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► To cite this version:

Anne-Laure Daniau, S. P. Harrison, P. J. Bartlein. Fire regimes during the Last Glacial. *Quaternary Science Reviews*, 2010, 29 (21-22), pp.2918-2930. 10.1016/j.quascirev.2009.11.008 . hal-03210048

HAL Id: hal-03210048

<https://hal.science/hal-03210048>

Submitted on 28 Apr 2021

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Fire regimes during the last glacial

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Ms. For Quaternary Science Reviews, D-O Special Issue

Version: Oct 04th 2009

Abstract

Sedimentary charcoal records document changes in fire regime. We have identified 67 sites (30 sites with better-than millennial-resolution) which have records for some part of the last glacial to analyse changes in global fire regimes. Fire was consistently lower during the glacial than during the Eemian and Holocene. Within the glacial, Marine Isotope Stage (MIS) 3 is characterised globally by more fire than MIS 2. The signal for MIS 4 is less clear: there is more fire in the northern hemisphere and less fire in the southern hemisphere than during MIS 2 and 3. The records, most particularly records from the northern extratropics, show millennial-scale variability in fire regimes corresponding to the rapid climate changes associated with Dansgaard-Oeschger (D-O) cycles. Most of the D-O cycles during the last glacial and all of the Heinrich Stadials are apparent in the composite global record of the high-resolution sites: fire increases during D-O warming events and decreases during intervals of rapid cooling. Our analyses show that fire regimes show a lagged response to rapid climate changes of ca 100-200 years in the case of D-O warming events, ca 0-100 years in the case of D-O cooling events and ca 200 years in the case of Heinrich Stadials. The strong climatic variability experienced during the glacial resulted in important changes in fire regimes even though the base level of biomass burning was less than today.

Introduction

Fire is the major cause of vegetation disturbance and helps to maintain vegetation diversity, productivity and nutrient cycling (Bond and Keeley, 2005; Bowman et al., 2009). The occurrence of fire is largely controlled by climatic conditions (e.g. van der Werf et al., 2008; Archibald et al., 2008), and thus provides a mechanism (along with disease mortality) through which vegetation can track climatic changes, but is also influenced by vegetation structure and productivity (Marlon et al., 2006; Lavorel et al., 2007; Archibald et al., 2008). Thus, climate changes will influence the fire regime both directly, through influencing lightning ignition, fuel moisture and the prevalence of fire weather, and indirectly through changes in vegetation type and productivity. Changes in fire regime in turn feedback on climate through their effects on the global carbon cycle, atmospheric chemistry, vegetation cover, and other physical and biological properties of the land surface such as albedo, soil water holding capacity and soil nutrient status (DeBano et al., 1998; Bowman et al., 2009). The interactions between climate, vegetation and fire regimes are complex, and can be difficult to disentangle under modern conditions when fire regimes are influenced by human activities. Charcoal produced during biomass burning episodes is relatively inert: charcoal records preserved in sedimentary sequences can therefore be used to reconstruct long-term changes in fire regime. Examining changing fire regimes on geological timescales provides an opportunity to disentangle the effects of climate and vegetation changes during intervals when human influence was negligible or non-existent.

The fire regime of a given environment is generally described in terms of the

characteristic intensity, severity and frequency of fire and the magnitude of the resulting loss of biomass (Bond and Keeley, 2005; Hu et al., 2006). Although fire frequency can be reconstructed using high-resolution records and special techniques (see Whitlock and Bartlein, 2004), most palaeorecords only provide an indication of relative changes in biomass burning. Such changes are unlikely to have occurred without changes in other aspects of the fire regime, and here we assume that changes in biomass burning reflect changes in fire regime. Sedimentary charcoal records typically yield reconstructions of biomass burning on decadal- through centennial-to-millennial timescales (Haberle and Ledru, 2001; Carcaillet et al., 2002). The Global Palaeofire Working Group (http://www.bridge.bris.ac.uk/projects/QUEST_IGBP_Global_Palaeofire_WG) has created a database over 700 individual charcoal records worldwide (GCD_V2: Daniau and Tinner et al., in prep). Analyses of earlier versions of these data have focused on the past 21,000 years (see e.g. Power et al., 2008; Marlon et al., 2008; Marlon et al., 2009). Here we analyse longer records from the GPWG database to examine broad-scale regional changes in fire regime during the last glacial (73.5-14.7 ka), in particular highlighting those high-resolution records that document the biomass burning response to rapid climate changes.

Source and treatment of charcoal records

The GPWG global charcoal database contains sedimentary charcoal records from both marine and terrestrial cores (Power et al., 2008). The database includes descriptive data (metadata) about both the sites and the charcoal samples, and detailed information on site chronology. Version 2 of the database (GCD-V2: Daniau and

Tinner et al., in prep) contains charcoal records from nearly 700 sites; we have extracted the 67 records (11 marine and 56 terrestrial sites) that extend at least into Marine Isotope Stage 3 (MIS 3, > 27500 cal yr B.P.) for our analysis (Table 1, Fig. 1). Among these sites, 36 are of low resolution (less than one sample per ka for MIS 3 or MIS 4) and 31 of high resolution (more than 1 sample per ka). There are two cores from one of these high-resolution sites (Xere Wapo, Sites 722, 808) and we retained the single core (Site 722) that covered the longest interval for our analyses.

