Changes in fire regimes since the Last Glacial Maximum: an assessment based on a global synthesis and analysis of charcoal data

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Abstract Fire activity has varied globally and continuously since the last glacial maximum (LGM) in response to long-term changes in global climate and shorter-term regional changes in climate, vegetation, and human land use. We have synthesized sedimentary charcoal records of biomass burning since the LGM and present global maps showing changes in fire activity for time slices during the past 21,000 years (as differences in charcoal accumulation values compared to pre-industrial). There is strong broad-scale coherence in fire activity after the LGM, but spatial heterogeneity in the signals increases thereafter. In North America, Europe and southern South America, charcoal records indicate less-than-present fire activity during the deglacial period, from 21,000 to ∼11,000 cal yr BP. In contrast, the tropical latitudes of South America and Africa show greater-than-present fire activity from ∼19,000 to ∼17,000 cal yr BP and most sites from Indochina and Australia show greater-than-present fire activity from 16,000 to ∼13,000 cal yr BP. Many sites indicate greater-than-present or near-present activity during the Holocene with the exception of eastern North America and eastern Asia from 8,000 to ∼3,000 cal yr BP, Indonesia and Australia from 11,000 to 4,000 cal yr BP, and southern South America from 6,000 to 3,000 cal yr BP where fire activity was less than present. Regional coherence in the patterns of change in fire activity was evident throughout the post-glacial period. These complex patterns can largely be explained in terms of large-scale climate controls modulated by local changes in vegetation and fuel load.

Keywords Palaeoenvironmental reconstruction · Biomass burning · Palaeofire regimes · Charcoal · Data-model comparisons

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1 Introduction

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 3,000-year intervals since the Last Glacial Maximum (LGM, conventionally centered 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 preparation). 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, palaeo-ecological 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 ($n = \sim 50$–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 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.

The database currently contains charcoal records from 405 sites (Fig. 1), 355 of which are used in this analysis and 33 of which have records back to the LGM (here defined as $21,000 \pm 500$ calendar yr BP). Records of the mean charcoal value at a site over a 1,000-year long interval were extracted for seven time slices (i.e., 3,000; 6,000; 9,000; 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 $3,000 \pm 500$, $6,000 \pm 500$, $9,000 \pm 500$, $12,000 \pm 500$, $15,000 \pm 500$, $18,000 \pm 500$, and $21,000 \pm 500$ cal yr BP. These values

● = sites used in current analysis (GCD 15 Oct 2007: b16e.txt)
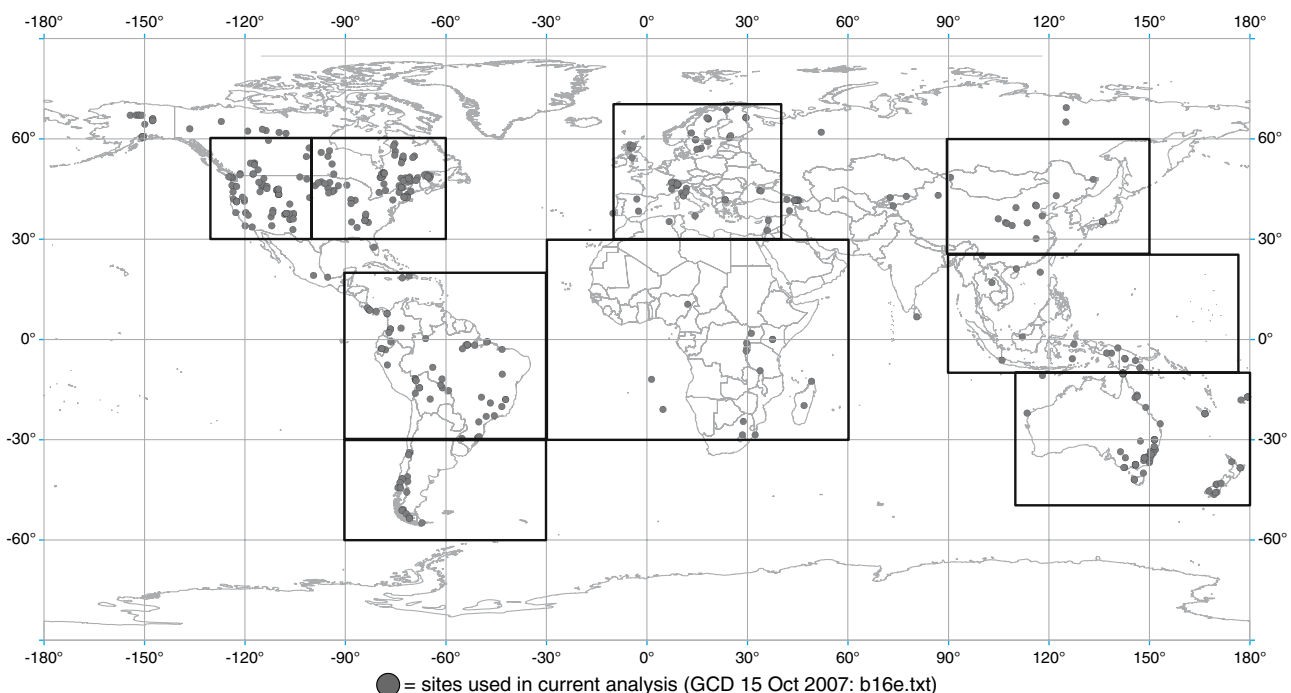
Fig. 1 Inventory of global charcoal records currently used in the analysis (*gray dots*) and regional delineations (*black boxes*) where sites were averaged together for regional summaries of fire regimes (see Fig. 5a)

