temperature reconstructions based on multiple climate proxies. These reconstructions differ in (a) data type, (b) data origin (terrestrial and/or marine), (c) inclusion or exclusion of extratropical data sources, (d) the number of data points used in the final reconstructions, (e) temporal length, (f) seasonal specificity (summer vs. mean annual temperature), and (g) statistical methods used to derive the reconstruction. Nevertheless, as shown in the Intergovernmental Panel on Climate Change (IPCC) Fourth Assessment[18], the reconstructions show broad coherence over the past 1300 years on centennial time scales, and the existence of multiple reconstructions makes it possible to derive confidence limits based on the degree of coherence between the available reconstructions. Here we use the mean of the 10 available reconstructions, smoothed with a lowess curve. We also show the uncertainties in the reconstructions.

There have been some attempts to reconstruct historical changes in global mean temperatures[39], and in southern hemisphere temperatures[40]. Given the paucity of data available from the southern hemisphere covering more than the past few decades, comparison of such data with the reconstructed fire histories is unwarranted. However, we did compare *simulated* global changes in warm season temperatures and dry season precipitation rates from the National Center for Atmospheric Research Climate System Model (NCAR CSM) (Fig. 22).

To explain the variations in global biomass burning, we also compared the charcoal data with population and land-cover data. World population and land-cover data were obtained from the HYDE 3.0 database[27] and were summarized using lowess curves in a manner similar to the charcoal and climate data. Confidence intervals (95%) are also shown (Fig. 11).

CHAPTER V

CONCLUSIONS

The results from the studies presented in Chapters II, III and IV fill a critical gap in our knowledge and understanding of the role of fire in the Earth system on millennial- to orbital-time scales. Fossil charcoal from millions of years ago has informed our understanding of fire on geological time scales, and global fire activity during recent decades and centuries has been actively studied through the lenses of satellite data, historical records and dendrochronological evidence. Paleofire reconstructions based on sedimentary charcoal records help bridge the large gap between these geological and modern studies on fire activity.

In the following sections, I review what we have learned about global fire history from the Last Glacial Maximum through the late Holocene in the context of the broader paleoecological literature. I then discuss a research agenda that can be implemented by conducting more meta-analyses of the new Global Charcoal Database. I conclude with a list of key lessons learned from this dissertation.

Global Fire History

Last Glacial Period

At the peak of the last glacial period about 21,000 years ago, ice sheets in the Northern Hemisphere were at their maximum extent and conditions were generally colder and drier than today (Braconnot et al., 2007). Maps of global paleofire activity around this time, though limited by data availability, show that the Americas, Europe and Australia were experiencing low levels of fire activity (although the existence of *some* charcoal in almost all records indicates that fire was very rarely entirely absent (Power et al., 2008)). Southeastern Asia and tropical Africa were exceptions, however, and record higher-than-present levels of burning during the LGM, probably due to greater-than-present fire frequencies associated with dry conditions (van der Kaars et al., 2001) and in some areas the expansion of grasslands (in East Africa and Eastern Asia) (Wooller et al., 2000; Wang et al., 2005).

The factors most limiting to fire during the glacial period are unknown, but two explanations are likely (Harrison et al. in prep). First, climate conditions were generally cold and dry, which limited vegetation growth and thus fuel availability, particularly in the northern extra-tropics. Exceptions to this pattern occurred in areas under the influence of an intensified jet stream (Thompson and Anderson, 2000) or stronger westerlies (Moreno et al., 1999), where wetter conditions could have decreased fire activity. Second, an overall decrease in the strength of the hydrological cycle under global cooling may have also reduced convective activity and lightning in some places, just as global warming is projected to increase the strength of the hydrologic cycle (Trapp et al., 2007; Trapp et al., 2009).

The Last Glacial/Interglacial Transition

As global temperatures increased, the ice sheets began to retreat, sea levels rose, and vegetation communities expanded into recently deglaciated environments. Sites with greater-than-present fire activity became increasingly evident after about 14,000 years ago in the Americas and Australia, although most locations there and in Europe continued to register low levels of burning (Power et al., 2008). In some parts of the world, particularly in the Northern Hemisphere, several large and rapid climate changes occurred, including ones at 12.9 ka and 11.7 ka, which marked the beginning and end of the Younger Dryas Chronozone (YDC) (Alley, 2000). The end of the YDC in particular has been widely recognized as a disturbingly large, fast, and widespread warming event. The vegetative impacts of these abrupt shifts have been studied extensively in North America (e.g. (Shuman et al., 2002)), South America (e.g. Heusser, 1998), and Europe (e.g. Walker, 1995; Isarin and Vandenberghe, 1998; Renssena et al., 2001) to gain insights into how ecosystems respond to such events. Following this tradition, Marlon et al. (2009) synthesized 35 paleofire records from North America to examine changes in fire activity during the transition period that encompasses several of these important events. The authors also tested the hypothesis that a comet explosion above the Laurentide Ice Sheet 12,900 years ago caused wildfires to sweep the entire continent, obliterating forests and contributing to the demise of megafauna and Paleoindian populations (Firestone et al., 2006; Firestone et al., 2007).

The study found:

- No evidence of continental-scale burning at any time during the last glacial/interglacial transition. This result does not say anything about whether a comet impact occurred, but it does help rule out two things: 1) continental-scale fires, and 2) widespread fire-related impacts as a mechanism for destroying the food supply of megafauna and/or Paleoindians.

- A strong upward trend in biomass burning during the last glacial/interglacial transition due to increasing Northern Hemisphere temperatures. Biomass burning was clearly not constant as suggested by a recent study based on isotopes from atmospheric methane coupled with a computer model (Fischer et al., 2008).

- Three distinct intervals with unique fire-fuel-climate conditions that indicate a strong correlation between millennial-scale climate changes and fire activity. The intervals were: 1) a pre-YDC interval with increasing biomass levels and fires due to higher-than-previous temperatures (i.e. the Older Dryas and Bolling/Allerød chronozones); 2) the YDC interval with no increase in biomass burning levels as a result of generally cold conditions; and 3) a post-YDC interval with increasing fires as a consequence of increasing summer temperatures but despite decreasing biomass levels.

- A short-term (decadal- to centennial-scale) increase in both biomass burning and fire frequency during the abrupt warming at the end of the YDC. The pollen data show no widespread change in biomass levels at this time, but this probably reflects the limitations of the pollen data resolution and/or the analysis methods. Previous research has demonstrated that vegetation mortality can increase during rapid climate changes (Tinner et al., 2008), but there is no clear evidence for this from the relatively coarse pollen data in the study. More detailed analyses on the pollen data could help determine whether fuel build-up from increased plant mortality (e.g. Mantgem et al., 2009; Adams et al., 2009) may have contributed to the increased fire activity.

Overall, these results show that the relationship between increasing temperatures and increasing fire identified in Power et al. (2008) hold true for North America during the last glacial/interglacial transition. Moreover, they provide evidence of past increases in fire during periods of rapid warming that are consistent with the widespread patterns of burning observed in recent decades.

The Late-Glacial Period and Early Holocene

Between 11,000 and 10,000 cal yr BP, global paleofire maps indicate that for the first time there were about an equal number of sites showing less-than-present burning as there were showing more-than-present burning, with at least four distinct spatial clusters of higher-than-present fire activity. South-central Europe, for example, along with the southern part of the St. Lawrence region in North America, southern South America, and southeastern Australia all have clusters indicating higher-than-present fire activity. In some areas, such as eastern North America and Europe, increased aridity due to the collapse of nearby ice sheets and the associated atmospheric circulation and regional climate changes probably caused the observed increases in fire activity (Carcaillet and Richard, 2000). In Patagonia (and possibly in southeastern Australia), increased burning was a result of widespread warmth and aridity due perhaps to a lower-than-present latitudinal insolation and temperature gradient and a consequent southward shift of storm tracks that deprived the region from moisture (Whitlock et al., 2007). Central South America also shows many sites with higher-than-present fire activity. In contrast to the increased coherence in high fire activity, Asia and Africa began to show increased heterogeneity in burning at this time.

The Early to Late Holocene

The early to late Holocene (i.e., 10,000 – 2000 cal yr BP) was not a primary focus of the analyses of this dissertation, although there were substantial and important shifts in global climate and many interesting changes in large-scale atmospheric circulation patterns that beg additional research. Nevertheless, some conclusions can be drawn about the interval based on the Power et al. (2008) time-slice maps and time-series curves.

In the charcoal anomaly maps, for example, heterogeneous spatial patterns (Ch. II, Fig. 5 panel c and Fig. 6a, b) are increasingly evident as the Holocene progresses. The apparent increase in the variability of fire activity may be due in part to the greater number of sites recording fire through the Holocene, thus reflecting the natural variability of fire across space. However, changes in global climate conditions triggered by insolation changes were also undoubtedly a contributing factor, both directly and indirectly. The melting of the large ice sheets in the Northern Hemisphere, for example, which was triggered by insolation changes, was essentially complete by 6000 cal yr BP. Thus the ice sheets were no longer exerting a strong control on atmospheric circulation patterns and regional climate conditions. As a result, interactions between fire and vegetation began to play a much greater role in determining the spatial and temporal distribution of fires across the landscape. In addition, insolation changes that affected ocean and land-surface conditions globally would have caused feedbacks to atmospheric circulation (Kohfeld and Harrison, 2000), which in turn would have directly triggered some changes in fire activity.

Not only did the spatial heterogeneity of fire increase, but overall levels of burning tended to be lower in the middle to late Holocene than they were in the early Holocene (Power et al., 2008). Maps of mid-Holocene global vegetation patterns (Prentice et al., 1996) indicate that there was a widespread expansion of more mesic taxa (forest into shrub/grassland), indicating wetter-than-previous climate conditions that would naturally prevent ignitions and fire spread. A shift to more mesic taxa also implies an increase in less flammable vegetation, which would further reduce fire (e.g. in temperate deciduous forests) (Mutch, 1970). Studies in the summer-dry area of the northwestern U.S. demonstrate how decreasing summer insolation coupled with changes in atmospheric circulation promoted cooler and wetter-than-previous summer conditions that in turn reduced the high fire activity previously observed during the early Holocene (Marlon et al., 2006; Whitlock et al., 2008). Also, at least in the northern hemisphere, if less intense seasonal differences in insolation created relatively mild climate conditions, local factors such as landscape topography and geology would become more important in controlling fire regimes, which would also increase spatial heterogeneity.

The Past 2000 Years

Environmental changes during the past 2000 years are particularly relevant today and have received increasing amounts of attention for a variety of reasons. First, this period encompasses the rapid expansion of human civilizations and the enormous impacts on the environment associated with this expansion (Turner II, 1990). Second, written records and observational data that document environmental changes and their impacts on people have become increasingly available and abundant. As such, the past 2000 years provides critical context for examining the impacts of humans on the environment. Third, two relatively large global-scale climate shifts occurred during this time – the "Medieval Climate Anomaly" and the "Little Ice Age," both of which had substantial impacts on people around the world. Fourth, although climate shifts occurred, many broad-scale or background environmental conditions, such as seasonal insolation levels, ice sheet extents, vegetation distribution and community composition, were relatively stable during the past 2000 years (excluding the last century or two), making it a useful and appropriate base period for evaluating decadal- to centennial-scale processes.

Paleofire data during the past 2000 years were analyzed by Marlon et al. (2009) to reconstruct global and regional changes in biomass burning and to assess the relative importance of climatic versus human influences during the interval. The global biomass burning trend shows a gradual decline from 1 A.D. to 1750 A.D., an abrupt increase from 1750 to 1900, and then a rapid decline towards the present. The trend parallels observed and simulated decreases in global and Northern Hemisphere temperatures (Jansen et al., 2007); but there is no clear parallel to simulated precipitation data, which show little consistency. Trends in population and land-use data also cannot explain the 2000-year decline, because at relatively low densities, population and fire activity should exhibit a positive relationship (Archibald et al., 2009). Our data show the opposite – biomass burning was declining while population densities and land-use change increased for most of the past two millennia. However, synchronous increases in population, land-use and biomass burning did occur after 1750 A.D. This parallel increase in population growth, land-use and burning lasted until ~1870 A.D., at which point the rate of population growth and land-use changes slowed and biomass burning began a sharp decline towards present. During this time temperatures increased rapidly. The reversal in the biomass burning trend at 1900 towards decreasing fire activity thus illustrates a threshold response of fire to population growth whereby increasing population densities cause increasing burning up to a point, after which further increases in population lead to a reduction in fire activity. Such a relationship has been demonstrated in recent research in South Africa (Archibald et al., 2009) and in a modeling study (Pechony and Shindell, submitted), but the interactions between fire, people and land-use is surely more complicated. Further research is needed to sort out the nature of these relationships in different locations and across vegetation types. The most recent reduction in global biomass burning (i.e. during the past few decades) may be considered mainly an extra-tropical signal, because data from regions where broad-scale and well-documented "deforestation fires" (Cochrane, 2003; Malhi et al., 2008; Morton et al., 2008) are occurring are almost entirely absent from the database.

The results from Marlon et al. (2008) provide important long-term context for human influences on burning. The data show that humans have had enormous impacts on global fire regimes in the past 150 years – far greater than any previous impacts at this scale. Widespread fire activity in the 1800's due to land clearance and logging caused widespread burning is well documented in historical records (Heyerdahl et al., 2001; Williams, 2006). The reduction in fire from grazing (e.g. Savage and Swetnam, 1990), agriculture, urbanization and direct suppression is likewise widespread and well-known (Pyne, 1995; Engelmark et al., 1998). The paleofire data syntheses reveal important new details, however, about the anomalous nature, extent and magnitude of these changes in a long-term context.

Future Research Directions in Synoptic Paleoecology

The development of the new Global Charcoal Database (GCD) is an example of research that promotes our ability to address a suite of complex problems at the in coupled human-environment dynamics. Research opportunities that have arisen from the creation of the database can be organized into four categories, including:

1) the analysis of issues relating to scale and pattern in paleoecology
2) targeting the collection of new paleofire data
3) the synthesis and exploration of paleofire data and the development of new hypotheses
4) hypothesis-testing with paleofire data

Research does not necessarily fit neatly into a single category or follow a particular sequence. Nevertheless, the categories highlight four important aspects of paleoecological research that are now possible given the existence of the GCD.

Analysis of Scale and Pattern in Paleoecology

Synoptic paleoecology deals with biotic and abiotic components and processes of ecosystems that occur at multiple spatial and temporal scales. As such, the concept of scale as it relates to the hierarchical nesting of systems (Urban et al., 1987) is central to research in this field, although it is rarely examined explicitly or in depth (but see

Williams et al., 2004). Among other things, scale can be used as a tool to 1) identify similarities and differences in shared elements and processes that vary by scale; 2) describe how particular scales highlight some aspects of phenomena and obscure others; 3) identify and explain scale-dependent processes; and 4) examine relations between scales. The creation of biogeographic databases facilitates and promotes such research, which is currently lacking in the environmental sciences. In his editorial remarks to the millennium issue of the *Journal of Biogeography*, Thompson Webb III reflects on the issue of scale in biogeography (Webb III, 2000). He imagines tools that would allow us to view biogeographic data with a zoom lens in space, time and taxonomy, or the "peopling" of earth (Watts, 1985) from different spatial and temporal scales. He also emphasized the constructed nature of our views and reconstructions of the past, which are uniquely shaped by the lenses (e.g. data) that we use. Webb illustrates this latter point by recalling a discussion in Paul Colinvaux's text book (Colinvaux, 1973; Colinvaux, 1993) that contrasted Alfred Russell Wallace's evolutionary interpretation of maps of continental flora and fauna (Wallace, 1876) with Allen's mapping of ecological life zones in North America as controlled by climate (Allen, 1871).

The importance of scale and spatial patterning to landscape processes is now well-recognized (Turner, 1989), and insights from this work may apply to the scaling-up of paleofire research from landscape to global scales. Research on the effects of changing spatial scales and the patterning of fire may be particularly helpful in understanding the controls on long-term fire activity. Because meta-analyses by their nature expand the spatial extent of traditional paleofire studies, additional information is needed about how representative different fire histories are of broader levels of fire activity (e.g. Gavin et al., 2006). Also, more data are needed on how particular vegetation, topographic, landscape and other characteristics associated with particular sites or regions affect long-term changes in fire activity. Although such aspects are considered when interpreting fire-regime changes at particular sites (Gavin et al., 2003; Brunelle, 2005), only modern, more spatially-extensive studies based on dendrochronological, historical or satellite data (e.g. Grau, 2001; LaFon and Grissino-Mayer, 2007; Yang et al., 2007) have focused specifically on how the importance of these spatial elements and patterns change with scale.