Charcoal records are obtained using several different methods (Power et al., 2008) and as a result are recorded as multiple different measurement types in the GCD. For our analyses, we use charcoal abundance or charcoal concentration. It would be possible to use the original charcoal measurements and the age models in the database to calculate charcoal influx through time. However, estimates of influx are strongly influenced by the quality of the age model and the number of age control or tie-points on some of the longer records is limited. For most of the individual sites used in our analyses, visual inspection of the record of last climatic cycle shows little difference between the records of abundance/concentration and those of influx in terms of temporal evolution; there are large differences however for the loess and marine records (Lijiayuan section: Site 109; Lingtai section: Site 110; Weinan section1: Site 111; MD95-2042: Site 542; MD04-2845: Site 849; ODP1078C: Site 1104; MD97-2140: Site 1105, see Tab. 1), most probably associated with rapid deposition events which cannot be resolved using the dating available. At such sites, we believe that the abundance or concentration data provide the more reliable estimate of changing fire regimes. We have used abundance or concentration data for the other sites for consistency.

Charcoal values vary by orders of magnitude between and within sites; the data have to be standardized to facilitate comparisons between sites and through time. We have adopted a procedure (see Power et al., in press for a fuller description) that involves homogenization of the variance using a Box-Cox transformation, and then rescaling the transformed data as z-scores using a base period of 0.2– 132 ka so that all sites have a common mean and variance.

To place the changes in biomass burning during the glacial in context, we compare the long term signal to records for the preceding (Eemian) and subsequent Holocene interglacials. These comparisons were made using box plots of the transformed charcoal data from the 30 high-resolution sites (Table 1). The age limits for MIS 2 (27.8-14.7 ka), MIS 3 (59.4-27.8 ka) and MIS 4 (73.5-59.4 ka) are defined in Sánchez-Goñi and Harrison (this volume); the limits for the Eemian were defined as 128-115 ka (following Kukla et al., 2002) and those for the Holocene as 11.7 ka to present (following Wolff et al., this volume).

Composite curves were constructed by fitting a lowess curve to the pooled transformed and rescaled data (e.g. Marlon et al., 2008), using a fixed window width, target points for which fitted values were obtained at 20-year intervals, and the tricube weight function with one “robustness iteration.” In the earlier part of the time interval examined here the data become increasingly sparse, and so the window width was increased so that at least 30 data points contributed to each fitted value. The charcoal data were summarized using window half-widths of 400 years for comparison with the NGRIP oxygen-isotope record. The data were detrended by subtracting a lowess

curve calculated using a 2000-year window to represent the longer-term variations in the slowly varying level of the charcoal. These particular parameter values were selected to avoid both oversmoothing, as would result from selecting a large window width, and undersmoothing, leading to a composite curve that was susceptible to influence of individual data points. Other combinations of parameter values produce essentially the same results and conclusions.

To characterize the typical variations in charcoal concentration around D-O warming events, D-O cooling events and Heinrich Stadials (HS), a superposed epoch analysis (SEA: Muscheler and Beer, 2006) was performed on the detrended charcoal data (Singh and Badruddin, 2006), aligning the charcoal deviations on the ages of the D-O warming events, and the start of HS as described in Sánchez-Goñi and Harrison (this volume) and Wolff et al. (this volume). The start of 19 most rapid cooling events was determined by inspection using the ages of the unsmoothed oxygen isotope data (see Table 2).

The data were summarized using 80-year-wide “bins.” This results in less noisy SEA plots, but has side effect of making the peaks more symmetrical about the event (defined as occurring at 0 years). The NGRIP oxygen-isotope data (Wolff et al., this volume) were summarized using 20-year-wide bins (the resolution of the data), and also using 80-year-wide bins, for comparison with the SEA of the charcoal data. Confidence intervals were calculated using a Monte-Carlo simulation (2000 replications) under the null hypothesis of random occurrence of events (Prager and Hoenig, 1992). The SEA composite or summary curves, formed in this application by the median values of the aligned segments of data that fall in a particular bin, show the aggregate or typical response of the variable being examined around the time of

the key events. If there is a consistent response in a time series to some externally determined event, then this is shown in an SEA by the composite curve falling outside of the confidence band. This band shows the variability expected in the SEA composite curves when there is no consistent response of the time series to the event occurrence (as simulated by repeated generation of the same number of random events as in the record of interest).

Broadscale changes in fire regimes between interglacial and glacial climates

The spatial distribution of the charcoal sites is uneven (Fig. 1). There is only one high-resolution site from North America, four sites from southern Europe and three sites from the Chinese Loess Plateau. There are four from tropical South America, two sites from tropical Africa and seven sites from SE Asia and tropical Australasia. There is only one site from extratropical South America but seven sites from Australasia. Nevertheless, the coverage is sufficient to provide a preliminary picture of changes in fire regime during the last glacial.

Globally (Fig. 2a), as indicated by median values of abundance/concentration, biomass burning was higher during the Holocene and the Eemian than the last glacial. This pattern is characteristic of both the northern (Fig. 2b) and southern (Fig. 2c) hemispheres. Within the glacial, biomass burning in the northern hemisphere was lowest in MIS 2, and lower in MIS 3 than the relatively short interval of MIS 4 (Fig. 2b). In the southern hemisphere, biomass burning was highest in MIS 3, and MIS 4 has less biomass burning than MIS 2 (Fig. 2c). The box plots show that inter-site

differences are less in the northern than the southern hemisphere, and hence the global differences between MIS 2, 3 and 4 reflect the northern hemisphere pattern.