were compared with an estimate of the modern (pre-industrial) fire regime, based on the mean charcoal value for the period 100–1,000 cal yr BP. The choice of a 1,000-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 1,000-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 1,000-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 4,000 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.Ideo.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 (2,000 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 2,003

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 ten orders of magnitude among and within sites (Fig. 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 Fig. 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_j'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:

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

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 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 (Fig. 3e)

Fig. 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
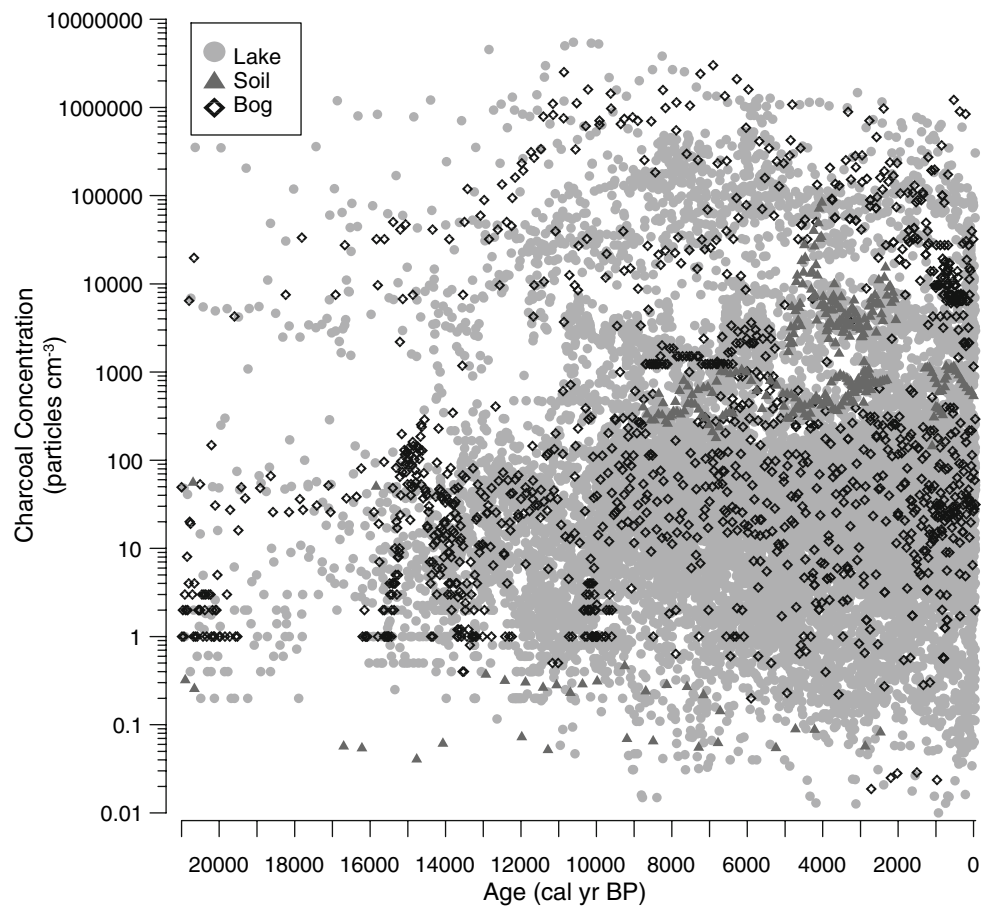
$$z_i = (c_i^* - \bar{c}_{(4\,\mathrm{ka})}^*)/s_{c(4\,\mathrm{ka})}^*$$

where, $\bar{c}_{(4\,\mathrm{ka})}^*$ is the mean minimax-rescaled and Box-Cox transformed charcoal value over the interval 4,000 to 100 cal yr BP, and $s_{c(4\,\mathrm{ka})}^*$ 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 (100–4,000 cal yr BP) was used to calculate the mean and standard deviation for each site. The choice of the last 4,000 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 4,000-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 1,000-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 1,000-year window and the mean Z score for the "modern", where "modern" is defined as the interval 1,000 to 100 cal yr BP).