Addressing issues of scale in paleoecology and in the environmental sciences more broadly is more than a fascinating topic of study – it is a crucial need. Our inherent difficulties in grasping and visualizing phenomena at multiple spatial scales, and particularly at scales larger than those of the landscapes in which we operate or the temporal scales of days and years at which we live our lives, has become a major obstacle to appreciating, and thus responding to, many of the most dangerous threats we collectively face. Global climate change, ocean acidification, biodiversity loss; such issues are not easily grasped or visualized even by those trained and practiced at examining phenomena at broad spatial scales. Improving our ability to collect, organize, visualize, explore and analyze Earth system information across various scales is therefore imperative.

Collection of New Paleofire Data

A fundamental aim of paleofire research is to document and understand pyrogeography – or the variations in fire through time and across space. To meet this goal, data are needed from many different environments, across climate gradients, vegetation gradients, in continental and coastal regions, in moist and arid environments, in places where human activities have been intensive and in places where it was limited.

The GCD has highlighted places where data are most needed and can be used to help target new data collection efforts. Data are currently most abundant in developed countries, particularly in North America, Europe and Australia, for obvious reasons. Data are particularly needed in the tropics (e.g. in Indonesian peats, where vast carbon stores exist and fires have been particularly severe in recent years), and in boreal and tundra regions, for example, because of the high sensitivity of Arctic environments to climate change and the potential for changes in high-latitude fire regimes to have substantial feedbacks to the global climate system. Similarly, data from grasslands/savannas are currently very limited and would be particularly helpful in improving our understanding of long-term fire dynamics in these vegetation types as well as in reconstructing more robust and representative regional paleofire syntheses.

In comparison to pollen data, charcoal data collection is relatively easy, fast and inexpensive, and analysis methods and conceptual models are rapidly improving (Conedera et al., 2009; Higuera et al., 2007). There are ample opportunities for re-coring existing sites, particularly where other paleoenvironmental data exists but where high-resolution charcoal data are currently lacking. Many opportunities also exist for re-analyzing older paleofire datasets with new approaches to address unresolved questions.

Synthesis and Exploration of Paleofire Data

Of equal importance to data collection is the need to gather and organize the substantial volumes of data that have already been collected. This is not an easy or straightforward task, and the skills it requires are unfortunately not part of the practices traditionally taught to new paleoecologists. As a result, synthesizing paleoecological and paleoclimatological data tends to require collaborative and interdisciplinary efforts such as those described in this dissertation. Syntheses and explorations are needed to 1) document patterns and trends in paleofire activity; 2) explain these trends through comparisons with other datasets; 3) conduct data-model comparisons. New hypotheses arise naturally from all of these processes.

Document Patterns and Trends in Paleofire Activity

A natural starting point for research employing the GCD is the creation of new "composite" curves that reflect changes in the average level of biomass burning for some geographic region. A better characterization of regional- to global-scale fire activity will usually provide support for or against existing hypotheses or lead to the development of new ones regarding the potential controls on the patterns. Developing hypotheses can in turn raise questions about the likely pathways or mechanisms by through which the primary controls operate.

The paleofire data may be explored and analyzed in different ways and in conjunction with other datasets at each of these stages, and with some luck, eventually deepen our understanding of the nature and causes of paleoenvironmental change. The new knowledge and understanding can then support the application of research findings to current and future environmental challenges, which guide the development of new research efforts and bring the process full circle.

As just one example, a new synthesis effort aimed at creating a summary of Holocene fire activity for Europe (none currently exists) would provide important baseline information about the relative importance of climate, vegetation and people on fire regimes in Europe during the Holocene. Clusters of high or low fire activity observed during particular intervals would provoke questions about prevailing climate conditions, vegetation changes and human influences. Intervals and places with relatively heterogeneous fire patterns would help rule out the dominance of any particular control at those times. A well-developed record of past human activity and climate changes here means that the region lends itself to an analytical approach based on multiple working hypotheses. Similar analyses could be performed for any geographic region where there are enough data to warrant it. This kind of study would improve our knowledge of the geography of fire – where and when it was most important in the past, why and what impacts it may have had on people, vegetation and climate.

Explaining Patterns and Trends through Comparisons with Other Datasets

Comparative analyses are a powerful tool for explaining patterns identified through the generation of time-slice maps or composite curves described above. The availability of the new Global Charcoal Dataset will greatly facilitate such work. For example, as presented in Chapter IV, (Marlon et al., 2008) explained global trends in biomass burning during the past 2000 years through comparisons with temperature data from ice-core records and population/land-use data from archaeological sources. In Chapter III, (Marlon et al., 2009) used a similar approach to demonstrate the influence of temperature (based on ice-core data) and fuels (based on pollen data) on paleofire activity. Tinner and Lotter (2001) used comparisons between pollen and ice-core data to demonstrate a clear vegetation response to an abrupt climate change at 8.2 ka. It would be interesting to explore the role of fire in this transition.

Comparisons of charcoal, pollen, sedimentary and archaeological data allowed researchers to identify likely impacts of early colonizers on island vegetation and fire regimes in the tropics, including Madagascar (Burney, 1987), the West Indies (Burney et al., 1994), and the Hawaiian Islands (Burney and Burney, 2003). Further comparisons of charcoal data and *Sporormiella*, a dung fungus used as a proxy for megafaunal biomass, allowed Burney et al. to propose a link between changes in human activity (including burning) with megafaunal extinctions in Madagascar and elsewhere (Burney et al., 2003; Burney and Flannery, 2005); see also (Raper and Bush, 2009). Integrating such data with additional GCD records from other islands in the tropics and beyond (e.g. Stevenson, 2004; Hannon et al., 2005; McGlone et al., 2007) could provide a robust perspective on the role of human disturbance in island ecosystems more generally (Burney, 1997). Finally, charcoal data syntheses can be used as a benchmark for comparison with individual black carbon records, for example from sediments and from ice cores. Similar trends in the two independent datasets would provide further support for the use of black carbon records as a proxy for broad-scale biomass burning.

The efficiency, quality and depth of the research outlined above depend heavily on the availability and accessibility of charcoal data but also of pollen data. Because charcoal data reflects trends in biomass burning or fire frequency, information about vegetation and fuel is critical for interpreting and explaining the results. Although the Global Pollen Database provides easy access to many time-series of pollen data, the database has not been updated since 2000 and is not currently being maintained (i.e., new records are not being added). Moreover, it includes only a small fraction of the existing data from Europe, Asia, Africa and South America. Thus, much work remains to create a more comprehensive, integrated and truly *global* pollen database.

Data-Model Comparisons

There are three modeling communities interested in broad-scale fire activity: 1) those interested in vegetation dynamics, who treat fire as a disturbance agent that alters the distribution and structure of vegetation (e.g. Thonicke et al., 2001; Spessa et al., 2003); 2) those interested in atmospheric chemistry, who treat fire as a source of trace gas and aerosol emissions (e.g. Koch and Hansen, 2005); and 3) modelers focusing directly on the process of fire itself (Flannigan et al., 2001; Bond et al., 2005; Flannigan et al., 2008). Each of these communities approach and simulate fire or its likelihood in different ways and produce different results. Thus, validation datasets are important to these communities. Satellite and climate data provide targets for simulating modern fire activity, fire risk or emissions, but paleofire data can provide a strong test for models that are used to simulate past fire activity responses to large climate changes.

In order to prepare a paleo dataset for data-model comparison, the data must be cleaned, verified, standardized, organized and made publicly available and *accessible*, particularly to non-specialists. Modelers are generally unfamiliar with paleoecological field methods, chronologies, analytical and standardization techniques and other aspects that influence how the data are interpreted. As a result, a substantial amount of work is required to process and synthesize the individual records in the Global Charcoal Database before they become usable in a data-model comparison. Developing and documenting a protocol for data-model comparisons could facilitate the use of charcoal records in future comparisons. Automating some aspects of the data preparation and synthesis and/or incorporating standardized data within the database itself would also facilitate the use of the charcoal data by communities both within and outside of paleoecology.

Testing Hypotheses: Disentangling Climate-Humans-Fire-Vegetation Interactions

Paleofire data have been used to test a number of hypotheses regarding the interactions between climate, people, vegetation and fire, which are varied, complex and challenging to separate. Hypotheses have often focused specifically on the controls, impacts and feedbacks of fire regimes. Meta-analyses of paleofire data offer an additional tool for addressing such hypotheses, which is a critical step in developing strategies to manage fire and resources.

Controls of Fire Activity

Paleoecologists have used several approaches to understand the controls on fire. A classic example is that of Cygnet Lake in Yellowstone National Park, where infertile rhyolitic soils have helped maintain fire-prone lodgepole pine forests around the lake for millennia (Millspaugh et al., 2000). The site is remote and was presumably never used by people, thus its fire history reflects the changing influence of climate alone. Other studies have addressed hypotheses about climate versus vegetation or geology as the dominant control on fire regimes by strategically selecting sites distributed across strong climatic (Brunelle, 2005; Whitlock et al., 2007) or geological (Briles et al., in prep) gradients, or near modern vegetation ecotones (Camill et al., 2003; Umbanhowar Jr et al., 2006; Higuera et al., 2008) and examining their fire-regime responses across these gradients and ecotones (which may be stable or changing) through time. Meta-analyses could be used to do similar studies that span broader-scale gradients, or combine datasets from multiple authors and studies that were originally collected and analyzed for different purposes.

Hypotheses regarding the impacts of human activities on past fires can also be approached using meta-analyses. For example, ongoing debates regarding the nature and extent of aboriginal burning on the landscape in Australia, the Americas, and elsewhere (Singh et al., 1981; Denevan, 1992; Vale, 2002) have increasingly relied on composite analyses of sediment cores (Haberle, 2001) or soil charcoal dates (Bush et al., 2008) to distinguish between human and climatic controls on fire. In many cases data are insufficient to provide conclusive evidence in support of humans or climate as the dominant control on fire, but increasing spatial coverage and resolution of charcoal records (Black and Mooney, 2006; Walsh et al., 2008) should foster important progress in this area.

Impacts of Fire Activity

As with the controls on fire, the impacts of fire are complex and difficult to measure, yet our ability to model fire depends in part on how well we understand it. Fire impacts can be thought of as a feedback process in many cases, either to the climate system directly, or to vegetation and then climate. Further paleofire work is needed to determine the impacts and feedbacks (or lack thereof) that may have resulted from increased human ignitions on vegetation and climate (Ruddiman and Thompson, 2001), decreased human ignitions on vegetation and climate (Nevle and Bird, 2008), and on the ecological impacts of changing fire regimes (Bond and Keeley, 2005; Bond et al., 2005).

After the publication of Denevan's important work "*The Pristine Myth*" in 1992 (Denevan, 1992), which increased attention on the complex and widespread impacts of Native populations throughout the Americas, fire has increasingly been described as one of the primary tools used by Native Americans to transform pre-historic landscapes. This line of thought has stimulated a number of ongoing debates that revolve around the question of whether humans or climate were a stronger control on past fire regimes (Boyd, 1999; Whitlock and Knox, 2002; Baker, 2002), but questions regarding the impacts and feedbacks of fire have also generated debate (Brook, 2009). Most noteworthy is the hypothesis popularized by Bill Ruddiman, which states that biomass burning along with other human activities such as rice cultivation, animal husbandry and deforestation began to alter global climate several thousand years ago by increasing atmospheric CO_2 and CH_4 concentrations (Ruddiman and Thompson, 2001). The more widely-held view is that human activities did *not* substantially affect greenhouse gas concentrations before 1750 A.D. Paleofire data are highly relevant to this question, as they can provide evidence of trends in past burning at the spatial and temporal scales needed to resolve the questions.

Recently, a book titled "*1491: New Revelations of the Americas Before Columbus*" further popularized the notion that indigenous people had extensive impacts on the pre-Columbian landscape (Mann, 2005). Subsequently, Nevle and Bird (2008) used charcoal data to support their argument that a pandemic-related *reduction* in human-caused burning after 1492 A.D. led to massive reforestation in the tropics and an associated decline in atmospheric CO_2. Marlon et al. (2008) however used a meta-analysis of charcoal records from the region in comparison with climate data to argue that the long-term changes in climate matched the patterns of burning better than changes in population density. They also noted that fire activity began to decline prior to 1492 A.D., which is consistent with the climate data but not with the population data. Additional work with the more comprehensive network of charcoal data now available, however, in conjunction with pollen-based evidence of vegetation change could be helpful in resolving the debate (Bush et al., 2008; Mayle and Power, 2008).

Interactions of Fire and Vegetation

Given the range of complex vegetation-fire interactions observed today, which vary across spatial and temporal scales (Bessie and Johnson, 1995; Schoennagel et al., 2004; Balch et al., 2008), it is not surprising that disentangling such interactions in the paleo record is a critical but challenging task (Gavin et al., 2007). Vegetation changes that impact fuels and fire may be comprised of variations in the type, distribution, productivity, structure and effects of these on fuel distribution, load, connectivity, flammability (Bond and van Wilgen, 1996). Efforts to determine the causal direction of changes in fire and vegetation (e.g. leads and lags), for example, are hampered by limitations inherent to the data, such as low temporal and spatial resolution and a lack of independent records of climate change (Delcourt and Delcourt, 1991). Marlon et al. (2009), for example, was unable to identify any systematic vegetation responses in western North America during the YDC, despite distinct and rapid shifts in climate and fire regimes at that time. However, the pollen analyses used in that study were more coarse than charcoal data in terms of their temporal and taxonomic resolution, and it is unclear whether the apparent lack of response to shifts in climate and/or fire was due to the low resolution of the data, the analytical approach (e.g., summarizing arboreal pollen proportions at a very broad spatial scale), or whether there was simply little vegetation change. The latter hypothesis would be surprising in light of the detailed studies in eastern North America that showed large and rapid responses of vegetation to climate shifts around the YDC (Shuman et al., 2002). Yet it is possible that climate changes were less dramatic in western North America, or that the community compositions and/or mountainous terrain in the west may have allowed vegetation reorganizations to occur in more limited areas, which would in turn be poorly recorded in the pollen record.

Despite the challenges involved in reconstructing fire-vegetation interactions from pollen and charcoal data, several approaches have been successful. For example, Tinner et al. (2006) was able to identify cause-and-effect relationships between vegetation and fire in Alaska using correlation analysis (Tinner et al., 2006), and careful analysis of high-resolution data revealed the impacts of changes in fuel structure on fire regimes in the same region (Higuera et al., 2008). Now that hundreds of records are available and accessible through the global pollen and charcoal databases, performing comparative analyses at multiple spatial scales is a less daunting task. Integrating these datasets with additional independent datasets of climatological, hydrological and other paleoenvironmental changes should lead to new insights about the ecological response to a wide range of forcings and feedbacks, including abrupt climate changes, the sequence and pathways of past changes, leads and lags in fire-vegetation interactions, and the role of fire in causing shifts between dynamic stable states of community composition (Umbanhowar Jr et al., 2006; Umbanhower, 2004). Understanding the synergistic effects of fire with other disturbances (e.g. rapid climate warming or insect-related diebacks that cause a rapid growth in fuels) is a particularly urgent research need (as it is with modeling) in light of the recent trends in tree mortality, insect outbreaks, etc. described above.

While fire-vegetation interactions are clearly important, particularly at local scales, it should be noted that recent observations (Gillett et al., 2004; Westerling et al., 2006) as well as paleo data (Marlon et al., 2009) suggest that vegetation and fuel characteristics may be less important during periods of relatively large and rapid climate changes than during intervals of relatively stable climate conditions. As Moritz et al. (2004) point out, the fuel-age structures of particular stands that are so carefully examined to predict fire behavior during most years are essentially meaningless in the face of extreme fire weather, when wildfires sweep across diverse ecosystems regardless of community composition, distribution, structure and other fuel characteristics that would normally be important determinants of fire behavior.