These analyses show that biomass burning is generally high during interglacials and low during glacials, a result that is consistent with analyses of the past 21,000 years (Power et al., 2008) which show low biomass burning during the glacial and deglaciation and higher biomass burning during the Holocene. The higher global temperatures associated with MIS 3 (relative to those of MIS 4 and 2) are sufficient to produce an increase in biomass burning relative to MIS 2. There is less differentiation between MIS 3 and MIS 2 in the southern hemisphere, which is consistent with the fact that the maximum expression of cold, dry conditions occurs in the later part of MIS 3 and into MIS 2 (Allan and Lindsay, 1998; Vandergoes et al., 2005). Low levels of biomass burning in the southern hemisphere during MIS 4 are consistent with cold conditions. The high biomass burning in the northern hemisphere during MIS 4 is largely a reflection of the records from the Loess Plateau (Wang et al., 2005); European sites show fire regimes comparable to MIS 3.

There is considerable variation in the level of biomass burning around these median values. The amplitude of this variability, as shown by the 25 and 75 percentile limits on the box plots (Fig. 2), is a reflection of the extreme differences in fire regime within a given period. The variability in biomass burning at a global scale (Fig. 2a) is greater during the glacial than during the Eemian and Holocene. Within the glacial, MIS 3 shows the greatest absolute variability in fire regime. These patterns in the amplitude of variability are consistent with the marked millennial-scale climate variability during MIS 3 (Bond et al., 1993; North Greenland Ice Core Project

Members, 2004) and the fact that climate variability within interglacials is more muted than during glacials (Overpeck and Webb, 2000).

Millennial-scale variability in fire regimes during the glacial

There are considerable differences in the patterns shown by individual records during the glacial (Fig. 3), and few sites show all of the features apparent in the global and hemispheric averages. Nevertheless, most of the records show millennial-scale variability at least for part of the glacial. The exceptions are Lake Pata (Site 432: Bush and Oliveira, 2006); Siberia-93-1 (Site 59: Mourguiart and Ledru, 2003), Xere Wapo C (Site 722: Stevenson and Hope, 2005) and Galway Tarn (Site 437: Vandergoes, unpub.) which show no fire during most of the length of the record – all these sites are sampled at high resolution and have sufficient dating control to document millennial-scale changes if they existed. Most of the records show that millennial-scale variability was more pronounced during MIS 3 than during MIS 2 and 4. The variability during MIS 3 is, in general, less apparent in the tropics than the extratropics (Fig. 3).

The millennial-scale variability expressed in certain charcoal records has been interpreted as a reflection of D-O climate variability through changes in fuel availability (see e.g. MD95-2042: Site 542, Danialu et al., 2007; MD04-2845: Site 849, Danialu et al., 2009). In these two sites, each of the Greenland Interstadials (GI) is marked by an increase in biomass burning contemporaneous with the expansion of forest; the reduction in biomass burning during Greenland Stadials (GS) is

contemporaneous with the development of more open vegetation. Other sites show variability consistent with registration of several different D-O cycles (e.g. Valle di Castiglione: Site 846, Magri, 2008; Allom Lake: site 597, Donders et al., 2005). In these sites, the GI are also characterised by increased biomass burning and the GS by reduced biomass burning. However, many of the records display only a few peaks that appear to correlate to specific warming events (e.g. Lagaccione: site 845, Magri, 2008; Lingtai section: site 110, Wang et al., 2005). In some cases, the peaks occur earlier or later than expected (e.g. Lake Wangoom 20m Core: site 237, Edney et al., 1990; Caledonia Fen: site 859, Kershaw et al., 2007). The lack of coherence, as indeed the coherence of individual peaks, reflects lack of dating control particularly during the earlier parts of the records. Given this, it is not possible to be certain that the apparent reversal of the relationship between warmer phases of the D-O cycle and increased biomass burning shown by some records (see e.g. Native Companion Lagoon: site 819, Chalson, 1991; Lijiayuan section: site 109, Wang et al., 2005) is a reflection of geographic differences in registration or not.

We have constructed a composite global record of fire regime, based on all sites, through the last glacial (Fig. 4). This record shows the generalised pattern of the response of biomass burning to rapid climate changes, and presumably the climate-driven changes in vegetation (see Fletcher et al., this volume; Takahara et al., this volume; Jiménez-Moreno et al., this volume; Heßler et al., this volume; Kershaw et al., this volume), associated with D-O cycles through the glacial. All of the GI except GI-11, GI-12, and GI-17, are marked by prominent increases in the composite charcoal concentration curves at or shortly after the warming event. There are other peaks in the composite curve that occur at times other than the D-O warming events:

some of these are associated with cool intervals (e.g. before and after GI-18), but some of these correspond to unnamed warming events, such as those at 30.7 ka and 69.4 ka. There are prominent troughs in the composite curve corresponding to each of HS (Fig. 4) and at the beginning of the Younger Dryas chronozone.