2.3 Creation of 3,000-year steps of time-slice maps (1,000-year window)

Time-slice anomaly maps of modern (1,000 to 100 cal yr BP) mean Z score minus palaeo mean Z score were created at 3,000-year intervals (Fig. 4a-g) from 21,000 cal yr BP to present. The anomalies were classified into five roughly equal-frequency groups ranging from those > +1.15 (strong positive anomalies, dark red) +1.15 to +0.365 (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 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

Fig. 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
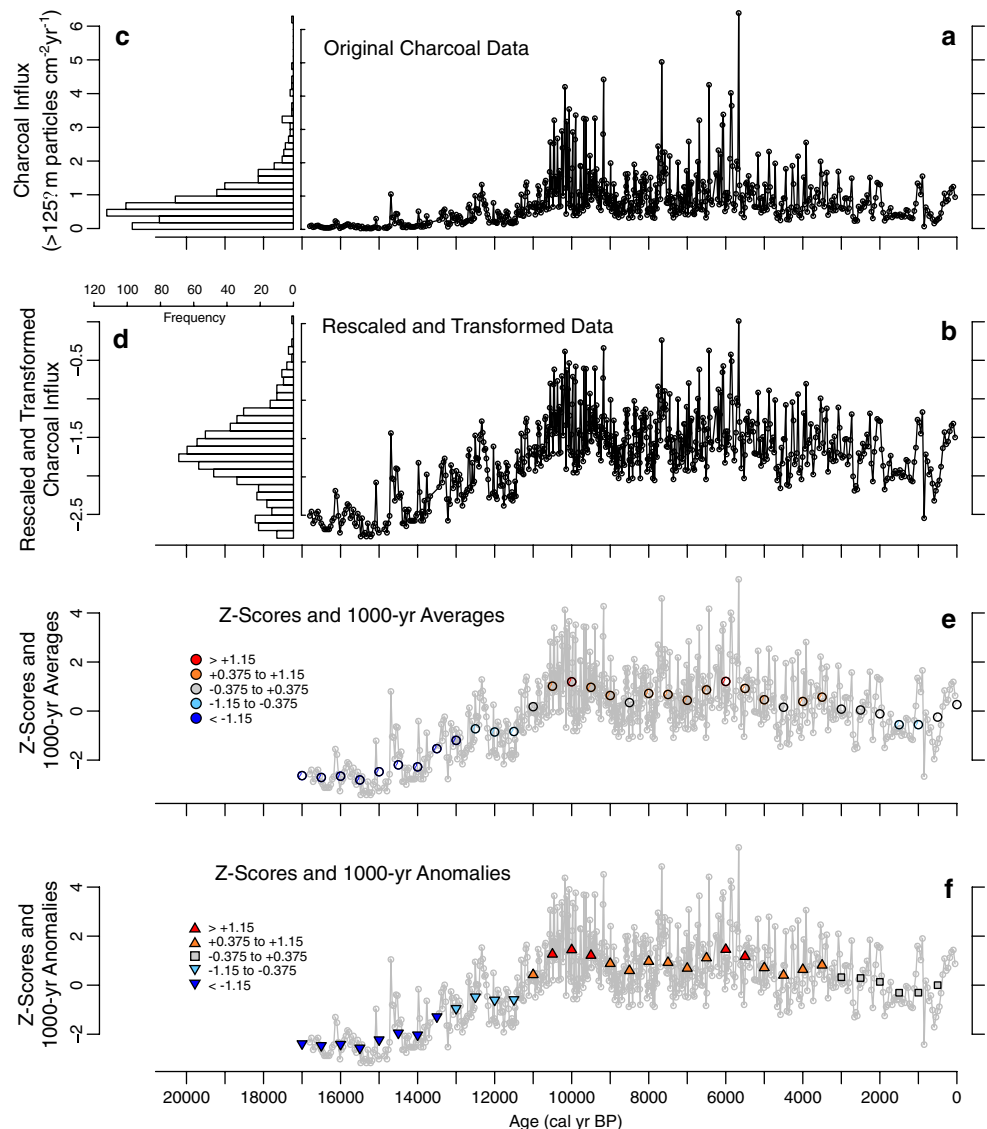
(Fig. 1). There are relatively few charcoal records available for the LGM and the early phase of the deglaciation. Although there are >200 sites with records for the past 10,000 cal yr BP (i.e., most of the Holocene, that is about the last 11,500 cal yr), some regions (e.g., the boreal forest zone of 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. 15,000–12,000 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. We discuss the spatial patterns of charcoal anomalies at each time step relative to present (here defined as the interval 1,000–100 cal yr BP; Fig. 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 (Fig. 4b). Within these broad regions, however, there is spatial heterogeneity in the change in fire regimes. For example, sites in southern