Fire Feedbacks to Climate

Further paleofire data syntheses may also be useful to modern efforts aimed at understanding present-day feedbacks between fire and the carbon cycle. Investigating carbon cycle dynamics is generally accomplished using models, but paleodata can be used to understand the underlying processes, to test the models, and to generate new hypotheses. For example, in their recent *Science* paper, Bowman et al. (2009) state that with the exception of deforestation fires, the long-term carbon dioxide emissions from wildfires are assumed to exist in a steady state because the re-growth of vegetation during the process of succession, which should absorb any carbon dioxide that was emitted. Yet, just two paragraphs later the authors remind us that new global vegetation models have indicated that forests would *at least double in extent* in the absence of fire. If fire increases then, as it appears to be doing on very broad spatial scales – how can we assume that such changes would maintain emissions in a steady state? It seems plausible at least that a widespread increase in area burned or fire frequency could shift the dominant pathways of vegetation succession and thus produce long-lasting effects on the carbon cycle and climate. Perhaps such changes would not produce significantly large effects on global climate, but large effects on biogeochemical cycles at regional scales seem likely, and could have important implications for carbon sequestration and management.

In fact, modeling experiments have already shown that changes in vegetation-fire dynamics are likely to have lasting effects on the carbon cycle (Lenihan et al., 2003; Scholze et al., 2006; Lenihan et al., 2008). Lenihan et al. (2008) projected an increase in total annual area burned under both cooler- and less-dry and warmer-and-drier scenarios. In the cooler- and less-dry scenarios total annual biomass consumption by fire was greater as well, although in warmer-and-drier scenarios simulated biomass consumption was initially greater, but then declined to historical or below-historical levels (Lenihan et al., 2008). In a different modeling study, Lenihan et al. (2003) showed that fire played a role in mediating tree-grass competition in response to changes in simulated precipitation. Likewise, Scholze et al. (2006) suggested that a doubling of CO_2 may increase fire risk in many forested areas. Such changes may be short-lived, but others may be longer-lasting. Paleo records of fire and vegetation changes provide ample evidence of such shifts, and could be drawn upon to illustrate the range of possible or probable outcomes based on a variety of initial conditions (i.e. prevailing vegetation and climate conditions). Specifically, pollen syntheses could be examined to identify places and times when forests were shifting towards savannas/grasslands. Paleofire data from the same region could then be analyzed to determine the fire-regime responses before, during and after the transition. A comparative analysis of such shifts in several places would provide a range of potential outcomes for a given set of conditions. Although such data do not provide quantitative estimates of ecosystem carbon loss/gain due to changing vegetation and fire regimes, knowing the likely direction of change under a variety of circumstances could be helpful for evaluating whether the "steady state" assumption is valid, or on what temporal/spatial scale it may be invalid.

All of the approaches outlined above can yield new insights into the controls, impacts and feedbacks of fire activity in different environments and during particular intervals. The availability of hundreds of charcoal records in a central repository (the GCD) now makes it possible to employ meta-analyses in the approaches outlined above in order to examine fire-related processes at multiple spatial scales.

Lessons Learned

Three main lessons emerged from this dissertation:

1) Climate is the primary determinant of broad-scale changes in fire activity.
2) Human activities caused fire activity to rise dramatically after 1750 A.D. and then decline rapidly during the 20th century.
3) A strong climate-fire link, massive human impacts on modern vegetation and fire regimes, and multiple warming-induced ecosystem stresses (e.g. increased tree mortality and pest outbreaks) all point towards a "perfect storm" for unusually high fire activity across broad areas in coming decades.

Lesson One: Climate Is the Primary Control on Global Fire Activity

A particularly robust conclusion from the paleofire meta-analyses presented here indicates that warm conditions lead to more fire than cool conditions, and vice versa. Furthermore, fire appears highly sensitive to climate changes insofar as rapid warming (cooling) events are associated with rapid increases (decreases) in fire activity. For example, the abrupt warming at the end of the Younger Dryas is correlated with an increase in both biomass burning and fire frequency in North America. Similarly, the decline in global biomass burning that occurred around 1500 A.D. is associated with relatively rapid cooling in the Northern Hemisphere during the Little Ice Age.

The idea that climate conditions, and specifically temperature, is strongly correlated with fire activity is not surprising – such a relationship is evident today as one moves from the high latitudes towards the equator, or from winter into summer months in the extra-tropics, for example. What is less obvious is that such a relationship should hold through time – and particularly over centuries and millennia – as well as across space. Thus, the paleo record provides evidence that the current link between increasing temperatures and higher fire activity (Westerling et al. 2006) in the western U.S., for example, is consistent with climate impacts on fire regimes in the past. Westerling et al. (2006) provide details about specific fire-climate linkages, including the earlier onset of spring, which leads to longer fire seasons and greater fire frequency and duration. A longer fire season increases the amount of time that fuels have to dry out, and also expands the window in which extreme fire weather can occur (Moritz et al., 2004). Studies elsewhere have documented similar climate-fire relationships, and several regional simulations (Pitman et al., 2007) plus one global study of future fire risk (Scholze et al., 2006) add further support. Climate also controls fire regimes indirectly, by strongly influencing net primary productivity (NPP), which in turn determines fuel accumulation rates. A new study by Krawchuk et al. (2009) based on statistical models combined with a GCM, highlights the importance of NPP as a direct control on modern global fire regimes. This new study also suggests that future increases in climate-driven fire will be balanced by climate-driven decreases in fire at the global scale. However, these results contractict the bulk of evidence to date that indicates that warming temperatures are leading to increasing fire activity. Moreover, when considered in the broader context of multiple interacting ecosystem stresses (e.g. human ignitions, invasive species, insect outbreaks and dieback) that increase ecosystem vulnerability and reduce resiliency, it would be surprising to see broad-scale declines in fire risk, at least in the near future.

Marlon et al. (2008) showed a correlation between decreasing temperatures in the Northern Hemisphere and declining global biomass burning. This finding confirms earlier evidence from Power et al. (2008) that cooler global climate conditions are associated with relatively low levels of biomass burning. Both of these studies are consistent with many previous fire-history studies that find decreased fire activity associated with the Little Ice Age (about 550-150 cal yr BP Grove, 1988).

Lesson Two: Humans Dramatically Altered Global Fire Regimes After 1750 A.D.

The gradual, climate-driven decline in biomass burning from 1 to 1750 A.D. identified by Marlon et al. (2008) stands in sharp contrast to the abrupt variations of the following 250 years. The sharp reduction in fire activity since ~1870 A.D. seems particularly surprising in light of the large number of fires that people ignite today (TNC, 2004; FAO, 2007). Yet, when the scope and magnitude of 20^{th}-century land-use changes are considered (Klein Goldewijk and Ramankutty, 2004; Ramankutty et al., 2007) – including widespread grazing, forest clearance, and agriculture – the major decline in global fire activity during the past century appears highly probable if not inevitable.

The important role that human activities have played in determining fire activity in recent decades, however, should not obscure the continuing importance of climate as a major control on global fire activity. Even during the 20^{th} century, climate – both directly and through indirect effects on vegetation – continues to determine the spatial distribution as well as the seasonal to inter-annual variability of fires (van der Werf et al., 2006; Krawchuk et al., 2009). For example, fires are naturally more frequent in seasonally dry, sub-tropical regions than in the temperate or boreal forests of higher latitudes – regardless of human impacts on ignitions. Likewise, the basic shape of the distribution of fire activity throughout the year is determined by climate, not people, although human influences modify that shape to some extent (Bartlein et al., 2008). It is important to note, however, that the paleofire data from the most recent decades are not robust due to sedimentological issues, and do not reflect the changing influences of people on fire regimes due, for example, to deforestation in the tropics (Cochrane and Barber, 2009), to migration from rural areas to cities (associated with increases in fire activity due to fuel-build up on abandoned agricultural lands), or to human-induced global warming since ~1970 (FAO, 2007).

Lesson Three: A "Perfect Storm" For Future Fire Activity

The strong climate-fire link identified in Lesson One suggests that the current increases in fire due to global warming (Westerling et al., 2006) are consistent with the impacts that past warming events had on fire regimes. The widespread reduction in burning during the 20th century (Lesson Two) is also consistent with increasing fire activity today, at least in forested areas where fire suppression has contributed to fuel build-up. Even where fire suppression has been less important, ecosystems face extended fire seasons, increasing vegetation productivity (Mickler et al., 2002), increasing tree mortality (Mantgem et al., 2009), and pest outbreaks due to the effects of global warming. Each of these changes tend to promote fire individually; synergistically they appear to be creating a "perfect storm" for unusually severe fire activity in parts of the western North America, southern Europe, eastern Australia, and other regions. The rapid development of former wildland areas under such conditions will undoubtedly be catastrophic in some cases (Moritz and Stephens, 2008).

Summary

Paleofire research can provide unique insight into the role of fire in the Earth system by providing information about how fire has interacted with climate and vegetation in the past on time scales ranging from decades to millennia. Compiling the many disparate datasets on past fire activity from the literature was an essential first step in building our knowledge of fire processes on long temporal and broad spatial scales. The real value of the new dataset, however only became apparent through the initial analyses, however, which provided our first glimpse into global-scale patterns of biomass burning during the past 21,000 years. Now, many new research opportunities have been created that have the potential to vastly improve our conceptual models of complex climate-vegetation-fire interactions, and of the implications of past, current, and future human activities on long-term terrestrial and atmospheric fire-related processes. Perhaps most critically, the formation of the Global Charcoal Database has opened the door to using paleofire data to help improve broad-scale fire modeling, which is essential for testing hypotheses about the coupled response of fire and vegetation to climatic variation

APPENDIX A

LIST OF CONTRIBUTORS TO CHAPTER II

Power, M.J., Institute of Geography, School of Geosciences, University of Edinburgh, Edinburgh, UK

Marlon, J., Department of Geography, University of Oregon, Eugene, OR, USA

Ortiz, N., BRIDGE, School of Geographical Sciences, University of Bristol, Bristol, UK

Bartlein, P.J., Department of Geography, University of Oregon, Eugene, OR, USA

Harrison, S.P., BRIDGE, School of Geographical Sciences, University of Bristol, Bristol, UK

Mayle, F.E., Institute of Geography, School of Geosciences, University of Edinburgh, Edinburgh, UK

Ballouche, A., Laboratoire Paysages et Biodiversité, Université d'Angers, Angers Cedex 1, France

Bradshaw, R.H.W., Department of Geography, University of Liverpool, Liverpool, UK

Carcaillet, C., Centre for Bio-Archaeology & Ecology (UMR5059 CNRS/UM2/EPHE), Montpellier, France

Cordova, C., Department of Geography, Oklahoma State University, Stillwater, OK, USA

Mooney, S., School of BEES, University of New South Wales, Sydney, NSW, Australia

Moreno, P.I., Facultad de Ciencias, Departamento de Biología, Institute of Ecology and Biodiversity, Nuñoa, Santiago, Chile

Prentice, I.C., QUEST, Department of Earth Sciences, University of Bristol, Bristol, UK

Thonicke, K., BRIDGE, School of Geographical Sciences, University of Bristol, Bristol, UK

Tinner, W., Institute of Plant Sciences, University of Bern, Altenbergrain 21, CH–3013 Bern, Switzerland

Whitlock, C., Department of Earth Sciences, Montana State University, Bozeman, MT, USA

Zhang, Y., Key Laboratory of Vegetation and Environmental Change, Institute of Botany, Chinese Academy of Sciences, Beijing, China

Zhao, Y., BRIDGE, School of Geographical Sciences, University of Bristol, Bristol, UK

Ali, A.A., Chaire industrielle CRSNG-UQAT-UQAM en aménagement forestier durable, Université du Québec en Abitibi-Témiscamingue, Noranda, QC, Canada

Anderson, R.S., Center for Environmental Science and Education, Northern Arizona University, Flagstaff, AZ, USA

Beer, R., Institute of Plant Sciences, University of Bern, Altenbergrain 21, CH–3013 Bern, Switzerland

Behling, H., Department of Palynology and Climate Dynamics, Georg-August University, Göttingen, Germany

Briles, C., Department of Earth Sciences, Montana State University, Bozeman, MT, USA

Brown, K.J., Department of Quaternary Geology, Geological Survey Denmark and Greenland, Copenhagen, Denmark

Brunelle A., Department of Geography, University of Utah, Salt Lake City, UT, USA

Bush, M., Department of Biological Sciences, Florida Institute of Technology, Melbourne, FL, USA

Camill, P., Department of Biology, Carlton College, Northfield, MN, USA

Chu, G.Q., Institute of Geology and Geophysics, Chinese Academy of Sciences, Beijing, China

Clark, J., Department of Biology, Duke University, Durham, NC, USA

Colombaroli, D., Institute of Plant Sciences, University of Bern, Altenbergrain 21, CH–3013 Bern, Switzerland

Connor, S., Social and Environmental Enquiry, University of Melbourne, Melbourne, VIC, Australia

Daniau, A.-L., Institut de Préhistoire et Géologie du Quaternaire, Université Bordeaux 1, Talence Cedex, France

Daniels, M., Ecological Restoration Institute, Northern Arizona University, Flagstaff, AR, USA

Dodson, J., Institute for the Environment, Brunel University, Uxbridge, UK

Doughty, E., Harvard University, Harvard Forest, Petersham, MA, USA

Edwards, M.E., School of Geography, University of Southampton, Southampton, UK

Finsinger, W., Institute of Plant Sciences, University of Bern, Altenbergrain 21, CH–3013 Bern, Switzerland; Palaeoecology, Laboratory of Palaeobotany and Palynology, Institute of Environmental Biology, Utrecht University, Utrecht, The Netherlands

Foster, D., Harvard University, Harvard Forest, Petersham, MA, USA

Frechette J., Department of Earth and Planetary Sciences, University of New Mexico, Albuquerque, NM, USA

Gaillard, M.-J., School of Pure and Applied Sciences, University of Kalmar, Kalmar, Sweden

Gavin, D.G., Department of Geography, University of Oregon, Eugene, OR, USA

Gobet, E., Institute of Plant Sciences, University of Bern, Altenbergrain 21, CH–3013 Bern, Switzerland

Haberle, S., Australian National University, Research School of Pacific and Asian Studies, Canberra, ACT, Australia

Hallett, D.J., Department of Geography and School of Environmental Studies, Queens University, Kingston, ON, Canada

Higuera, P., Department of Earth Sciences, Montana State University, Bozeman, MT, USA

Hope, G., Australian National University, Research School of Pacific and Asian Studies, Canberra, ACT, Australia

Horn, S., Department of Geography, University of Tennessee, Knoxville, TN, USA

Inoue, J., Department of Biology and Geoscience, Graduate School of Science, Osaka City University, Osaka, Japan

Kaltenrieder, P., Institute of Plant Sciences, University of Bern, Altenbergrain 21, CH–3013 Bern, Switzerland

Kennedy, L., Department of Geography, Virginia Polytechnic and State University, Blacksburg, WV, USA

Kong, Z.C., Institute of Botany, Chinese Academy of Sciences, Beijing, China

Larsen, C., Department of Geography, University of Buffalo, Buffalo, NY, USA

Long, C.J., Department of Geography and Urban Planning, University of Wisconsin, Oshkosh, WI, USA

Lynch, J., Biology Department, North Central College, Naperville, IL, USA

Lynch, E.A., Biology Department, Luther College, Decorah, IA, USA

McGlone, M., Manaaki Whenua - Landcare Research, Lincoln, New Zealand

Meeks, S., Department of Anthropology, University of Tennessee, Knoxville, TN, USA

Mensing, S., Department of Geography, University of Reno, Nevada, NV, USA

Meyer, G., Department of Earth and Planetary Sciences, University of New Mexico, Albuquerque, NM, USA

Minckley, T., Department of Botany, University of Wyoming, Laramie, WY, USA

Mohr, J., College of Forestry, Oregon State University, Corvallis, OR, USA

Nelson, D.M., Institute for Genomic Biology, University of Illinois, Urbana, IL, USA

New, J., Department of Earth and Planetary Sciences, University of New Mexico, Albuquerque, NM, USA

Newnham, R., School of Geography, University of Plymouth, Plymouth, Devon, UK

Noti, R., Institut de Botanique, Université de Neuchâtel, Bern, Switzerland

Oswald, W., Department of Communication Sciences and Disorders, Emerson College, Boston, MA, USA

Pierce, J., Department of Geosciences, Boise State University, Boise, ID, USA

Richard, P.J.H., Département de Géographie, Université de Montréal, Montréal, QC, Canada

Rowe, C., Département de Géographie, Université de Montréal, Montréal, QC, Canada

Sanchez Goñi, M.F., EPHE, UMR-CNRS 5805, EPOC, Université Bordeaux 1, 33405 Talence, France

Shuman, B.N., Geology and Geophysics, University of Wyoming, Laramie, WY, USA

Takahara, H., Graduate School of Agriculture, Kyoto Prefectual University, Kyoto, Japan

Toney, J., Department of Geological Sciences, Brown University, Providence, RI, USA

Turney, C., GeoQuEST Research Centre, School of Earth and Environmental Sciences, University of Wollongong, Wollongong, NSW, Australia

Urrego-Sanchez, D.H., Department of Biological Sciences, Florida Institute of Technology, Melbourne, FL, USA

Umbanhowar, C., Biology and Environmental Studies, St. Olaf College, Northfield, MN, USA

Vandergoes, M., GNS Science, Avalon, New Zealand

Vanniere, B., LCE CNRS-University of Franche-Comté, Besançon, France

Vescovi, E., Institute of Plant Sciences, University of Bern, Altenbergrain 21, CH–3013 Bern, Switzerland

Walsh, M., Department of Geography, University of Oregon, Eugene, OR, USA

Wang, X., Institute of Geology and Geophysics, Chinese Academy of Sciences, Beijing, China

Williams, N., Department of Environment and Conservation, University of New South Wales, Sydney, NSW, Australia

Wilmshurst, J., Manaaki Whenua - Landcare Research, Lincoln, New Zealand

Zhang, J.H., Chinese Academy of Meteorological Science, Beijing, China

APPENDIX B

SUPPLEMENTARY INFORMATION TO CHAPTER III

This appendix contains additional methodological details, 1 table that describes the sites in the main text, and 3 figures. Fig. 15 is a site map. Fig. 16 shows the individual records of charcoal influx and proportions of arboreal pollen during deglaciation along with background trends in biomass burning and radiocarbon dates. Fig. 17 shows the peak frequencies and magnitudes of the high-resolution sites, as well as the number and location of radiocarbon dates in these records.