The general response of biomass burning as represented by the composite global record of fire regime can be seen in the superposed epoch plots for the charcoal and NGRIP oxygen isotope data (Fig. 5). The specific shape of the abrupt warming at the beginning of the GI, most rapid cooling that defines the GS and the more gradual cooling at the onsets of the HS are shown by the thin stepped lines in Figs. 5A, 5C and 5E, respectively. The majority of the warming takes place within a single 20-year sampling interval, the cooling within 200 years and within 500 years in case of the HS. The impact of smoothing (as has been done to construct the charcoal composite curve) is shown by the blue lines in Figs. 5A, 5C and 5E. The smoothing broadens the peaks and troughs in the SEA curves, such that the abrupt warming or cooling appears to begin a few hundred years earlier in the smoothed curves than in the unsmoothed curves. Nevertheless, the SEA curves for the smoothed oxygen-isotope data reflect a similar overall structure to the composite events as do the SEA curves for the unsmoothed data.

These plots show a strong non-linear correlation between biomass burning and temperature, characterised by a general increase in biomass burning during warming (Fig. 5A and B) and a decrease in biomass burning during cooling (Fig. 5C and D). Although biomass burning decreases markedly during the HS, the start of this decrease occurs well before the start of the cooling registered in the NGRIP oxygen isotope curve (Fig. 5E and F). However, this lead is not apparent during the D-O

cooling events which suggests that it is simply an artefact of defining the start of HS using ages defined by Sánchez-Gómez and Harrison (this volume) which are slightly different from the start of decrease of temperature of NGRIP. Note that the charcoal SEA summary curves do not exactly match the corresponding oxygen-isotope SEA curves, as they would be expected to do if biomass burning as represented by charcoal were a proxy for the temperature variations represented by the Greenland oxygen-isotope data. Instead, the comparison of the SEA curves show that biomass burning increases rapidly during D-O warming events, and reaches a peak at nearly the same time as temperature. In contrast, at the onset of the GS and HS, biomass burning decreases significantly and then returns to “background” levels while temperatures remain low (Figs. 5C and D; 5E and F). There is a significant peak in biomass burning corresponding to the D-O warming events, lagging the actual timing of the event by ca 100-200 years, roughly the same span that the smoothed oxygen-isotope data lags the unsmoothed data (Fig. 5B). The SEA curve lies outside of the confidence band at the onset of the GI and so there is evidence for a significant increase in biomass burning accompanying abrupt warming throughout MIS 2-4. During cooling events, biomass burning shows a significant trough ca 0-100 years after the onset of the start of the most rapid cooling of the GS (Fig. 5D), peaking ca 200 years after the onset of the cooling phase in case of the HS (Fig. 5F).

Discussion

This is a first attempt to document changes in global fire regimes during the glacial. Despite inter-site variability in the registration of changes in biomass burning, the

composite record shows millennial-scale variability in fire regimes consistent with a response to D-O climate variability and concomitant changes in vegetation. Some records show a clear record of this variability, while others appear to register only some of the D-O cycles.

There are comparatively few charcoal records which cover the last glacial, and fewer which were sampled at sufficiently high resolution to allow diagnosis of millennial-scale variability. The dating control on these records, particularly for MIS 4 and the first part of MIS 3 is, in general, poor – for the majority of records (60; see Table 1), the age model is based on extrapolation from available radiocarbon dates. In a small number of cases (3 records), other radiometric methods or luminescence dates were used to extend the age model. The age models for charcoal records from marine cores are principally based on radiometric methods and marine isotope stratigraphy. Orbital tuning of grain-size records was used to date the three cores from China. We have made no attempt to tune radiometrically-dated records to match the Greenland record nor have we converted existing orbitally-tuned age models to the GICC05 timescale. It is therefore encouraging that it is possible to extract a composite record and that this record shows millennial-scale variability throughout the glacial.

The most complete records of D-O variability in fire regimes are from analyses of charcoal in marine sediment cores. These records likely sample a larger source area than most terrestrial records (Lynch et al., 2007). Nevertheless, where it is possible to make comparisons with terrestrial charcoal records (see e.g. Moss and Kershaw, 2000) they appear to show similar patterns of change within the uncertainties of the chronology. Our analyses show that marine and terrestrial records show similar and

coherent millennial-scale variability. Marine records have several distinct advantages for the analysis of long-term and millennial-scale variability in fire regimes: they are quasi-continuous and thus more likely to yield extremely long records, and age-modelling is made easier because of the existence of isotopic measurements on the same samples. Unfortunately, many fewer marine cores have been analysed for charcoal than for pollen (see e.g. Table 1).

The available long records show that the incidence of biomass burning was lower during the glacial than during the Eemian and the Holocene, supporting the idea of a strong relationship between temperature and fire regimes. This is also supported by the fact that globally biomass burning was higher during MIS 3 than MIS 2. These relationships are consistent with the finding that biomass burning was low during the last glacial maximum and the deglaciation, and higher (though more spatially variable) during the Holocene (Power et al., 2008; Danialu, Tinner et al., in prep.). The dominance of the temperature signal on biomass burning at glacial-interglacial timescales is unsurprising, given that colder climates are in general drier because of the Clausius-Clapeyron relationship (Clausius, 1865). Colder, drier conditions would reduce vegetation productivity and thus the availability of fuel for burning. As climate warms, and hence the global hydrological cycle becomes more vigorous, vegetation productivity would increase leading to an increase in fuel and fire.