Fig. 4 a Global map of mean *Z* scores of charcoal values for 1,000 to 100 cal yr BP illustrates sites that have higher (*red*) or lower (*blue*) average charcoal values during the last millennia when compared to the last four millennia. The global anomaly maps (**b–h**) at 3,000-year time slices, beginning at 21,000 through 3,000 cal yr BP, permit comparisons of change in charcoal accumulation relative to present (1,000 to 100 cal yr BP). The continental palaeogeographies are mapped for 21,000 and 6,000 cal yr BP. The anomaly maps reveal both the considerable spatial heterogeneity as well as regional coherencies of global charcoal
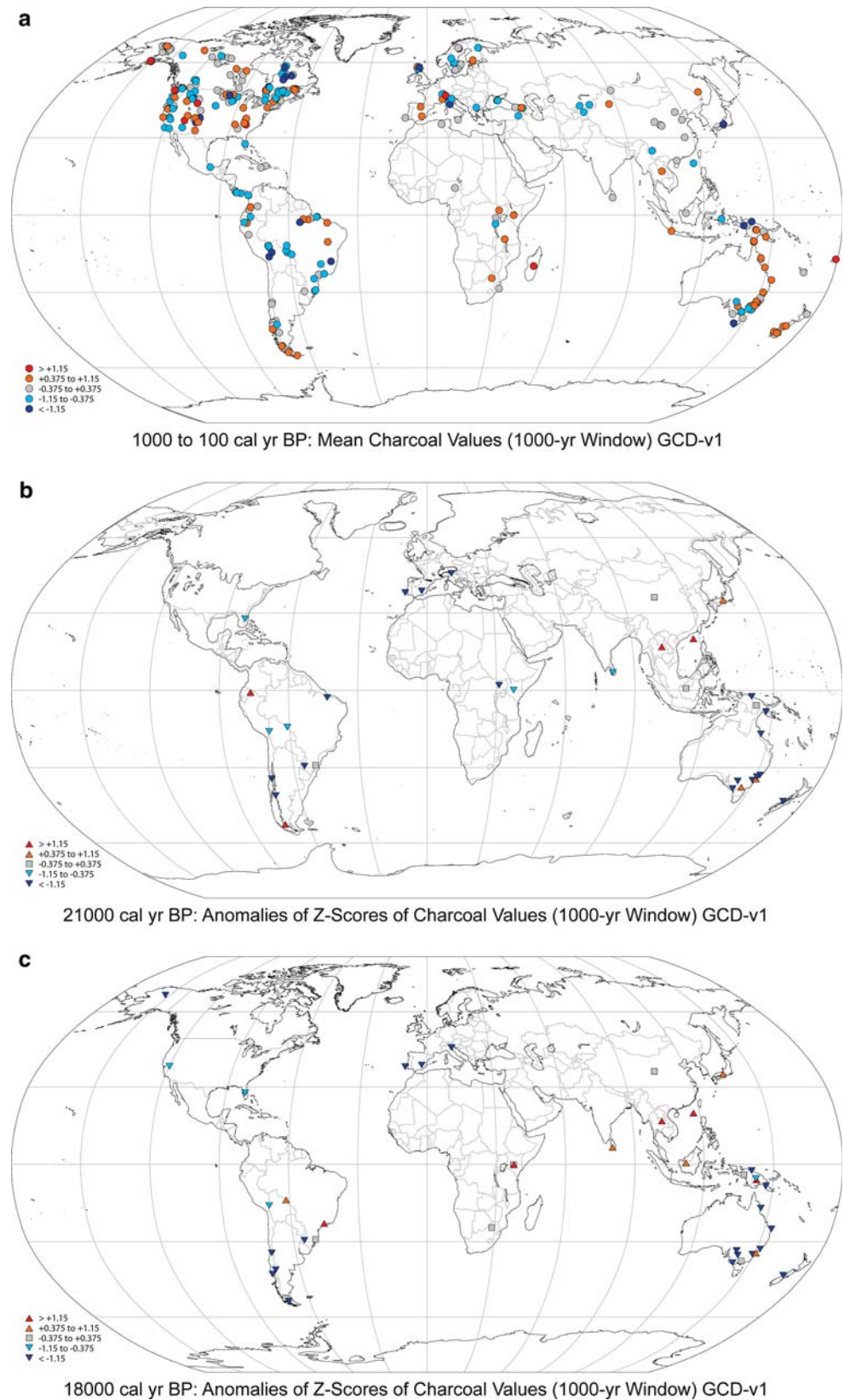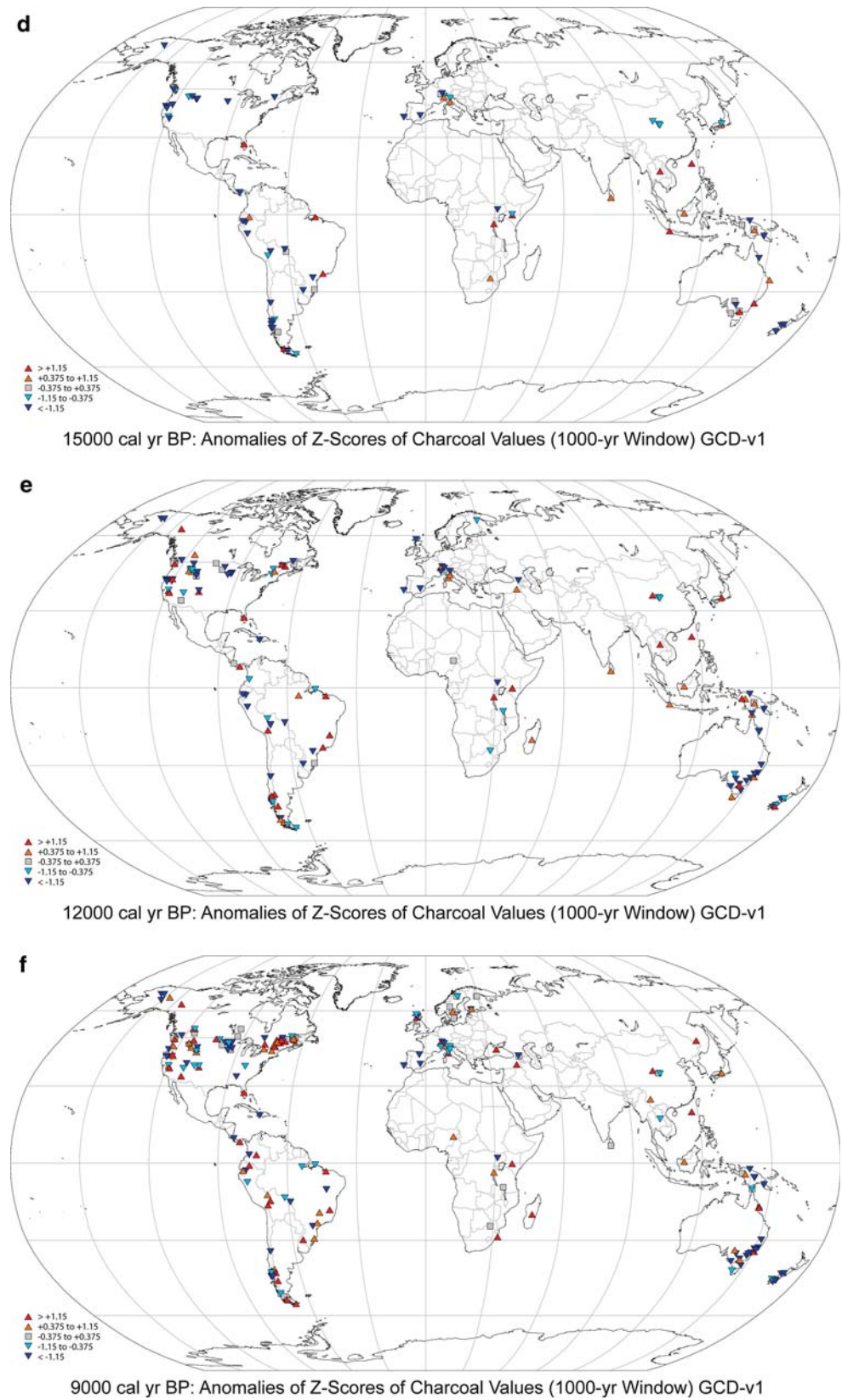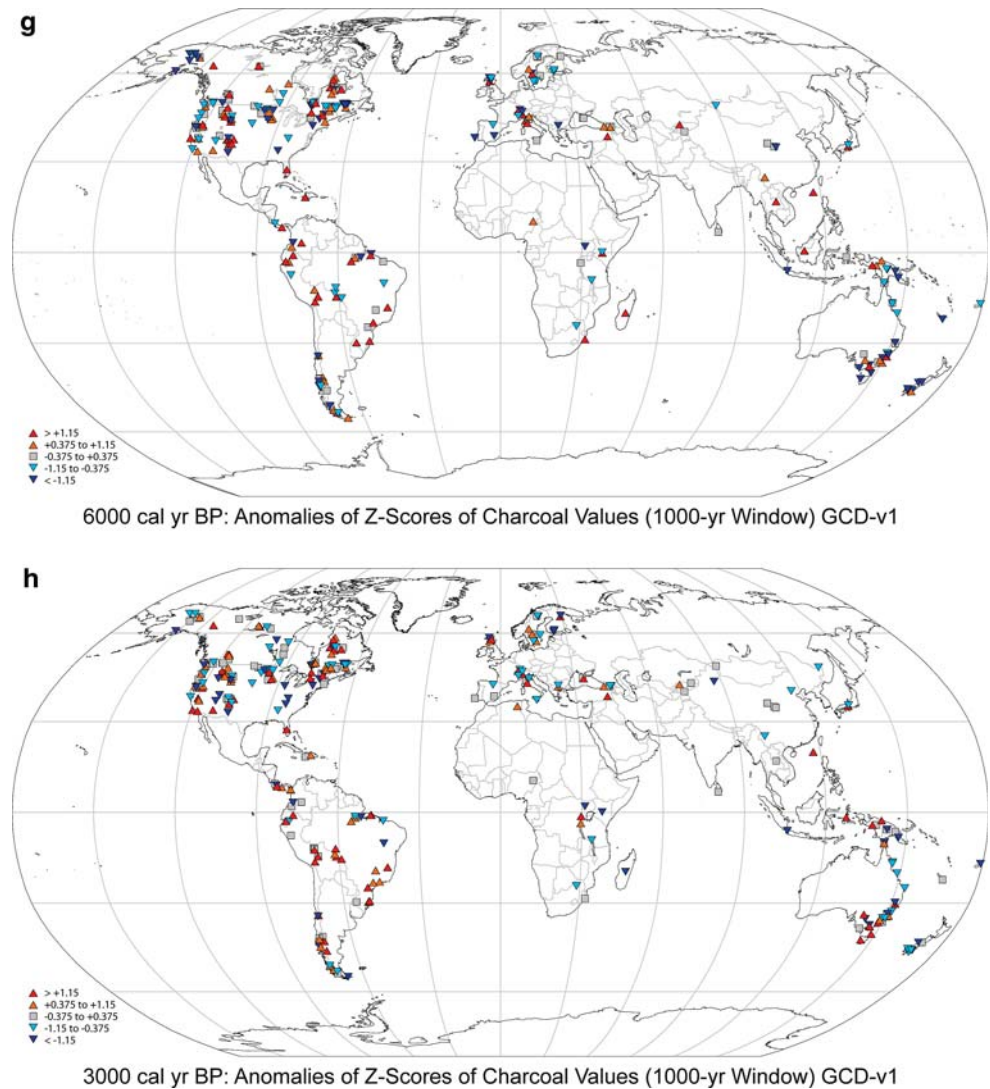