Table 1. Study sites used in the North American paleofire analysis.

ID	Site Name	Lat.	Long.	Elev. (m a.s.l.)	Age Range (cal yr BP)
1	Ruppert Lake, AK[1]	67.07	-154.23	230	13,078 - 10,000
2	Xindi Lake, AK[1]	67.11	-152.49	240	15,000 - 10,000
3	Lost Lake, AK[2]	64.30	-146.69	240	14,650 – 10,000
4	Whyac Lake, BC[3]	48.67	-124.84	15	15,000 - 10,000
5	Pixie Lake, BC[3]	48.60	-124.20	70	15,000 - 10,000
6	Boomerang Lake, BC[4]	49.18	-124.15	360	13,422 - 10,000
7	Enos Lake, BC[4]	49.28	-124.15	50	15,000 - 10,000
8	Walker Lake, BC[5]	48.53	-124.00	950	15,000 - 10,000
9	Porphyry Lake, BC[5]	48.91	-123.83	1100	14,979 - 10,000
10	East Sooke Fen, BC[3]	48.35	-123.68	155	13,685 - 10,000
11	Battle Ground Lake, WA[6]	45.80	-122.49	154	14,290 - 10,000
12	Little Lake, OR[7]	44.17	-123.58	210	15,000 - 10,000
13	Bolan Lake, OR[8]	42.02	-123.46	1637	14,545 - 10,000
14	Bluff Lake, CA[8]	41.35	-122.56	1921	15,000 - 11,065
15	Mumbo Lake, CA[10]	41.19	-122.51	1860	15,000 - 10,000
16	Dead Horse Lake, CA[11]	42.56	-120.78	2248	15,000 - 10,000
17	Swamp Lake, CA[12]	37.95	-119.82	1554	15,000 - 10,000
18	Siesta Lake, CA[13]	37.85	-119.67	2430	13,241 - 10,000
19	East Lake, CA[14]	37.18	-119.03	2863	14,634 - 10,000
20	Foy Lake, MT[15]	48.17	-114.36	1006	13,134 - 10,000
21	Burnt Knob Lake, ID[16]	45.70	-114.99	2250	15,000 - 10,000
22	Baker Lake, ID[15]	45.89	-114.26	2300	14,328 - 10,000
23	Pintlar Lake, MT[15]	45.84	-113.44	1921	14,732 - 10,000
24	Slough Creek Lake, WY[17]	44.93	-110.35	1884	13,362 - 10,000

25	Cygnet Lake, WY[18]	44.65	-110.60	2530	15,000 - 10,000
26	Crane Lake, AZ[19]	36.72	-112.22	2590	13,835 - 10,000
27	Hunters Lake, CO[20]	37.61	-106.84	3516	14,273 - 10,000
28	Como Lake, CO[19]	37.55	-105.50	3523	13,602 - 10,000
29	Chihuahueños Bog, NM[20]	36.05	-106.51	2925	15,000 - 10,000
30	Moon Lake, ND[21]	46.86	-98.16	456	13,794 - 10,000
31	Sharkey Lake, MN[22]	44.59	-93.41	305	13,037 - 10,000
32	Hertel, QC[23]	45.68	-74.05	70	13,000 - 10,000
33	Albion, QC[24]	45.67	-71.33	320	13,566 - 10,000
34	J'Arrive, QC[24]	49.25	-65.38	56	14,055 - 10,000
35	Lake Tulane, FL[25, 28]	27.59	-81.50	35	15,000 - 10,000

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Data Sources and Locations

Charcoal and pollen data sources, site locations and elevation, and temporal coverage during the Last Glacial–Interglacial Transition (LGIT) are provided in Table 2. Numbers in parentheses after site names are reference numbers. Site locations are shown in Fig. 15 and are coded to reflect the existence of a charcoal peak at 12.9 and 11.7 ka, as determined by the analysis techniques described below.

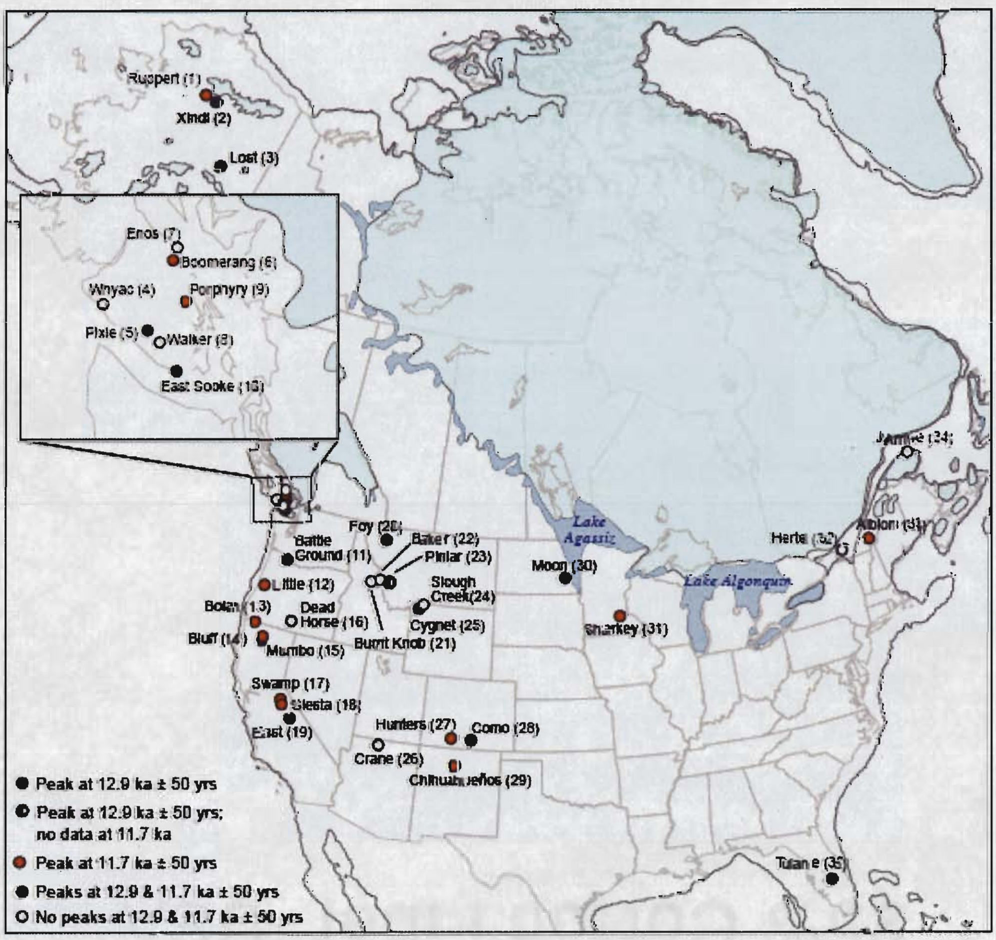

Fig. 15. Map of North America 13,000 years ago showing the locations of paleofire study studies.

Chronologies

Lake-sediment records of the kind selected for paleoecological studies have generally well-behaved sedimentation regimes with slowly varying sedimentation rates. This allows chronological control with fewer radiocarbon dates than required for discontinuous terrestrial records, or marine records with continuous or intermittent bioturbation of sediments. For example, the age of a synchronous vegetation change in eastern North America, the Tsuga (hemlock) decline, can be estimated with sub-century precision (4750 ^{14}C y BP, with a standard error of the mean of ~50 y) using networks of lake records like those used here (1, 2). However, owing to the reorganizations of the circulation of the atmosphere and ocean that are involved in the abrupt climate changes, larger than usual uncertainties arise in calibrating radiocarbon ages during the LGIT (3), and so we compared the results from our peak identification analysis based on both a narrow 100-y and wider 500-y window width. As described in the main text, peaks were not more likely to occur at 12.9 than at 11.7 ka in either case.

Analysis of Individual Charcoal and Pollen Records

Charcoal concentration data (particles cm^{-3}) were converted to influx values (particles cm^{-2} y^{-1}) by dividing charcoal values by sample deposition times (y cm^{-1}). For the low-resolution records (>50 y $sample^{-1}$), we used quantile regression to estimate background charcoal influx values as the 50th percentile (4). The degrees of freedom parameter (df) was 10 for all but 3 records (i.e., 73 for East Lake, 50 for Sharkey Lake, and 6 for Walker Lake).

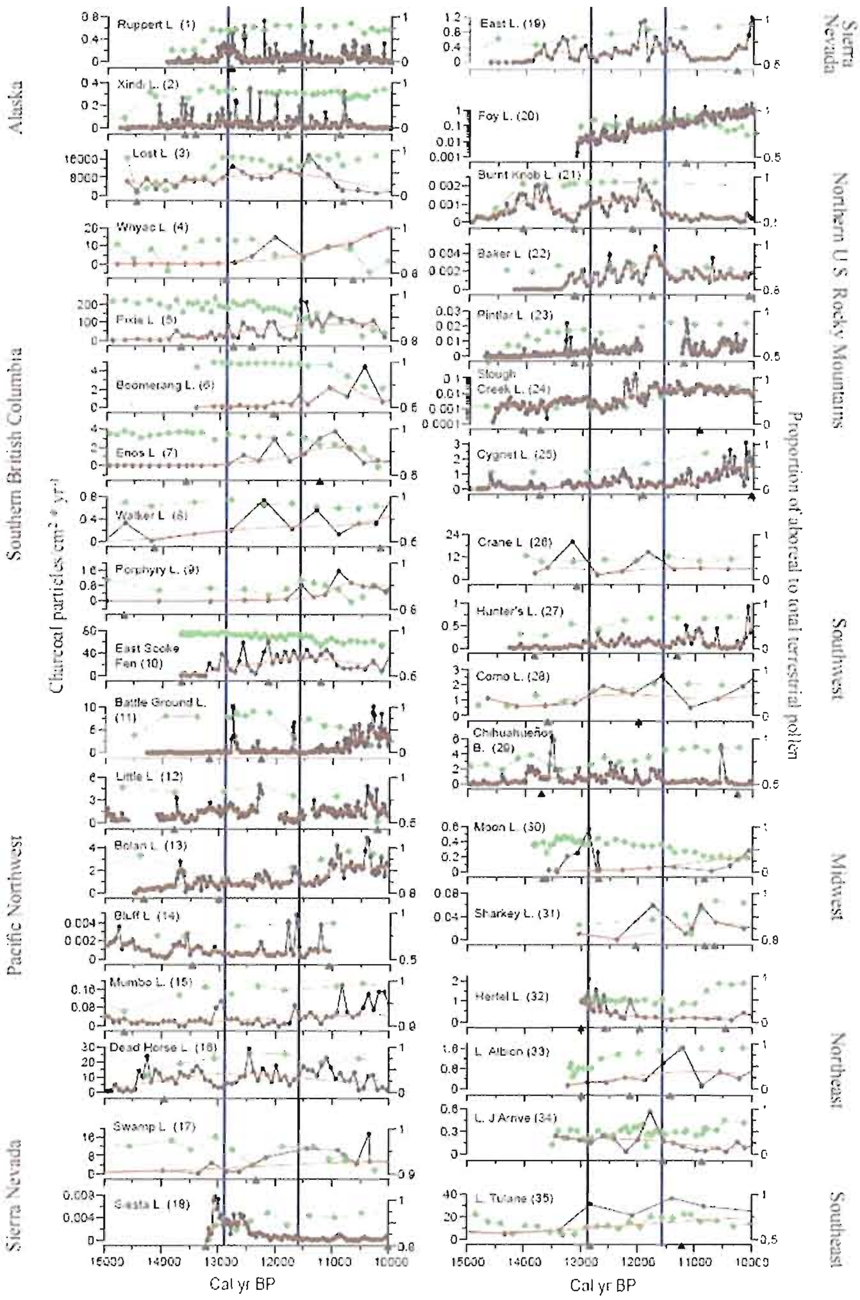

Fig. 16. Individual site records arranged by region. Each graph shows changes in charcoal influx (black), smoothed charcoal influx (red), and proportions of arboreal pollen (green) during deglaciation. Radiocarbon dates are represented by black triangles along x-axes. Site numbers are in parentheses following the site names.

In continuously sampled (high-resolution), macroscopic (typically >100 μm) charcoal records, large charcoal peaks above background represent individual, local fire events or clusters of events (fire episodes) as has been demonstrated by examination of the portions of the sedimentary records that overlap with dendrochronological or historical records of fire (5, 6). Lower resolution records based on microscopic charcoal (>100 μm), reflects burning at broader scales (7). Low-resolution records will integrate individual fire episodes, but increased fire activity can still be inferred from large peaks in low-resolution records (8). For Fig. 10 in the main text, any increase in charcoal influx above background within a defined time period (i.e., either ±50 or ±250 y) was considered a "peak."

The high-resolution (<50 y per sample) charcoal influx series were decomposed into background and peaks components using CharAnalysis (9), which allows us to reconstruct peak frequencies and to quantify peak sizes in addition to separating peaks from background charcoal levels (Fig. 17). Charcoal values were interpolated to constant time intervals based on the median resolution at each site. A robust lowess smoother was used to define background trends with a 500-y window width for all but 2 records (sites 18 and 13), which showed an improved signal-to-noise ratio with larger window widths (18). Site 18 was smoothed with a 600-y window and site 13 was smoothed with an 800-y window. Peaks were identified by calculating the residuals above a locally defined threshold. The peaks component was defined as the residuals after subtracting background values from the interpolated series, and charcoal peaks were identified by calculating a locally defined threshold value separating fire-related and non-fire-related variations in the peaks component (9). Only peaks that had a maximum charcoal count with a <5% chance of coming from the same Poisson distribution as minimum charcoal counts within the previous 75 y were considered, except for site 13, 20, and 24, where all peaks were counted due to the lack of the sample volume information required to perform the minimum count test (9, 10). Peak magnitudes were obtained by calculating the positive deviations above the background. Ratios of arboreal to non-arboreal pollen percentages (AP/NAP) were obtained by dividing the sum of arboreal and shrub pollen percentages (AP) by the sum of the total terrestrial pollen percentage [AP/(AP + NAP)]. Changes in AP were used as an indicator of major changes in woody fuel levels, not as a tool for reconstructing detailed changes in vegetation community composition, which is beyond the scope of this article.