A strong relationship between biomass burning and temperature, as recorded in the Greenland ice core, is apparent within the glacial. Global biomass burning is, in general, higher during GI and lower during GS. However, this signal is most marked in Europe. The composite biomass burning record presented here does not show the

antiphasing apparent between the Greenland and Antarctic ice core records (EPICA Community Members, 2006). This does not appear to be due to a paucity of southern hemisphere charcoal records, but could reflect the distribution of these records given that some southern hemisphere climate regimes are strongly linked to northern hemisphere temperature variations through the thermohaline circulation (Clark et al., 2002). Conversely, it may reflect problems with the lack of tie points to constrain the age models for sites in the tropics and southern extratropics.

Intervals of coolings associated with the end of D-O events and with HS showed an initial fire decline lasting 100-200 years followed then by slow increase of fire to levels similar to pre-cooling fire regimes. The long records also confirm that rapid warming during the glacial is accompanied by a marked increase in biomass burning. Observed changes in fire regime occur rapidly and lags D-O warming events by ca 100-200 years. Sanchez-Goni et al. (2009) have shown a similar time lag between D-O warming and vegetation changes in southwestern Europe. The association between rapid warming and increases in biomass burning seen during the glacial has also been observed during the last deglaciation. Marlon et al. (2009) studied changes in fire regime across North America during the last glacial-interglacial transition and demonstrated that the rapid warming at the termination of the Younger Dryas chronozone was marked by increases in both the incidence of fires (registered by charcoal peaks), and the overall level of biomass burning. They did not attempt to quantify whether there was a lag between the warming and the change in the fire regime, but none is apparent in their composite curves. The objectively identified times of change in the trend of charcoal influx over the interval were indistinguishable from the beginning and end of the Younger-Dryas chronozone, however, which

suggests little lag. According to Marlon et al. (2009), the rapid cooling at the onset of the Younger Dryas was associated with a moderate increase in biomass burning, but that increase could also be associated with the warming at 13.2 ka following the Inter-Allerød Cool Period. One might expect to see an increase in biomass burning associated with the apparently very rapid cooling that marks the transition from GI to GS in some of the shorter D-O cycles (e.g. D-O 6). We have not explored this possibility, largely because the existing records do not have sufficient sampling resolution during these short cooling events (see Fig. 3) but this would be an issue worth exploring by re-sampling existing cores. Overall, the specific changes in biomass burning across the Younger Dryas chronozone described by Marlon et al. (2009) appear replicated by the multiple instances of abrupt warming and cooling present in the longer records analyzed here.

Both fire regime and vegetation (see Fletcher et al., this volume; Takahara et al., this volume; Jiménez-Moreno et al., this volume; Heßler et al., this volume; Kershaw et al., this volume) records show a clear response to D-O climate variability. The signal in both is clearest in the northern extratropics, and more muted in the tropics and southern hemisphere (see Heßler et al., this volume; Kershaw et al., this volume). The similarities between the two types of record open up the possibility that changes in fire regime are not caused directly by changes in climate but rather are modulated through climate-induced changes in vegetation. There are several ways in which changes in vegetation can affect the fire regime. Firstly, changes in climate affect vegetation productivity, leading to increases/decreases in the availability of fuel. Increased productivity during intervals of rapid climate warming, for example, could explain the observed increase in biomass burning particularly in regions which were

fuel-limited (see e.g. van der Werf et al., 2008). Climate changes also lead to changes in the relative abundance of different plant functional types (PFTs), and thus produce differential mortality amongst these PFTs. This could also affect the amount of fuel available. When the change in climate is sufficiently large, there may be wholesale shifts in biomes. Transitions to biomes which are more open or characterised by fire-promoting species (e.g. savannas or grasslands: Bond et al., 2005) could lead to increases in biomass burning. Biome transitions are characteristic of glacial-interglacial transitions, but can also occur during D-O cycles (see Fletcher et al., this volume).

Increases in biomass burning both on glacial-interglacial timescales and during D-O cycles are linked to either shifts from open to forest vegetation (see e.g. Fletcher et al., this volume) or increases in tree abundance (see e.g. Takahara et al., this volume). Under modern conditions, reductions in forest cover and the creation of more open vegetation would *generally* lead to an increase in fire (van der Werf et al., 2008). This suggests that the changes in fire regime during the glacial are more likely to be related to changes in productivity. Both the globally cold and dry conditions and the low CO₂ concentration during the glacial limit productivity. We suggest that the climate changes that gave rise to vegetation shifts during the D-O warming events and GI would also have increased productivity overall, leading to increased fuel availability and hence more fire. Similarly, decreased fire in response to rapid cooling is a consequence of decreased fuel loads because of decreased vegetation productivity. The reasons for a return of fire to levels similar to pre-cooling fire regimes during D-O cooling events or HS are less clear. Vegetation responses to sufficiently large and rapid cooling may include forest die-back, which would create conditions for

increased fire.

One motivation for focusing on palaeofire regimes is to examine the relationships between climate, vegetation and fire when human influence on fire regimes is assumed to have been negligible. However, there have been speculations that the emergence of Modern Humans was associated with extensive use of fire (McBrearty and Brooks, 2000). Furthermore, it has explicitly been claimed that Aboriginal colonization of Australia around ca 45ka led to a major change in biomass burning (Kershaw, 1986; Kershaw et al., 2006). Modern humans arrived in Europe around ca 40 ka, but detailed analyses of charcoal records from western Europe show no changes in biomass burning over this interval (Daniau et al., in rev.) nor do the other European records included in this synthesis show a change in fire regime at this time. We also find no evidence of a significant shift in biomass burning in Australia around ca 45ka – again a result which is consistent with the increasing number of analyses of individual pollen and charcoal records from Australasia which show no significant change in fire regimes at the time of initial colonization (e.g. Stevenson and Hope, 2005; Dodson et al., 2005).