1000 to 100 cal yr BP: Mean Charcoal Values (1000-yr Window) GCD-v1

21000 cal yr BP: Anomalies of Z-Scores of Charcoal Values (1000-yr Window) GCD-v1

18000 cal yr BP: Anomalies of Z-Scores of Charcoal Values (1000-yr Window) GCD-v1

Fig. 4 continued

d

▲ > +1.15
▲ +0.375 to +1.15
■ -0.375 to +0.375
▼ -1.15 to -0.375
▼ < -1.15

15000 cal yr BP: Anomalies of Z-Scores of Charcoal Values (1000-yr Window) GCD-v1

e

▲ > +1.15
▲ +0.375 to +1.15
■ -0.375 to +0.375
▼ -1.15 to -0.375
▼ < -1.15

12000 cal yr BP: Anomalies of Z-Scores of Charcoal Values (1000-yr Window) GCD-v1

f

▲ > +1.15
▲ +0.375 to +1.15
■ -0.375 to +0.375
▼ -1.15 to -0.375
▼ < -1.15

9000 cal yr BP: Anomalies of Z-Scores of Charcoal Values (1000-yr Window) GCD-v1

Fig. 4 continued

6000 cal yr BP: Anomalies of Z-Scores of Charcoal Values (1000-yr Window) GCD-v1

3000 cal yr BP: Anomalies of Z-Scores of Charcoal Values (1000-yr Window) GCD-v1

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.

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 (Fig. 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