Trends in Charcoal Influx, Peak Frequency, and Arboreal Pollen

The estimation of trends in noisy data like the charcoal influx data involves a tradeoff between (i) fitting a relatively simple model, like a straight line or polynomial, which allows assessment of the significance of the trend to be made (11) and (ii) using a more flexible or "data-adaptive" model which may better represent more complicated or nonlinear forms of a trend, but which makes it harder to establish the overall significance of the trend (12). We use 2 approaches here: (i) a piecewise linear or segmented-regression model, which allows some flexibility in the fitted model, in particular changes in slope and intercept at some (possibly unknown) breakpoints, and (ii) a local regression or "lowess" approach, which makes no assumptions about the form of the overall trend.

The charcoal influx data were first transformed using the Box–Cox transformation to stabilize the variance of the data as described in Power et al. (13). The transformed values were converted to Z scores by subtracting the mean value and dividing by the standard deviation using a base period of 15–10 ka to allow comparisons among the records that feature widely varying average charcoal influx rates.

We used the "segmented" package (14) from the R-Project (15) to fit an overall linear trend to the charcoal influx data, allowing for changes in the slope and intercept of the trend line at several breakpoints, which were simultaneously estimated with the trend. There is a tradeoff between the number of breakpoints (and the length of the intervals they define) and the interpretability and robustness of the results. Too few breakpoints may lead to a less-good fit to the data, and greater heterogeneity of the intervals or episodes that are defined, whereas too many breakpoints lead to more complicated ad hoc interpretations of the results and to greater sensitivity of the results to the specific data being analyzed. We explored linear and polynomial (2nd- and 3rd-order) trends, and 2–4 break- points, with starting values for the breakpoints at even 1,000-y intervals from 11,000 to 14,000 y BP.

The best-fitting model with the fewest parameters was a segmented straight-line model with breakpoints at 12,820 y BP (SE = 128.0 y) and 11,550 yr BP (SE = 162.3 y). Because these breakpoint ages are indistinguishable from the beginning and end of the YDC (12,875 y BP and 11,660 y BP, respectively), we refit this model using the latter values as breakpoints using ordinary least squares "dummy variable" regression. This model is:

$$\text{Influx} = \underset{[0.569100]}{5.241697} - \underset{[0.000053]}{0.000452} \bullet \text{Age} \ (\text{Age} < 11{,}660 \text{ y BP})$$

$$= \underset{[1.034000]}{0.797223} + \underset{[0.000084]}{0.000071} \bullet \text{Age} \ (11{,}600\text{–}12{,}875 \text{ y BP})$$

$$= \underset{[0.625000]}{6.610190} - \underset{[0.000045]}{0.000512} \bullet \text{Age} \ (>12{,}875 \text{ y BP}).$$

where the values in square brackets are the standard errors of the regression coefficients, and $F = 137$ ($P < 2.2 \times 10^{-16}$), $R^2 = 0.1647$. Note that the slopes of the line segments before (-0.000452) and after (-0.000512) the YDC are virtually identical, but fitting a model that constrains them to be so adds little to the efficiency of the model. Note also that the slope of the line segment during the YDC is not significantly different from zero. This model yields the straight-line segments on Fig. 9C, and demonstrates the statistical significance of the overall trend in charcoal influx over the LGIT as well as the absence of a trend during the YDC.

A local-regression or lowess curve (16) was also fit to the data to show the long-term trends unconstrained by the specification of a particular model of the trend. The lowess curve-fitting procedure used the tricube weight function with a fixed-width window of 200 y (100-y half-width) as opposed to a variable-width window that "spans" a fixed proportion of the data points. Fitted values were obtained at "target points" spaced 10 y apart. (Note that this interval is not an expression of our belief in the chronological precision of the data, but simply allows us to graph the fitted values in a reasonable way.) A robustness iteration was used to minimize the influence of unusual points or outliers. We also calculated bootstrap confidence intervals for the lowess curve (1,000 replications) where the sampling-with-replacement was done by sites as opposed to individual samples, to assess the impact of the inclusion or exclusion of specific sites in our dataset. The lowess fitted values appear as the smooth curve in Fig. 9C, and the 5th and 95th percentiles of the bootstrapped fitted values define the shaded bands. Note that the segmented regression trend model and the lowess curve describe the same general trend in charcoal influx during the LGIT.

The density of charcoal peaks in the high-resolution charcoal records (Fig. 9D) was displayed by using a kernel density estimator (17). We selected a bandwidth of 100 y, which provides a compromise between over-smoothing the peak frequencies while still displaying local maxima in peak frequencies that are supported by peaks in multiple individual records. Bootstrap confidence intervals were obtained in the same way as for the influx data.

AP proportions were transformed using the "angular" or arcsine transformation, and a composite curve (Fig. 9E) was constructed by smoothing the transformed data in a similar fashion as the charcoal influx data. However, because the temporal resolution of the pollen data are typically less than that of the charcoal data, we used a larger window width (200-y half-width) to smooth these data. Bootstrap confidence intervals were again obtained as for charcoal influx.

The Increase in Charcoal at 13.2 ka

The charcoal increase at 13.2 ka is evident in 14 of the 33 sites recording fires by 13.1 ka (sites 1, 2, 10, 15, 17, 18, 19, 20, 21, 22, 23, 26, 30, and 35; Fig. 16) from 8 different regions. These sites span an elevation range of 8–2,863 m, with 5 sites located above 2,000 m. Similar increases in charcoal influx occurred previously at 3 sites (sites 2, 19, and 21), so the change was unprecedented in only 11 records. Of these 11 records, 13.2 ka marks the beginning of a discrete peak at 7 of them (sites 1, 15, 17, 18, 23, 26, and 30), versus an increase in baseline levels at the remaining 4 sites (sites 10, 20, 22, and 35). Fire frequency also increased to a local maximum at 13.2 ka, after a peak in AP at 13.4 ka (Fig. 10). In contrast, 20 sites show low charcoal influx at 13.2 ka, illustrating that burning was widespread, but not continent-wide at the time.

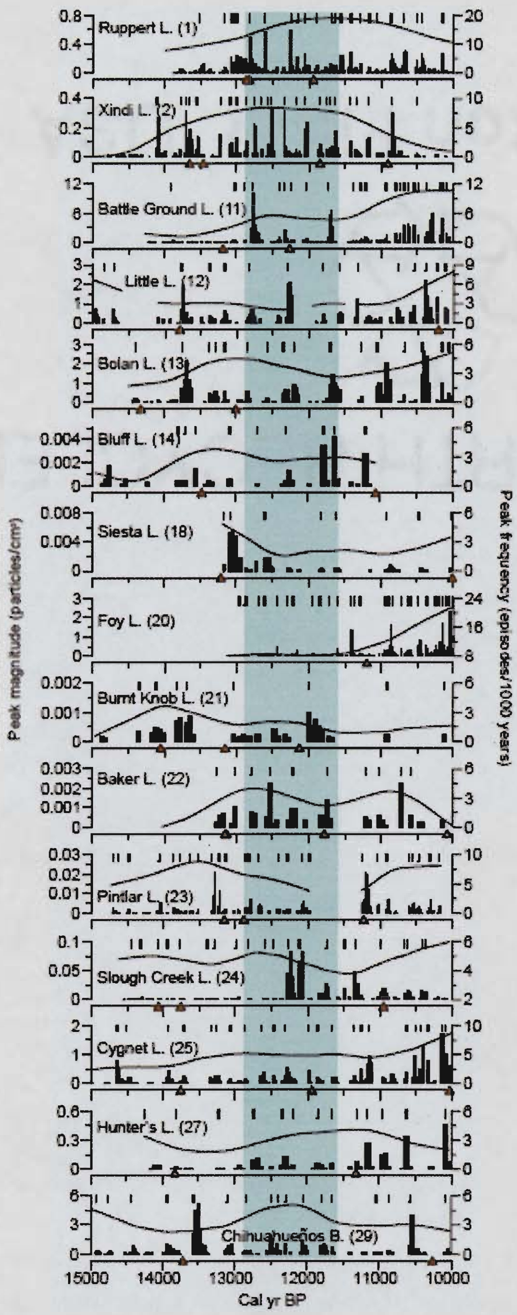

Fig. 17. Fifteen high-resolution paleofire records. Each graph shows changes in charcoal peak magnitude (black) and fire-episode frequency (smoothed black line) during deglaciation. Radiocarbon dates are represented by orange triangles.

APPENDIX C

SUPPLEMENTARY INFORMATION TO CHAPTER IV

The following supplementary material includes 1) information about data sources, including a list of the additional 36 charcoal records included in GCD v.2 and methodological details about our assessment of data coverage in climate and vegetation space; 2) details about the age controls and chronological precision of the records in the database; 3) an evaluation of potential factors affecting charcoal influx variations; 4) details about the reconstructed and simulated climate datasets used in the comparisons; and 5) the global, zonal, and regional biomass burning reconstructions for regions with sufficient sites along with associated statistics for each composite record.

Data Sources

We obtained 370 sedimentary charcoal records covering part or all of the last two millennia from the Global Charcoal Database (GCD version 1)[1]. The GCD contains charcoal records from different types of sites; we excluded records from marine sediments, which have unknown charcoal source regions, and alluvial fans and soils because these typically have poor temporal resolution and may reflect extremely local fires and not general biomass burning level. We also excluded charcoal records from archaeological sites because these typically reflect fuelwood use as opposed to fires in the surrounding landscape. To improve the regional coverage, an additional 36 sites (Table 2) were obtained from members of the Global Palaeofire Working Group or digitized from the published literature and have been incorporated into a new version of the GCD (GCD version 2).

Table 2. Additional charcoal records used in the global study with their GCD record locator (Site #). These data are available in GCD version 2: http://www.bridge.bris.ac.uk/projects/QUEST_IGBP_Global_Palaeofire_WG.

Number	Site ID	Site Name	Latitude	Longitude
1	417	Artxilondo	43.03	-1.14
2	418	Neublans	46.92	5.34
3	419	Chavannes	46.84	2.37
4	420	Pla de l'Orri	42.50	1.89
5	421	Cuguron	43.10	0.54
6	422	Peyre	44.99	2.72
7	483	Supulah Hill	-4.07	138.58
8	484	Wanda	-2.33	121.23
9	554	Canal de la Puntilla	-40.95	-72.90
10	597	Allom Lake	-25.23	153.17
11	598	Lake Sibaya	-27.21	32.37
12	600	Funduzi	-22.86	30.89
13	610	Lake Teletskoye	25.23	87.65
14	611	Gribljе Marsh	45.57	15.28
15	615	Mlaka	45.50	15.21
16	616	El Tiro Bog	-3.84	-79.15
17	621	Boa-1	19.07	-71.03
18	622	Laguna Azul	-52.12	-69.52
19	623	Crevice Lake	45.00	110.58
20	624	Les Comailles	47.66	3.22
21	672	Mizorogaike	35.06	135.77
22	673	Jagaike	35.24	135.46
23	674	Hatchodaira	35.23	135.83
24	676	Ofuke	38.65	135.18
25	682	Bolshoe bog	51.47	104.50
26	683	Duliha bog	51.52	105.00
27	684	Cheremushuka bog	52.75	108.08
28	685	Duguldzeiri River bog	54.45	109.53
29	686	Tompuda bog	55.13	109.77
30	699	Maralay Alas	63.10	130.58
31	700	Sugun Lake	62.08	129.48
32	701	Chai-ku Lake	62.00	130.07
33	712	Lake Lucerne	47.05	8.59
34	713	Bereket Basin	37.55	30.30
35	714	Lago di Pergusa	37.52	14.30
36	716	Xishuangbannan	21.50	101.50

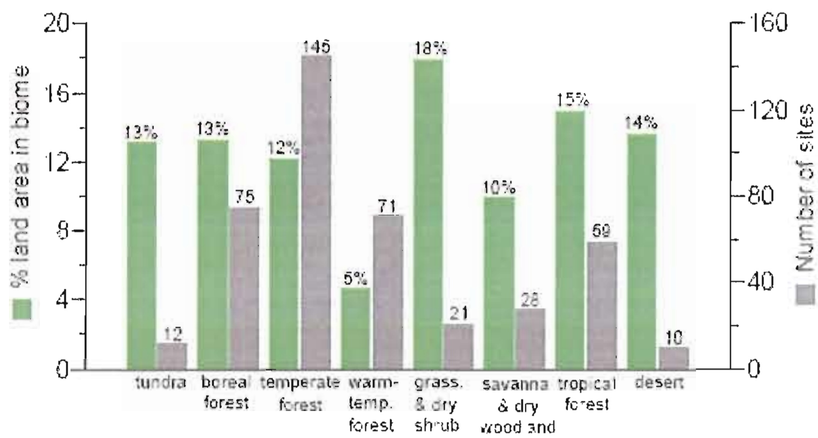

Fig. 18. Percentage of non ice-covered land are in each biome, and proportion of charcoal sites in each biome.

The distribution of charcoal sites was assessed in terms of climate and vegetation and was found to be reasonably representative of all major climatic zones and forest biomes (Fig. 13, main text). The climate at each sampling site and at each point on a global 0.5-degree grid, expressed in terms of growing-season warmth (growing degree days, GDD0, the accumulated temperature sum above 0°C) and plant-available moisture (the ratio of actual evapotranspiration to equilibrium evapotranspiration), was estimated from a climatology for 1961-1990 (CRU CL 2.0)[2] using algorithms embedded in the BIOME4 global equilibrium terrestrial biogeography-biogeochemistry model[3] and soil properties. We used the model-simulated biome at each site (CLIMATE 2.1) from BIOME4, which has been extensively tested and shown to produce a good simulation of vegetation distribution globally, to represent the natural vegetation cover. The 27 biomes distinguished by the model were grouped into broader units[4]. Estimates of percentage tree cover on the 0.5-degree grid based on remotely-sensed data[5] provided an alternative, independent representation of the actual vegetation cover.

Age Control and Chronological Precision of Charcoal Records

Records in the GCD have age models in calendar years, based on calibration of the original radiocarbon dates at each site using the Fairbanks et al. (2005)[6] calibration program (http://radiocarbon.Ideo.columbia.edu/research/radcarbcal.htm) and a standardized protocol for deriving the best-fit age model. The same procedure was adopted for the 31 new sites added to our data set. Specifically, (a) the core top was considered modern unless the original authors indicated that the surface was disturbed, (b) modern was assumed to correspond to an age of -50 cal yr BP (2000 AD) except in cases where the original publications specifically assigned a date (usually 0 cal yr BP, or 1950 A.D.). Age models were constructed using all available ages, including dated tephra layers and pollen stratigraphic ages, and were based on four possible age model styles: (1) linear interpolation, (2) a polynomial constrained to pass through zero, (3) an unconstrained polynomial fit, and (4) a cubic smoothing spline. The "best fit" age model was selected for each record, based on goodness-of-fit statistics and the appearance of the resulting curve (i.e., ensuring that new models did not introduce age reversals)[1].

Data Standardization and Normalization

We used the following procedure to limit the influence of sedimentation rate changes on the composite charcoal records (see Evaluation of Potential Factors Influencing Charcoal Influx Variations below), and to ensure comparability among a broad range of data types and charcoal quantification methods:

(1) All non-influx data (e.g., concentration expressed as particles/cm^3/yr; charcoal-to-pollen ratios) were converted to influx values (i.e. particles/cm^2/yr) or quantities proportional to influx, by dividing the charcoal values by sample deposition times (yr/cm).

(2) The variance of individual records was homogenized using the Box-Cox transformation:

$$c_i^* = \begin{cases} ((c_i' + \alpha)^{\lambda} - 1)/\lambda & \lambda \neq 0 \\ \log(c_i' + \alpha) & \lambda = 0 \end{cases}$$

where c_i^* is the transformed value, λ is the Box-Cox transformation parameter and α is a small positive constant (here, 0.01) added to avoid problems when c_i' and λ are both zero. The transformation parameter λ was estimated by using a maximum likelihood procedure[7], and was based on all charcoal values in each record spanning the past 2050 years.

(3) Because the Box-Cox transformation results in values that are comparable only among data sets with identical λ values, the transformed data were rescaled to the range (0, 1) using the minimax transformation:

$$c_i' = (c_i - c_{\min})/(c_{\max} - c_{\min})$$

where c_i' is the transformed value of the i-th sample in the record and c_i, and $c_{\max}$ and $c_{\min}$ are the maximum and minimum values over the past 2050 years.

(4) The transformed and rescaled values were converted to Z-scores by subtracting the mean value and dividing by the standard deviation of the past 2050 years, to express the data as anomaly values, or differences from a long-term average.