Future perspectives

This paper is a first attempt to synthesise charcoal records covering the last glacial and to explore changes in global fire regimes during a period characterised by millennial-scale climate variability and rapid warming events. Some interesting features have emerged from these analyses, but it is clear that the extant records provide fewer answers than questions.

There are comparatively few records: our analyses are based on an order-of-magnitude fewer sites than were used for analyses of the past 21 ka (Power et al., 2008; Daniaux and Tinner et al., in prep). A significant increase in the number of records would be required to derive statistically robust results at a regional scale, results that would permit us to examine the possibility of differential responses to out-of-phase climate changes between the northern and southern hemispheres (see Blunier et al., 1998; EPICA Community Members, 2006). The increasing interest in charcoal records from marine cores (see e.g. Beaufort et al., 2003; Thevenon et al., 2004; Daniaux et al., 2007; 2009) is driving the production of more long records. Further expansion of the number of records could be achieved by exploiting the large number of archived marine cores that cover the last glacial. However, this should be accompanied by systematic studies of paired terrestrial and marine records, to determine at what scale the marine cores homogenise geographic differences in fire regimes.

Comparatively few of the charcoal records have been sampled at sub-millennial resolution. This precludes analysis of phenomena which have been studied during the more recent past (see e.g. Marlon et al., 2009), such as the abrupt changes in biomass burning that accompany equally abrupt changes in climate, or the consideration of multiple hypotheses for explaining particular features in charcoal records (e.g. Marlon et al., 2008). It would be interesting, for example, to see if there is a short-term increase in biomass burning at the transitions marking the onsets of GSs in those D-O cycles which are marked by a decrease in temperature in 20-60 years (e.g. D-O 6). This could be done by augmenting the sampling resolution for existing cores, but new

high-resolution studies of specific D-O cycles would certainly be worthwhile.

Chronological control beyond the limits of radiocarbon dating is always problematic (Sirocko et al., 2007). In the case of charcoal records from marine sites, direct correlation of isotopic records with the Greenland isotope record is possible.

Unfortunately, oxygen isotope records from lake sites cannot be used in this way because they are affected by changes in multiple climate factors (Roberts et al., 2008). Tephra can be used to synchronise records from within a specific region (Lowe et al., 2001); however, this still leaves the issue of precise dating and correlation with other regions open. Luminescence techniques provide a mechanism for dating older sediments, but are subject to large error margins and have not proved particularly successful on the lacustrine sediments (see e.g. Thomas et al., 2003) from which most of the terrestrial charcoal records are obtained. However, it is possible that better results could be obtained using the newly-developed single-grain multiple aliquot techniques (Roberts et al., 2003; Roberts, 2007), and these approaches should certainly be tried in order to improve the chronological control on charcoal records. Orbital tuning techniques have been used to date charcoal records from the Chinese Loess Plateau (Wang et al., 2005). However, without independent means of dating, the use of orbital tuning in this way precludes any attempt to investigate leads and lags in the relationship between rapid climate changes and fire (see e.g. Kohfeld and Harrison, 2003). It is clear that many extant charcoal records would benefit from better dating, including the use of multiple dating techniques, although this may still not provide a complete solution to the problem of improving the chronology of long records.

We have speculated that changes in vegetation productivity may have been implicated in bringing about the observed changes in fire regime during the D-O cycles.

Palaeoenvironmental data document what has happened in the past (Kohfeld and Harrison, 2000), and provoke this kind of speculation. Regional syntheses of vegetation data during the last glacial, such as those presented by Fletcher et al.(this volume); Takahara et al.(this volume); Jiménez-Moreno et al.(this volume); Heßler et al.(this volume); Kershaw et al.(this volume), will allow a more rigorous comparison between fire and vegetation records than is currently possible. However, disentangling the direct effects of climate, of changes in productivity, and changes in vegetation type on fire regimes requires modelling of the coupled climate-vegetation-fire system. With the advent of fully dynamic vegetation-fire models (Lenihan and Neilson, 1998; Thonicke et al., 2001; Venevsky et al., 2002; Prentice et al., 2007), exploration of the relative importance of different mechanisms in driving changes in fire regimes through comparing carefully-designed experiments with observations now becomes possible. This approach has been used, for example, to quantify the relative impact of the direct effects of low CO₂ and of climate changes on glacial vegetation (Harrison and Prentice, 2003). Vegetation-fire models have been used to examine the consequences of LGM climate changes on fire regimes (Thonicke et al., 2005). Similar experiments could be run using output from climate model simulations of the D-O cycles (see e.g. Kageyama et al., this volume).

Palaeoenvironmental databases are a useful tool for exploring many facets of past climate changes (see Harrison, 2007). Our analyses of fire regimes during the last glacial were greatly facilitated by the existence of the GPWG charcoal database. Continued development of this database will provide more opportunities for the kind

of exploratory work we have done here. In conjunction with databases documenting changes in other aspects of the palaeoenvironment, most particularly vegetation changes, analyses of the charcoal records will provide insights into the relationships between climate, vegetation and fire, and ultimately a better understanding of the mechanisms of past climate changes.