(Fig. 4d). 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

Fig. 5 Global and regional summaries of average anomaly Z scores of charcoal values. **a** 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*. **b** The global average anomalies for Z scores of charcoal values for all records are shown at *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
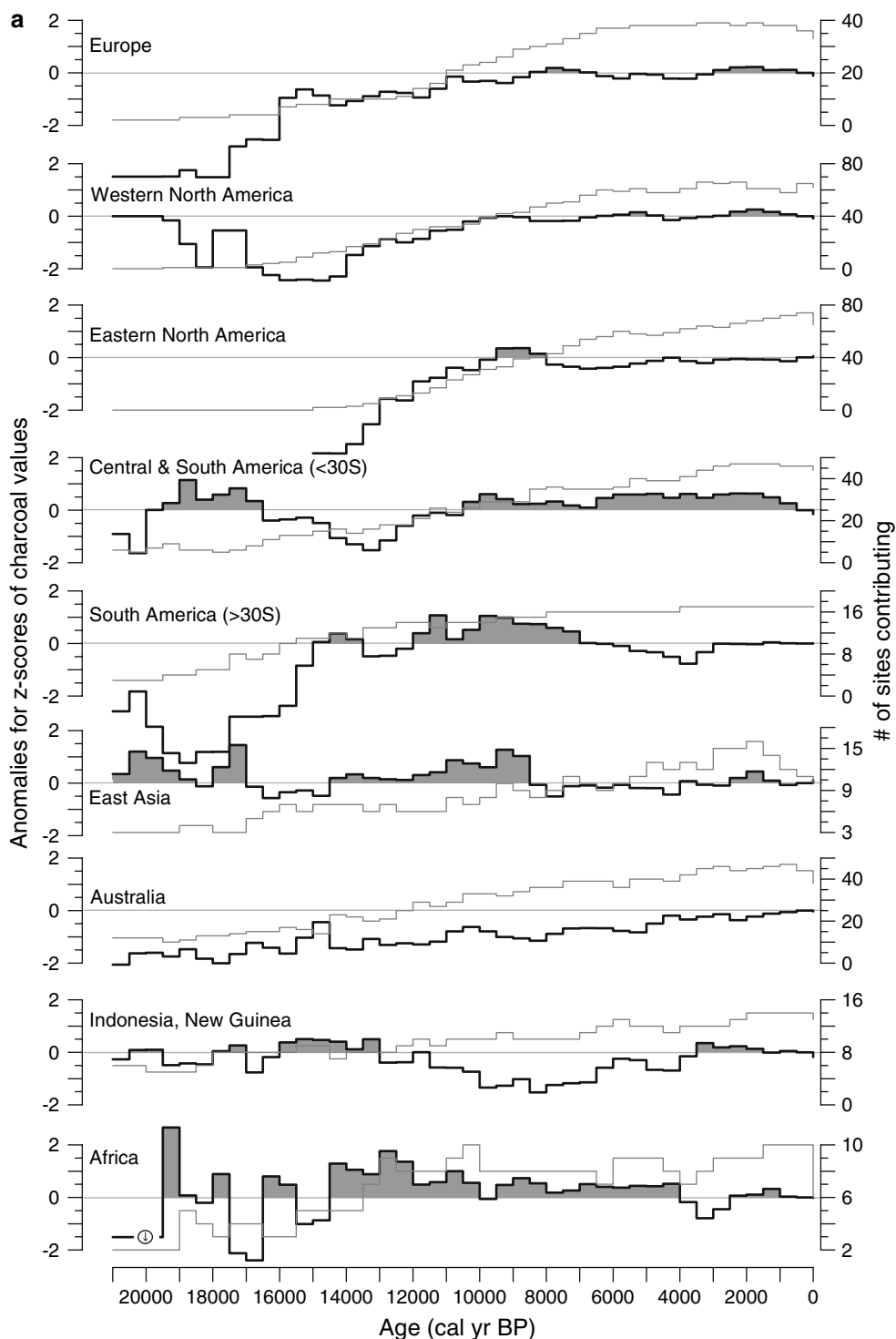
greater than present). In general, the mapped patterns show more spatial heterogeneity than during earlier periods (Fig. 4e), but the globally averaged Z score anomaly remains lower than present (−0.52). In the low latitudes of South America, areas of eastern Brazil continue to indicate greater-than-present 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, b; 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