Data Smoothing and Construction of the Composite Records

To construct the global and regional composite charcoal records, we tested several alternative processes (Fig. 19) and adopted a two-stage smoothing method. Individual records were "pre-smoothed" to ensure that records with unusually high sample resolution (e.g. Crawford Lake[8]) did not have a disproportionate influence in the composite record. Smoothing was performed using locally weighted regression, or "lowess"[9]. The lowess approach minimizes the influence of outliers, which helps filter noise from the charcoal data. In implementing lowess, we used a constant window width and fixed target points (in time), as opposed to a constant proportion or span of data points and fitting at each data point[10]. The lowess function used the customary tricube weight function, a first-degree or linear fit at each target point, and a single "robustness iteration." The presmoothing window width was 20 years, with a fit of order 0 (i.e., a locally weighted mean) and a robustness parameter of 0, thus including all data values that fall within the window when calculating the local mean. A single value was thus produced for each non-overlapping 20-year interval from 50 BC to 2000 AD, ensuring that the value for each 20-year window was based on all samples (for records with high temporal resolution) and also that data were not interpolated within low-resolution segments of records or extrapolated into missing segments of records. The lowess technique was also used to create the composite charcoal curves (Fig. 11a), based on a 200-yr window that emphasizes centennial-scale variations in the charcoal data. The smooth curves shown here are constructed by determining fitted values at 10-year intervals. (Window-widths here refer to the full span of the window, not the half-width).

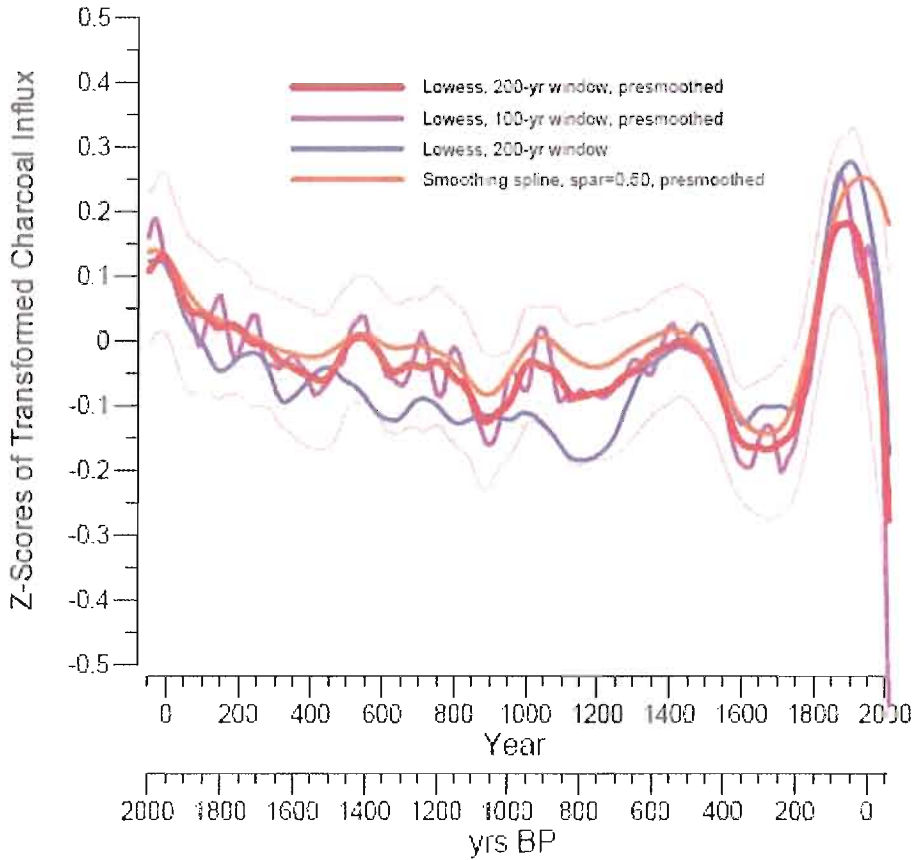

Figure 19. Alternative (global) composite curves, including fitted values determined using the presmoothed values and a 200-yr window (red), the presmoothed values and a 100-yr window (purple), the transformed and rescaled individual charcoal values (blue), and the fitted values from the application of a smoothing spline with a 0.50 smoothing parameter (orange).

Confidence intervals for the global and regional composites were generated by a bootstrap approach in which individual sites (not samples) were sampled with replacement over 1000 replications. The approach allows us to asses the sensitivity of the results to the inclusion or exclusion of individual sites. Bootstrap confidence intervals for each target point were the 2.5th and 97.5th percentiles of the fitted values for that target point. The number of samples, age control, and spatial distribution of records contributing to each summary series is illustrated in figures 23-32. We also assessed the degree to which the global composite record was influenced by the uneven spatial distribution of sites by examining alternative bootstrap approaches. We considered (1) sampling with replacement of individual samples (the conventional approach), (2) sampling with replacement of individual sites (the approach used here), and (3) stratified resampling based on sampling pre-defined regions (with replacement) and then sampling (with replacement) sites within the selected region (which upweights the influence of regions represented by fewer sites). The resulting curves (Fig. 20) fall within the confidence limits on our standard curve, have a similar temporal structure, and identify the same century-scale local maxima. This strongly suggests that the global summary curve is not appreciably influenced by the uneven distribution of sites among regions. In addition, we examined the relationship between the global composite records and potential factors affecting charcoal influx variations, such as sedimentation rates (Fig. 21) and found the composite to be robust to such influences.

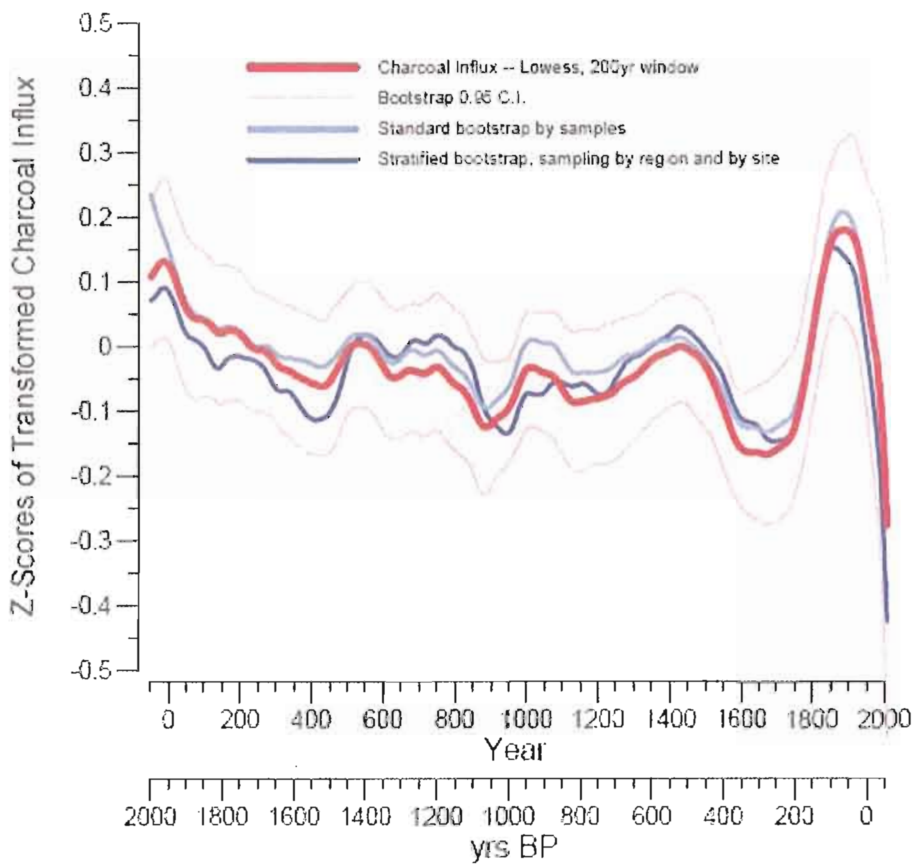

Figure 20. Evaluation of the impact of site distribution on the composite records.

Evaluation of Potential Factors Affecting Charcoal Influx Variations

In paleofire research in general, the possibility of changing sedimentation rates in lake cores motivates the examination of charcoal influx, as opposed to charcoal concentrations. However, the remarkable charcoal influx variations in the last few centuries warranted an examination of the effects of using charcoal concentration as opposed to influx, of varying sedimentation rates, and possible uncertainties in age model construction on the composite curves. Sedimentation rates are artificially high near the top of lake-sediment cores because of the high water content and possibly higher erosion rates after land-cover changes, and thus we evaluated the possibility that features observed in the charcoal curve in Fig. 11 (main text) were artifacts of these processes.

A simple model was created to assess the impact of changing sedimentation rates and down-core sediment dewatering on charcoal concentration and influx data (P.J. Bartlein and D.G. Gavin, unpublished results). Figure 21 shows composite curves across sites based on charcoal concentration, influx, and individual-sample sediment rates (all smoothed the same way as the charcoal influx data in Fig. 11, main text). Sedimentation rates show the expected increase over time, while charcoal concentration and influx show the characteristic upturn around 1750 A.D. and downturn after 1850 A.D. The only way the model can generate features similar to those observed in the influx data is by manipulating sedimentation rate or concentration in unrealistic ways.

Comparison of concentration data, influx data, and sedimentation rates (Fig. 21) shows that the long-term trend in the concentration data inversely reflects the long-term trend in sedimentation rate, which increases towards present. The distinct downward trend in the influx data cannot be accounted for by changes in sedimentation rate. Centennial-scale variability in both concentration and influx shows no consistent relationship with changes in sedimentation rate. The peak in charcoal concentration and influx at ca. 1850 A.D. (Figure 11, main text, Fig. 21) is coincident with a steady increase in sedimentation rate and is therefore not an artifact due to changes in sedimentation rates.

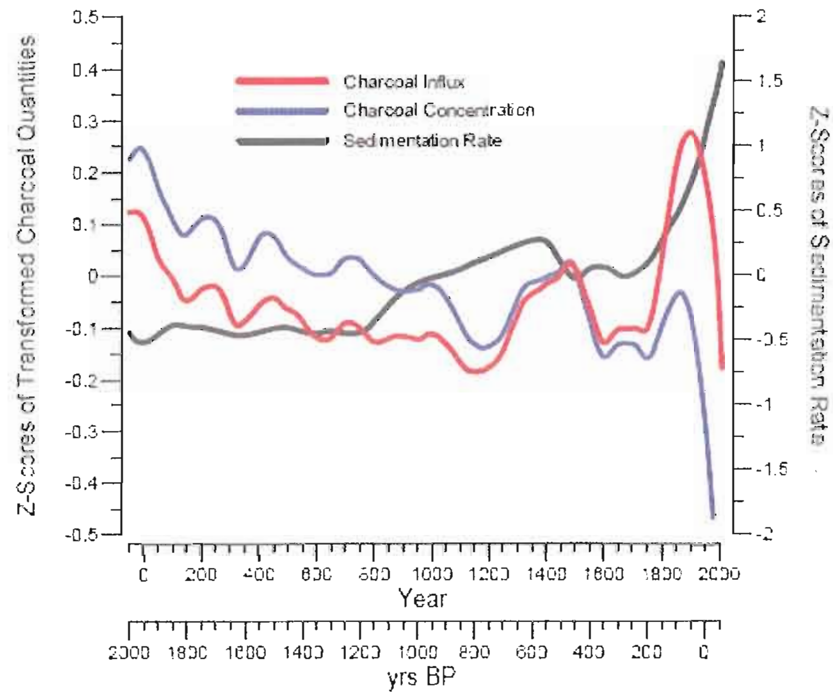

Fig. 21. Smoothed values of charcoal concentration, influx, and sedimentation rate. The concentration, influx, and sediment rate of individual samples were summarized by curves constructed by locally weighted regressions (i.e. "lowess curves") using a constant window width of 200 years, the tricube weight function, and one "robustness iteration."

The variations in charcoal accumulation are thus highly robust to differences in pre-treatment methods. However, additional factors increase the uncertainty of the composite reconstructions for recent decades, including the assignment of an age for the "modern" core top (i.e. 1950 A.D. or 2000 A.D.), the use of a different method (Pb210 versus 14C) to date recent decades, the high water content of surface samples, and potentially large changes in erosion and sediment disturbance due to human activities. To highlight the increased uncertainty in our reconstructions in recent decades, changes in biomass burning post-1950 are shown as a dashed line. Other sources of data, e.g. forestry records, fire-scar or tree-ring records, or remote-sensing, are likely to provide more accurate tools for reconstructing changes in fire regimes than sedimentary records over this period[3].

Comparison with NCAR Climate Simulations

Climate-model simulations provide an alternative realization of long-term climate changes. We have compared the composite charcoal record to results from a 1000-year simulation[11] with the National Center for Atmospheric Research Climate System Model (NCAR CSM), Version 1.4, a global coupled atmosphere-ocean-sea ice-land surface model without flux adjustments[12, 13]. The NCAR CSM1.4 model was forced over the period from 850 to 2000 A.D. using observation-based records of solar irradiance, spatially explicit aerosol loading from explosive volcanism, the greenhouse gases CO2, CH4, N2O, CFC-11 and CFC-12, and anthropogenic sulfate aerosols, plus a recurring annual cycle of ozone and natural sulfate aerosol. Other forcing factors, such as orbital parameter changes (very slow forcing, constant 1400 A.D. conditions were used) and vegetation/land use changes (highly regional in impact), were not included. The solar irradiance history over the past 1150 years was based on a recent 10Be history and solar irradiance was scaled to a reduction during the Maunder Minimum (1645-1715 A.D.) of 0.25% relative to present[14]. The volcanic forcing was established by converting ice-core aerosol proxies to latitudinal and temporally varying atmospheric aerosol fields. Atmospheric CO2, CH4 and N2O concentrations were individually prescribed based on ice-core measurements and direct atmospheric observations. The CFC-11, scaled to take into account the radiative forcing by other halocarbons and SF6[15], and CFC-12 concentration histories are based on historic emission data and recent measurements. The results are given as detrended residuals after subtracting a millennial-scale spline fit for individual months of the annual cycle at each model grid point obtained from a control integration. Further details of these simulations (model setup and simulated climate) are available elsewhere[11].

We compared the composite global fire reconstruction for the past 1000 years with simulated global temperature and precipitation from the NCAR CSM1.4 simulation. There is broad similarity between the simulated temperature record and the charcoal-based fire record (Fig. 22) over much of this interval. However, in general the simulated changes are smoother than those shown in the global fire reconstruction and the feature corresponding to the "Little Ice Age" interval in the fire reconstruction is more distinct. The marked upward trend in fire activity after ca 1750 A.D. corresponds to a marked upward trend in the simulated temperature curve, supporting our suggestion that climate change could have contributed to the observed increase in fire. The most striking disparity between the two records is post-1950 when charcoal decreases while climate continues to warm. This difference supports our suggestion that this feature of the composite global fire reconstruction is driven by direct human action rather than by climate. There is no clear relationship between the composite global fire reconstruction and simulated precipitation changes. This suggests that the principal climatic driver for changes in biomass burning at a global scale is changes in temperature, rather than precipitation.

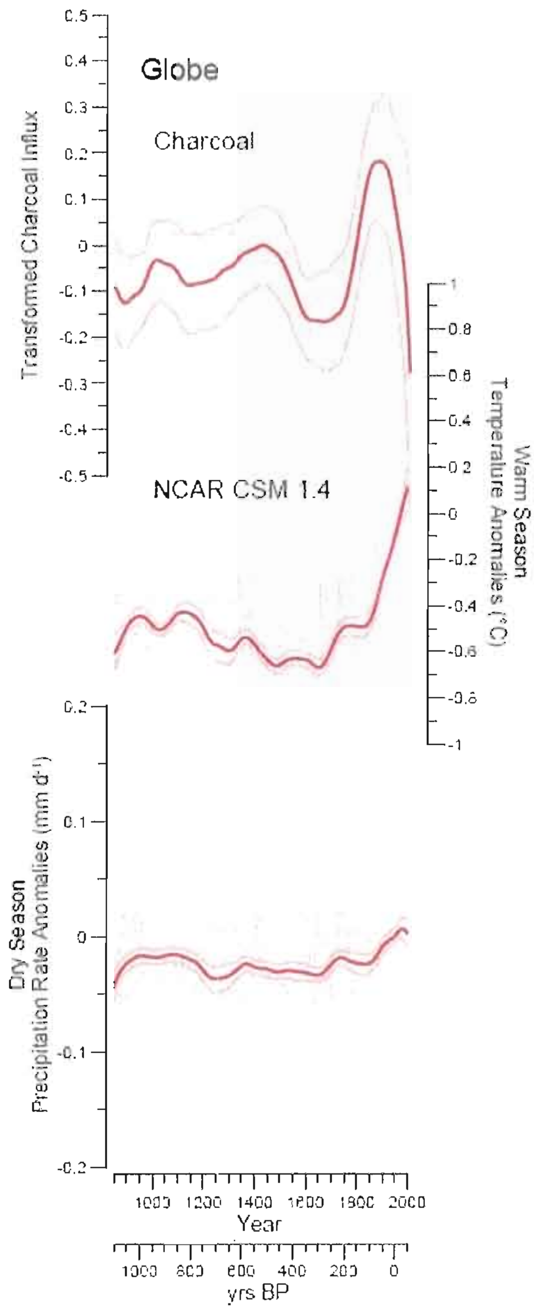

Fig. 22. (A) Reconstruction of global biomass burning, with NCAR CSM1.4 simulated (B) warm season temperature anomalies, and (C) dry season precipitation rate anomalies. Simulated data are for land only and are also shown smoothed and with 95% confidence intervals in the same manner as the charcoal records.