Acknowledgements

This article is a contribution to the QUAVIDA working group of the ARC-NZ Network for Vegetation Function and to the Global Palaeofire Working Group (GPWG) of the International Geosphere-Biosphere Project Cross-Project Initiative on Fire. The GPWG is supported by the UK Natural Environment Research Council's QUEST (Quantifying Uncertainty in the Earth System) programme. Data analysis was supported by the QUEST-DESIRE project (SPH), and by the US National Science Foundation Paleoclimatology Program (ATM 0714146, PJB). ALD is supported by a QUEST International Research Fellowship. We thank our colleagues in the GPWG who have made these analyses possible through their contributions to the GPWG Global Charcoal Database. We thank Masa Kageyama, Andrew Scott, Pierre Friedlingstein and Maria-Fernanda Sánchez-Goñi for their valuable comments which greatly improved this manuscript.

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Table 1: Charcoal records from the GCD_V2 covering all or part of the last glacial (MIS 4, 3 and 2). The sites are classified according to sampling resolution: high (> 1 sample per ka) or low (< 1 sample per ka).

ID_site (GCD_V2)	Site name	Latitude (decimal degrees)	Longitude (decimal degrees)	Elevation (m)	Site type	Presence of Eemian (E) and/or Holocene (H)	Glacial coverage (MIS)	Dating control during MIS 4-2	Sampling resolution during MIS 4- 2	Reference	Pollen record available
59	Siberia 93-1	-17.8333	-64.7189	2920	terrestrial	H	2, part of 3	10 ¹⁴ C dates	High	Mourguiart and Ledru, 2003	yes
83	Morro de Itapeva	-22.7833	-45.5733	1850	terrestrial	H	2, part of 3	3 ¹⁴ C dates	Low	Behling, 1997	yes
85	Catas Altas	-20.0833	-43.3667	755	terrestrial	----	2,3	6 ¹⁴ C dates	Low	Behling and Lichte, 1997	yes
90	Cambara do Sul	-29.0525	-50.1011	1040	terrestrial	H	2, part of 3	3 ¹⁴ C dates	High	Behling et al., 2004	yes
95	17940 core	20.11667	117.3833	-1727	marine	H	2, part of 3	8 AMS dates	Low	Sun et al., 2000	yes

109	Lijiayuan section	36.11667	104.8583	1850	terrestrial	E, H	2,3,4	Orbital grain size curves tuning	High	Wang et al., 2005	no
110	Lingtai section	35.00917	107.5092	1400	terrestrial	E, H	2,3,4	Orbital grain size curves tuning	High	Wang et al., 2005	no
111	Weinan section1	34.06667	109	650	terrestrial	E, H	2,3,4	Orbital grain size curves tuning	High	Wang et al., 2005	no
120	Crominia	-17.2785	-49.4181	710	terrestrial	H	2, part of 3	3 ¹⁴ C dates	High	Salgado-Labouriau et al., 1997	yes
140	Tagua Tagua	-34.5	-71.1667	200	terrestrial	H	2, part of 3	8 ¹⁴ C dates	High	Heusser, 1990	yes
188	Maxus-Core5	-0.7	-76.45	3	terrestrial	H	2, part of 3	1 AMS, 2 ¹⁴ C dates	Low	Athens and Ward, 1999	no
233	Banda Sea Core SHI-9014	-5.76667	126.9667	-3163	marine	E, H	2, 3, 4	7 AMS, 3 tie-points based on marine isotope stratigraphy	Low	van der Kaars et al., 2000	yes
234	ODP Site 820	-16.6333	146.3	-280	marine	E, H	2,3,4	3 AMS dates, 1 tie-point based on MIS boundary	High	Moss and Kershaw, 2000	yes
236	Lake George	-35.0656	149.4181	673	terrestrial	1 sample in E, H	2,3,4	3 ¹⁴ C dates	High	Singh et al., 1981	yes
237	Lake Wangoom 20m	-38.35	142.6	100	terrestrial	H	2, part of 3	3 ¹⁴ C dates	High	Edney et al., 1990	yes

	Core										
240	Tom Burns Missen Core	-37.3833	145.8167	1075	terrestrial	H	2, part of 3	1 date ¹⁴ C	Low	McKenzie, 1997	yes
243	Lake Baraba Thirlmere Lakes	-34.2342	150.5397	305	terrestrial	H	2,3	1 AMS date	High	Black et al., 2006	yes
244	Redhead Lagoon	-32.9944	151.7208	65	terrestrial	H	2,3,4	4 ¹⁴ C dates	High	Williams, N. (unpublished)	no
248	Cuddie Springs	-30.3795	147.3117	127	terrestrial	E, H	2,3,4	14 AMS dates, 1 OSL	Low	Field et al., 2002	yes
250	Core Fr10/95-GC-17	-22.0458	113.5018	-1093	marine	H	2, part of 3,4	3 AMS dates, 4 tie- points based on marine isotope stratigraphy	High	Van der Kaars and De Deckker, 2002	yes
252	Penrith Lakes	-33.7139	150.6774	18	terrestrial	H	2, part of 3	2 ¹⁴ C dates	Low	Chalson, 1991	yes
323	Sacred Lake	0.045906	37.52885	2350	terrestrial	H	2, part of 3	6 ¹⁴ C dates	Low	Scott, 2000	no
348	ODP-668B	-20.9269	4.77056	-2693	marine	E, H	2,3,4	3 tie-points based on MIS boundary	Low	Bird and Cali, 2002	no
349	Pemerak Swamp	0.7888	112.05	40	terrestrial	H	2, part of 3	2 AMS dates, 1 ¹⁴ C date	Low	Anshari et al., 2001	no
350	Piston Core RC24-07	-11.9167	1.35	-3899	marine	H, E	2,3,4	2 tie-points based on MIS boundary	High	Verardo and Ruddiman, 1996	no