Fig. 5 continued
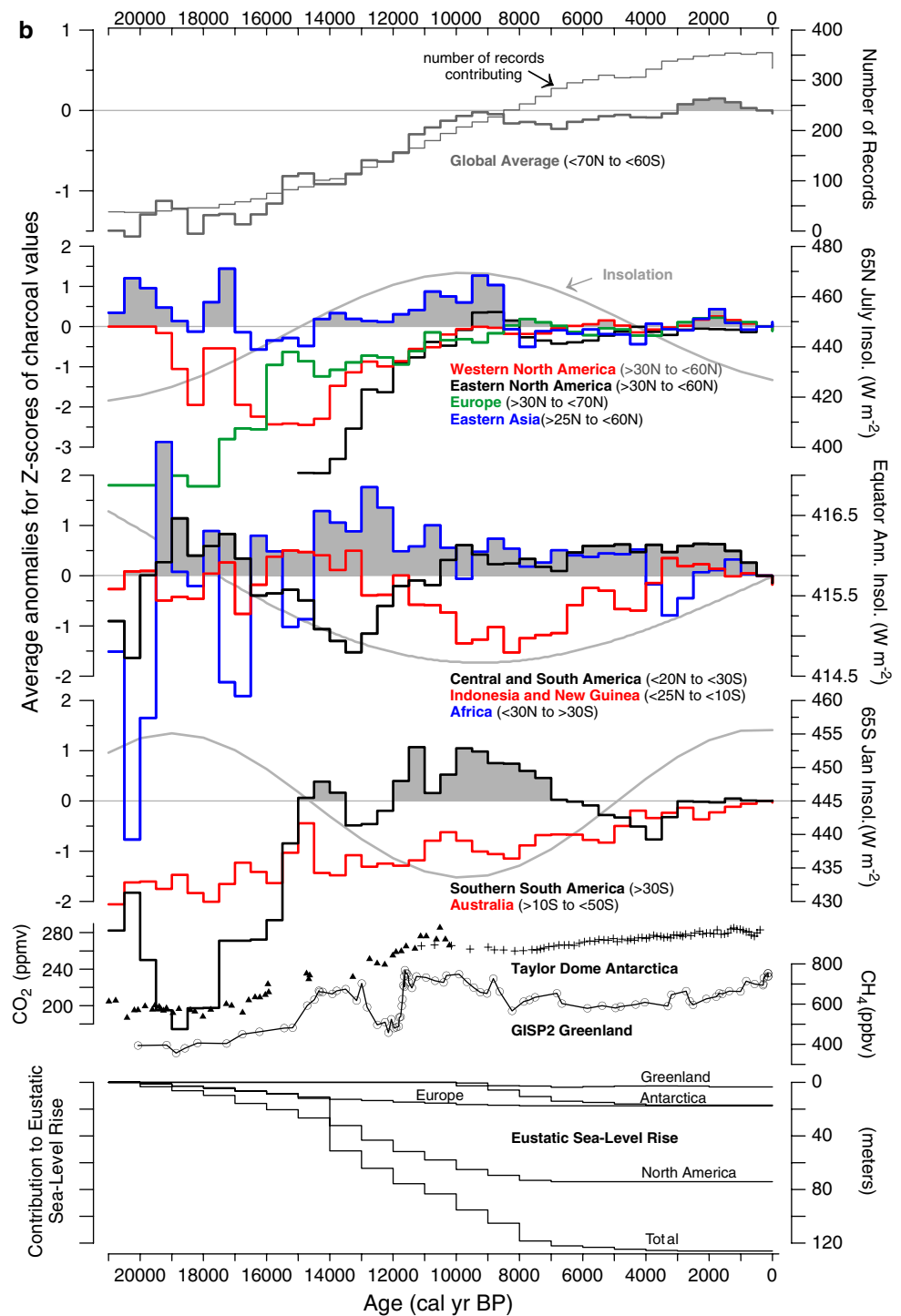
New Guinea, northeastern Australia and from western Colombia, Ecuador and Peru, show less-than-present fire. In southeastern Australia, a shift to less-than-present fire occurred at most mainland sites, although Tasmanian records indicate greater-than-present fire and sites in New Zealand continue to 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 9,000 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 9,000 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 (Fig. 5a) 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 (Fig. 4f). 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 9,000 cal yr BP. Sites in the northern mid-latitudes of Europe show heterogeneity in fire regimes at 9,000 cal yr BP.

At 6,000 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 (Fig. 4g). 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 (Fig. 5b). 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 6,000 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 9,000 cal yr BP.

By 3,000 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 3,000 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 3,000 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 3,000 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 (Fig. 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 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. 120 m 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 6,000 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-year 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-year intervals (not shown) that supplement those in Fig. 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 (Fig. 5a). Comparison of these grouped time series with time series of large-scale climate controls (Fig. 5b) suggest that the global charcoal record since 21,000 cal yr ago can be divided into four relatively distinct intervals: (1) a glacial interval (typified by patterns at 21,000–16,000 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 (15,000–12,000 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 11,000–7,000 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 yr BP)

Although there are relatively few charcoal records for the LGM and subsequent millennia, they show a consistent pattern of low fire (Fig. 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 (Fig. 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 W m^{-2}) during the LGM (Fig. 5b), 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 (Fig. 5b) 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. 15,000–12,000 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) (Fig. 5b). 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. 11,000–7,000 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 (Fig. 5a, b). 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 8,000 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 nonlinear 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) (Fig. 5b). Increased CO_2 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. 2004). Methane records from the GISP2 ice core reveal increased variability around 12,000 cal yr BP (Fig. 5b), 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 9,000 cal yr BP, with significant decreases in eastern Asia, Indonesia, eastern North America, and Africa by ∼ 8,000 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 6,000 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 (Fig. 5b), 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 6,000 cal yr BP. For example, in Australia, a period of maximum precipitation between 7,000 and 5,000 cal yr BP (Harrison and Dodson 1993) may have been responsible for reduced fire at 6,000 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 6,000 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 7,000–5,000 cal yr BP that resulted from combined effects of intensified land-use activities and centennial-scale shifts to warmer and drier climatic conditions.

By 3,000 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 3,000 cal yr BP compared to 6,000 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 3,000 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 3,000 cal yr BP.

5 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 (1) the monotonic increase in biomass burning between the LGM and present, and (2) 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 (Fig. 4), characterizing the glacial interval from 21,000 to 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 (Figs. 4a–h, 5a–b).

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 behavior. We anticipate that the maps (Fig. 4) and regional time series (Fig. 5) 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 6,000 cal yr BP, but especially from 3,000 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 behavior of vegetation and fire (Marlon et al., in preparation). 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.

6 List of contributors

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