Global, Zonal, and Regional Biomass Burning Reconstructions and Associated Statistics

The global composite reconstruction of biomass burning (Fig. 23b) is based on transformed, rescaled, and standardized charcoal values (Fig. 23a). Although the number of sites contributing to the global curve varies through time, there are never less than 80 sites comprising the average for each time interval (Fig. 23c). Most records in the global composite have a sampling resolution greater than one sample per 50 years, and many have a resolution of one sample per 10-20 years (Fig. 23f). There is reasonable dating control throughout the period of interest and only five of the 20-year time windows lack radiometric dates (Fig. 23d). These analyses suggest that the variations in the global composite are unlikely to be artifacts related to sample density.

The robustness of the zonal reconstructions of fire history (Fig. 12, main text) can be assessed on the basis of figures 25 to 28. As might be expected given the land area within each zone (Fig. 12, main text), the number of sites contributing to the reconstruction, sampling density, and age control are all higher for the northern hemisphere extratropics than for other zonal reconstructions. Nevertheless, there are more than 10 sites contributing to all zonal reconstructions in nearly every time interval, and the sampling resolution and dating control are sufficient to ensure confidence in the reconstruction of long-term trends and centennial-scale variability. These reconstructions are discussed in the main text.

We examined composite records of fire history for a number of regions, defining regions on the basis of broad-scale atmospheric-circulation controls on modern and past climate regimes. Some regions have too few sites to provide a coherent picture of regional changes. Here, we present results for regions with more than 10 sites contributing to most 20-year windows during at least the past 1000 years (Figs. 29-32). The sample size per 20-year interval and the proportion of global ice-free land area are also shown.

The regional composites show that the initial long-term downward trend characteristic of the global composite fire history is expressed in western North America (Fig. 28), Asia (Fig. 31) and Central and tropical South America (Fig. 32). The trend is weak in the composite records for eastern North America (Fig. 29) and Europe (Fig. 30). The records from Asia (Fig. 31), like the zonal records of the tropics (Fig. 25) and southern extratropics (Fig. 26), suggest variable levels of biomass burning prior to ca 1800 AD. With the exception of Central and tropical South America (Fig. 32), all regions display a marked upturn in the second half of the 19th century, although the timing of this increase varies. The marked downturn during the first half of the 20th century is clear in the regional composite records from western North America (Fig. 28), Central and tropical South America (Fig. 32), and Asia (Fig. 31). Although there is a wealth of information in these regional fire history records, analyzing these patterns and their relationships with regional climate, population and land-use changes, is beyond the scope of this paper.

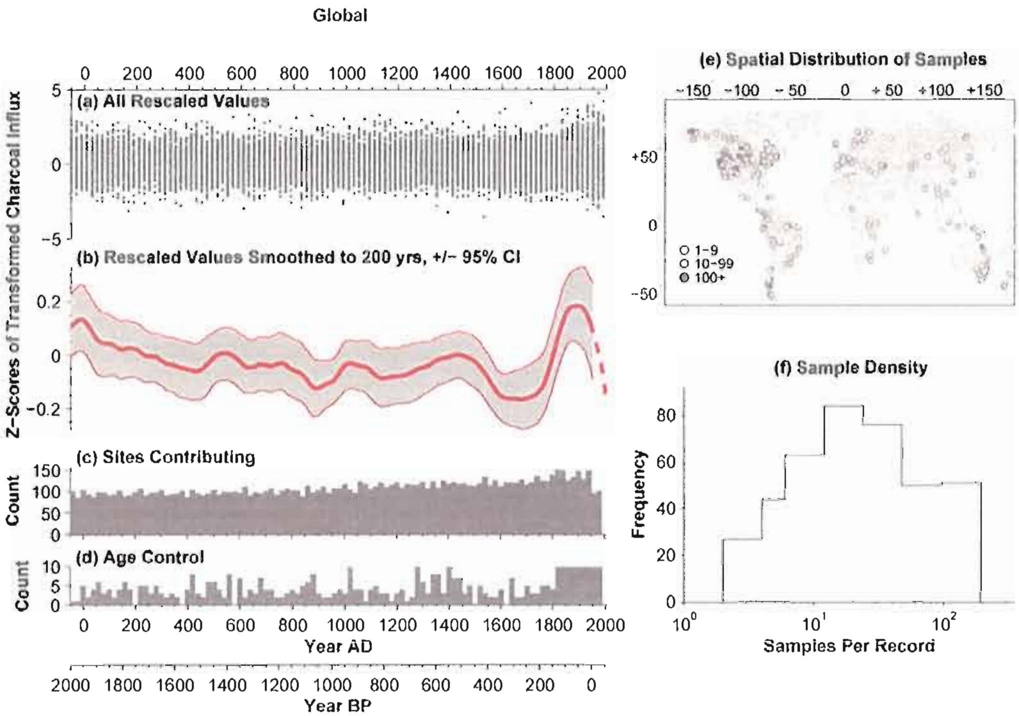

Fig. 23. Composite charcoal data for the globe. (A) Rescaled and transformed charcoal influx values on equal 20-year intervals. (B) Composite charcoal curve with confidence intervals. (C) Number of sites contributing at least one sample to each 20-year interval in the composite record. (D) Number of 14C or 210Pb dates contributing to chronological control for each 20-year interval in the composite record. (E) Spatial distribution of sites contributing to the composite record, with the number of samples at each site reflected by the shade of each dot. (F) Histogram of sample density within each individual record contributing to the composite record. Note the log scale on the x-axis.

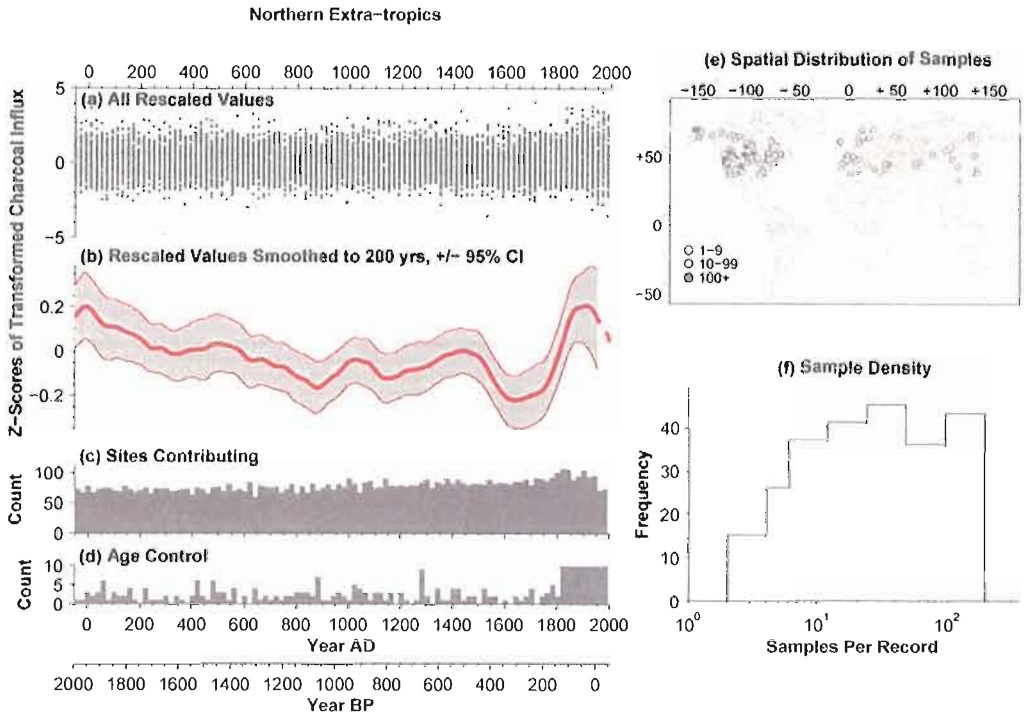

Fig. 24. Composite charcoal data for the northern extra-tropics. (A) Rescaled and transformed charcoal influx values on equal 20-year intervals. (B) Composite charcoal curve with confidence intervals. (C) Number of sites contributing at least one sample to each 20-year interval in the composite record. (D) Number of 14C or 210Pb dates contributing to chronological control for each 20-year interval in the composite record. (E) Spatial distribution of sites contributing to the composite record, with the number of samples at each site reflected by the shade of each dot. (F) Histogram of sample density within each individual record contributing to the composite record. Note the log scale on the x-axis.

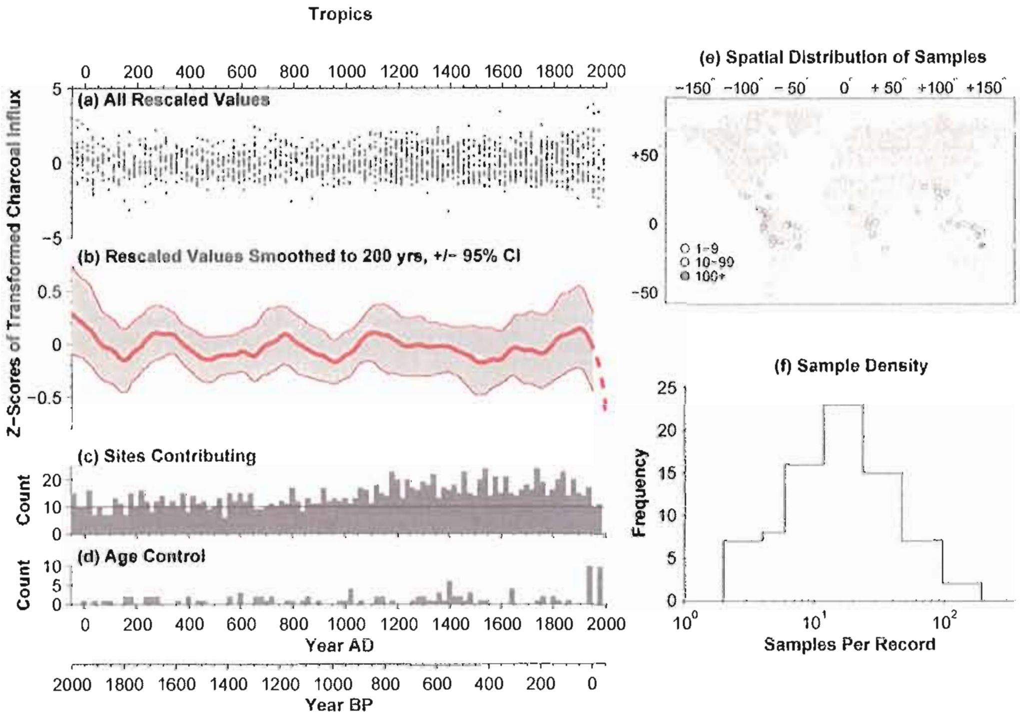

Fig. 25. Composite charcoal data for the tropics. (A) Rescaled and transformed charcoal influx values on equal 20-year intervals. (B) Composite charcoal curve with confidence intervals. (C) Number of sites contributing at least one sample to each 20-year interval in the composite record. (D) Number of 14C or 210Pb dates contributing to chronological control for each 20-year interval in the composite record. (E) Spatial distribution of sites contributing to the composite record, with the number of samples at each site reflected by the shade of each dot. (F) Histogram of sample density within each individual record contributing to the composite record. Note the log scale on the x-axis.

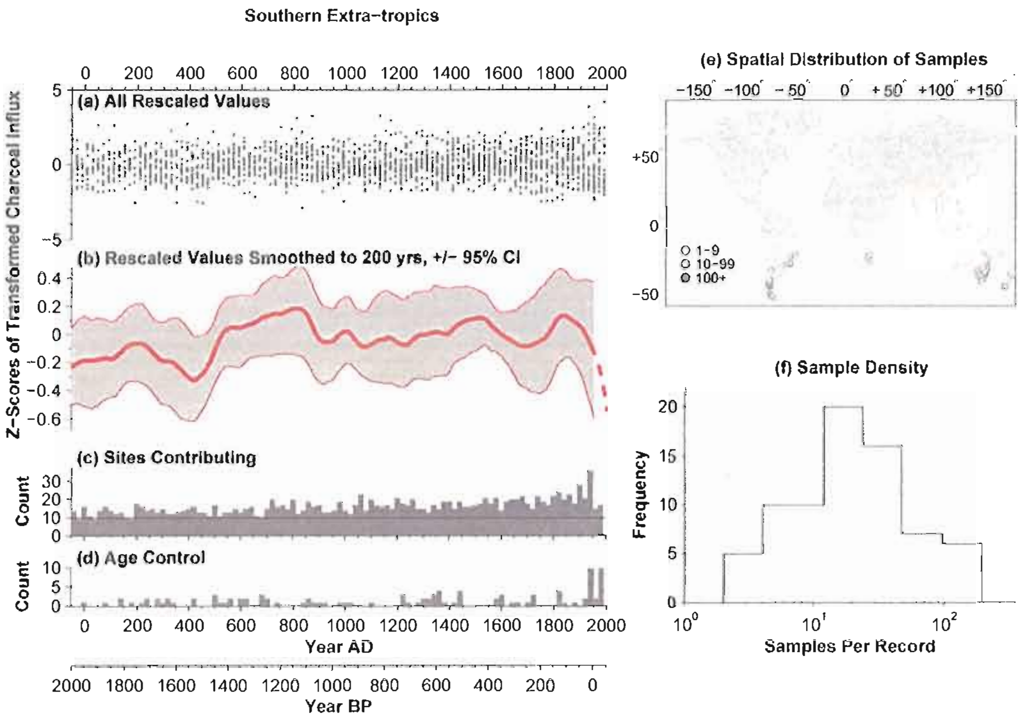

Fig. 26. Composite charcoal data for the southern extra-tropics. (A) Rescaled and transformed charcoal influx values on equal 20-year intervals. (B) Composite charcoal curve with confidence intervals. (C) Number of sites contributing at least one sample to each 20-year interval in the composite record. (D) Number of 14C or 210Pb dates contributing to chronological control for each 20-year interval in the composite record. (E) Spatial distribution of sites contributing to the composite record, with the number of samples at each site reflected by the shade of each dot. (F) Histogram of sample density within each individual record contributing to the composite record. Note the log scale on the x-axis.

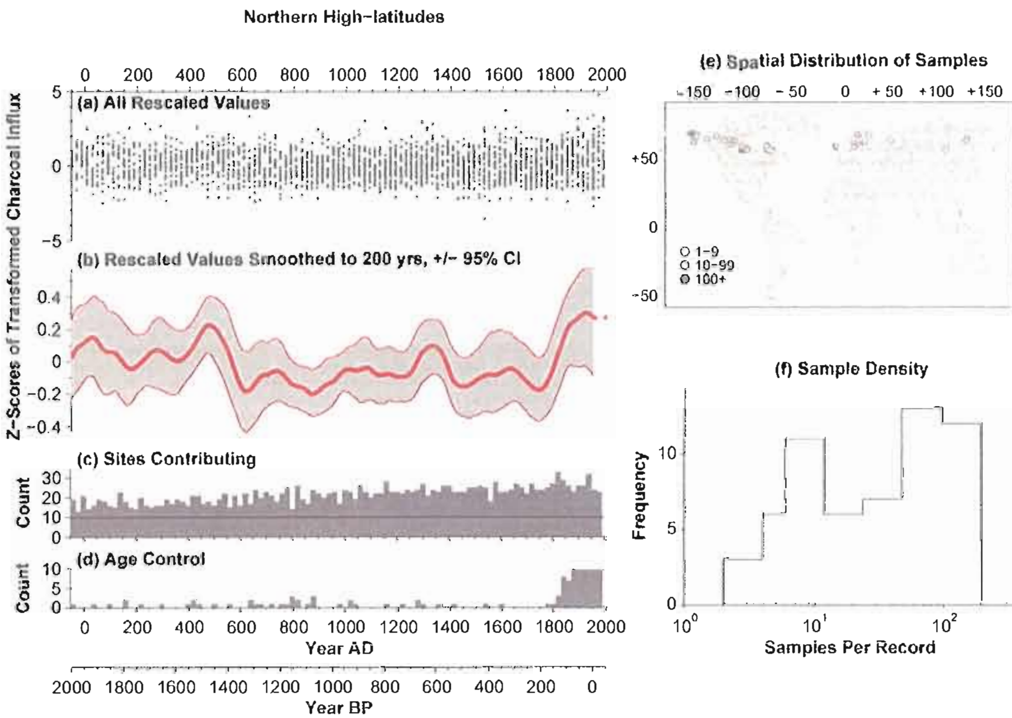

Fig. 27. Composite charcoal data for the northern high latitudes. (A) Rescaled and transformed charcoal influx values on equal 20-year intervals. (B) Composite charcoal curve with confidence intervals. (C) Number of sites contributing at least one sample to each 20-year interval in the composite record. (D) Number of 14C or 210Pb dates contributing to chronological control for each 20-year interval in the composite record. (E) Spatial distribution of sites contributing to the composite record, with the number of samples at each site reflected by the shade of each dot. (F) Histogram of sample density within each individual record contributing to the composite record. Note the log scale on the x-axis.

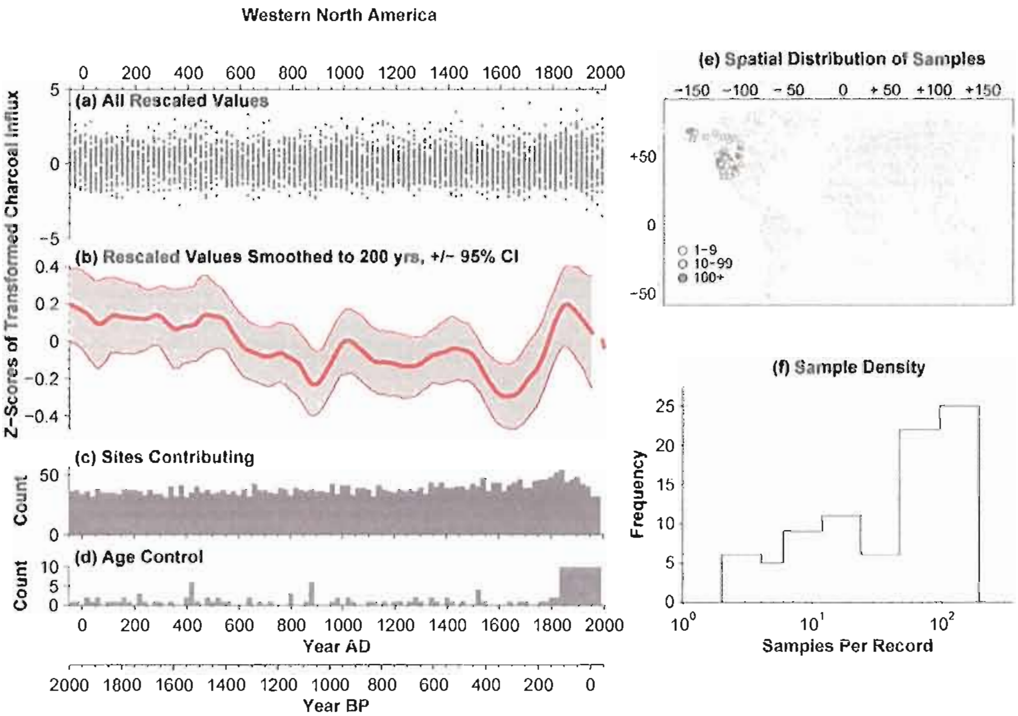

Fig. 28. Composite charcoal data for the western North America. (A) Rescaled and transformed charcoal influx values on equal 20-year intervals. (B) Composite charcoal curve with confidence intervals. (C) Number of sites contributing at least one sample to each 20-year interval in the composite record. (D) Number of 14C or 210Pb dates contributing to chronological control for each 20-year interval in the composite record. (E) Spatial distribution of sites contributing to the composite record, with the number of samples at each site reflected by the shade of each dot. (F) Histogram of sample density within each individual record contributing to the composite record. Note the log scale on the x-axis.

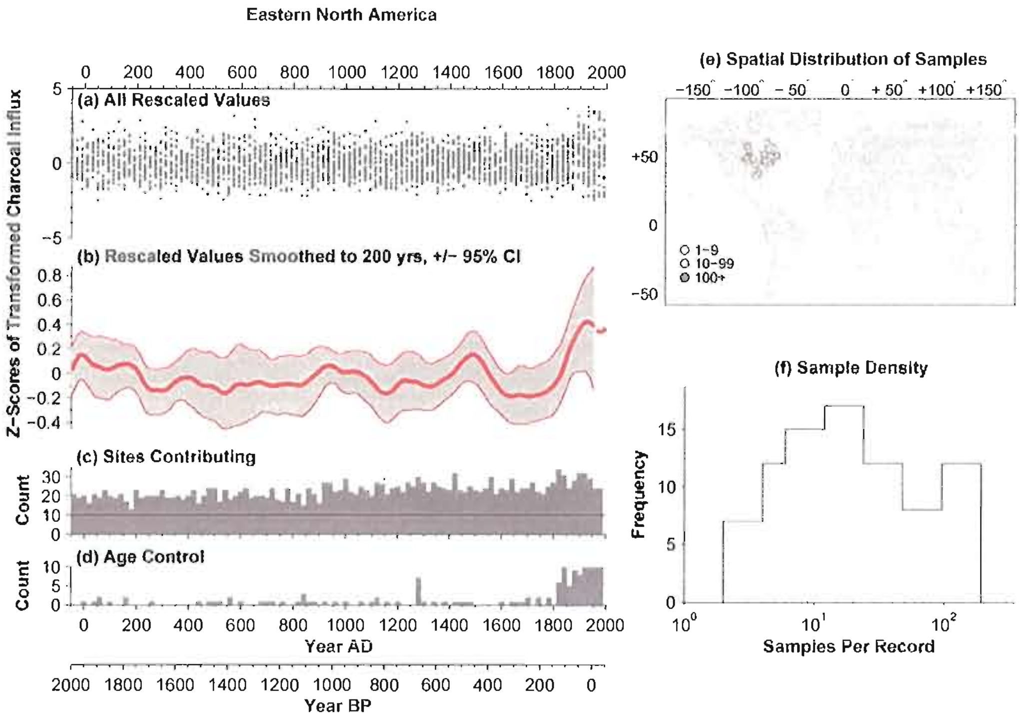

Fig. 29. Composite charcoal data for the eastern North America. (A) Rescaled and transformed charcoal influx values on equal 20-year intervals. (B) Composite charcoal curve with confidence intervals. (C) Number of sites contributing at least one sample to each 20-year interval in the composite record. (D) Number of 14C or 210Pb dates contributing to chronological control for each 20-year interval in the composite record. (E) Spatial distribution of sites contributing to the composite record, with the number of samples at each site reflected by the shade of each dot. (F) Histogram of sample density within each individual record contributing to the composite record. Note the log scale on the x-axis.

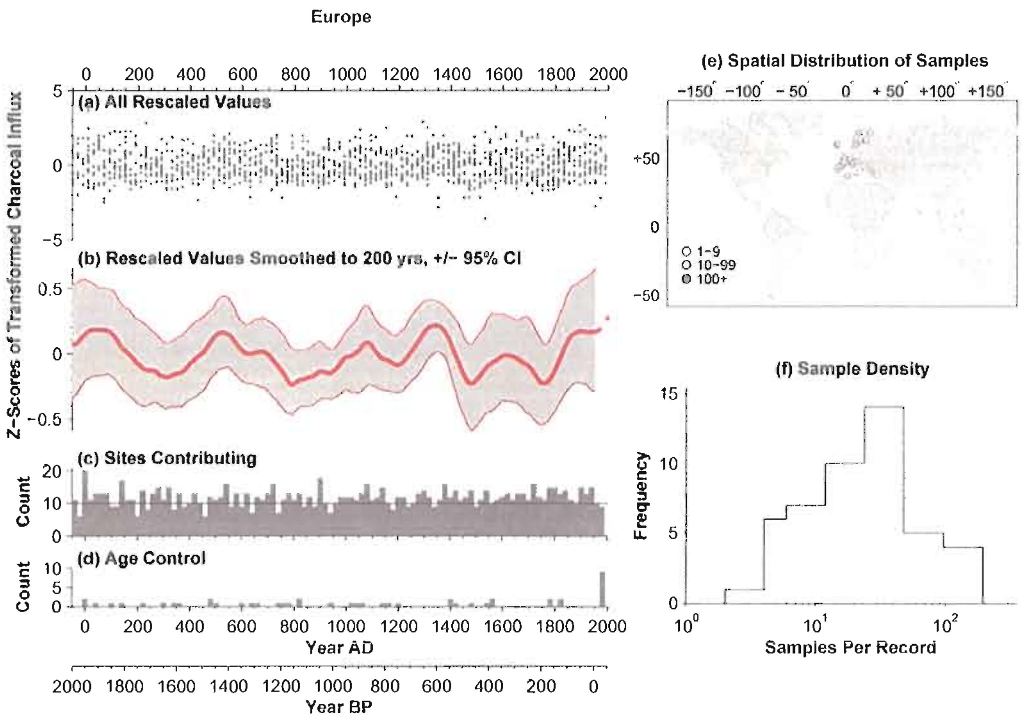

Fig. 30. Composite charcoal data for the Europe. (A) Rescaled and transformed charcoal influx values on equal 20-year intervals. (B) Composite charcoal curve with confidence intervals. (C) Number of sites contributing at least one sample to each 20-year interval in the composite record. (D) Number of 14C or 210Pb dates contributing to chronological control for each 20-year interval in the composite record. (E) Spatial distribution of sites contributing to the composite record, with the number of samples at each site reflected by the shade of each dot. (F) Histogram of sample density within each individual record contributing to the composite record. Note the log scale on the x-axis.

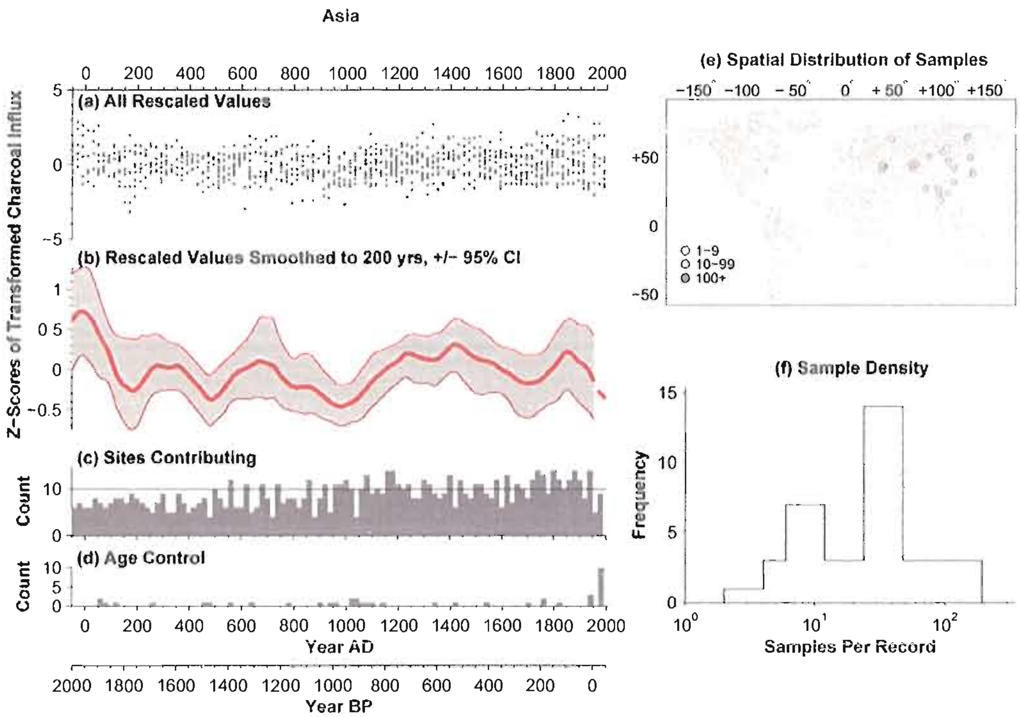

Fig. 31. Composite charcoal data for Asia. (A) Rescaled and transformed charcoal influx values on equal 20-year intervals. (B) Composite charcoal curve with confidence intervals. (C) Number of sites contributing at least one sample to each 20-year interval in the composite record. (D) Number of 14C or 210Pb dates contributing to chronological control for each 20-year interval in the composite record. (E) Spatial distribution of sites contributing to the composite record, with the number of samples at each site reflected by the shade of each dot. (F) Histogram of sample density within each individual record contributing to the composite record. Note the log scale on the x-axis.

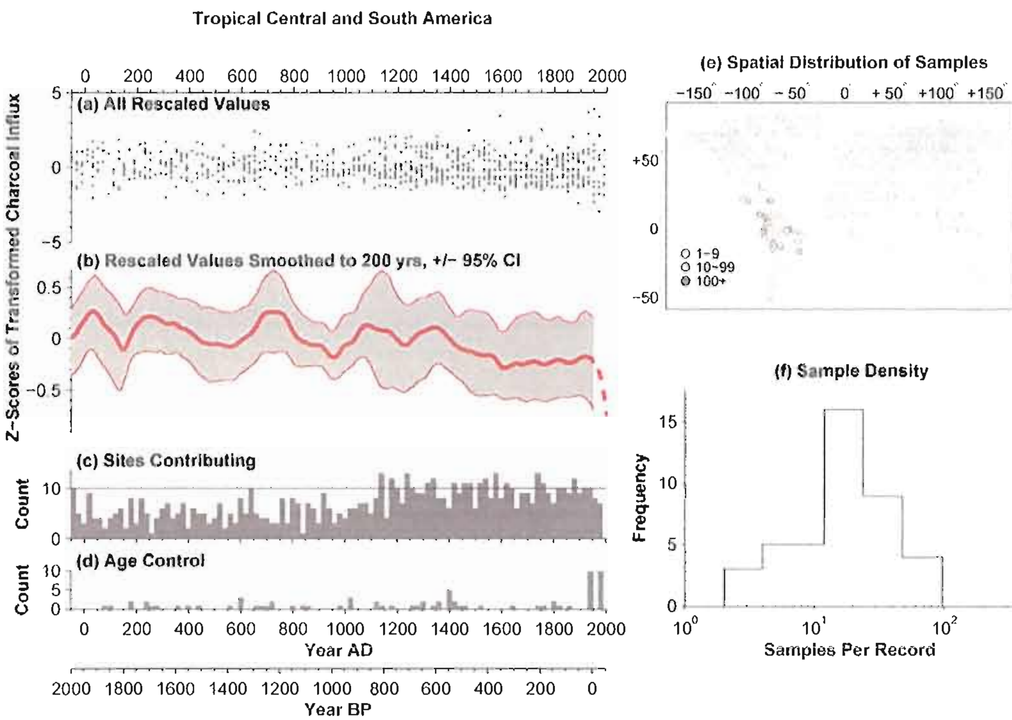

Fig. 32. Composite charcoal data for tropical Central and South America. (A) Rescaled and transformed charcoal influx values on equal 20-year intervals. (B) Composite charcoal curve with confidence intervals. (C) Number of sites contributing at least one sample to each 20-year interval in the composite record. (D) Number of 14C or 210Pb dates contributing to chronological control for each 20-year interval in the composite record. (E) Spatial distribution of sites contributing to the composite record, with the number of samples at each site reflected by the shade of each dot. (F) Histogram of sample density within each individual record contributing to the composite record. Note the log scale on the x-axis.

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Chapter III

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