351	Rusaka Swamp	-3.43333	29.61667	2070	terrestrial	H	2, part of 3	1 date ¹⁴ C	Low	Bonnefille et al., 1995	no
355	Kashiru Swamp	-3.46667	29.56667	2240	terrestrial	H	2, part of 3	2 AMS dates	Low	Aucour et al., 1994	yes
360	Lake Rutundu	-0.04217	37.46488	3140	terrestrial	H	2, part of 3	8 AMS dates	Low	Wooler et al., 2003	yes
374	Nong Pa Kho	17.01	102.93	180	terrestrial	H	2, part of 3	2 AMS dates	High	Penny, 2001	yes
432	Lake Pata	0.266667	-66.0667	300	terrestrial	H	2, part of 3	4 ¹⁴ C dates, 7 AMS dates	High	Bush and Oliveira, 2006	yes
437	Galway Tarn	-43.4083	169.8733	130	terrestrial	3 samples in H	2, part of 3	n/a	High	Vandergoes, M. (unpublished)	no
438	Okarito Pakihi	-43.2417	170.2167	70	terrestrial	E, H	2,3,4	2 AMS, 2 TL	Low	Newnham et al., 2007b ; Vandergoes et al., 2005	yes
450	Mubwindi Swamp	-1.08333	29.75417	2100	terrestrial	H	2, part of 3	5 ¹⁴ C dates, 1 AMS	Low	Marchant et al., 1997	yes
451	Lake Biwa Takashima-oki Core	35.245	136.0538	85	terrestrial	E, H	2,3,4	7 ¹⁴ C dates, 1 tephra	Low	Inoue et al., 2001	yes
479	Lac Suprin	-22.18	166.59	230	terrestrial	----	2, part of 3	4 ¹⁴ C dates	High	Hope and Pask, 1998	yes

483	Supulah Hill	-4.07	138.58	1580	terrestrial	H	2, part of 3	2 ¹⁴ C dates	Low	Hope, 1998	yes
484	Wanda	-2.33	121.23	440	terrestrial	H	2, part of 3	3 ¹⁴ C dates, 1 AMS	Low	Hope, 2001	yes
488	Kohuora	-36.57	174.52	0	terrestrial	H	2, part of 3	8 ¹⁴ C, 2 AMS	Low	Newnham et al., 2007a	yes
542	MD95-2042	37.75	-10.1667	-3146	marine	E, H	2,3,4	16 AMS, 7 tie-points of planktonic $\delta^{18}\text{O}$ tuning, IRD	High	Daniau et al., 2007	yes
557	Lake Tulane	27.5862	-81.5033	35	terrestrial	H	2, part of 3	9 ¹⁴ C dates	Low	Watts and Hansen, 1988;Grimm et al., 1993	yes
560	Matsabory Ampozalana	-12.5	49.15	0	terrestrial	----	part of 2 and 3	4 dates ¹⁴ C	High	Burney, 1987	no
597	Allom Lake	-25.2333	153.1667	100	terrestrial	H	Part of 3	5 ¹⁴ C dates	High	Donders et al., 2005	yes
683	Duliha bog	51.51667	105	NOTK	terrestrial	H	2, part of 3	5 AMS dates	Low	Takahara, H. and Krivonogov, S. (unpublished data)	no
684	Cheremushuka bog	52.75	108.0833	NOTK	terrestrial	H	2, part of 3	3 AMS dates	Low	Shichi et al., 2009	yes
722	Xere Wapo C	-22.29	166.97	220	terrestrial	H	2,3,4	6 AMS dates	High	Stevenson and Hope, 2005	yes

733	Lake Hordorli	-2.533	140.55	680	terrestrial	H	2,3, part of 4	5 ¹⁴ C	Low	Hope, 1996	yes
769	Little Lake	44.16802	-123.584	703	terrestrial	H	2, part of 3	2 AMS dates	High	Long, C., (unpublished data)	yes
787	Kosipe A	-8.4667	147.2	1960	terrestrial	H	2, part of 3	2 AMS dates	Low	Hope, G. (unpublished data)	yes
788	Kosipe C	-8.4667	147.2	1960	terrestrial	H	2, part of 3	1 date ¹⁴ C	Low	Hope, G. (unpublished data)	yes
807	Xere Wapo B	-22.29	166.97	220	terrestrial	E, H	2,3,4	1 ¹⁴ C date, 2 AMS	Low	Hope and Pask, 1998; Stevenson and Hope, 2005	yes
808	Xere Wapo D	-22.29	166.97	220	terrestrial	----	3,4	2 AMS dates	High	Stevenson and Hope, 2005	yes
819	Native Companion Lagoon	-27.6754	153.4107	20	terrestrial	H	2, part of 3	10 AMS	High	Chalson, 1991	no
832	Egg Lagoon	-39.65	143.95	20	terrestrial	E, H	2,3,4	n/a	Low	D'Costa, 1997	yes
845	Lagaccione	42.56667	11.85	355	terrestrial	H	2,3,4	3 ¹⁴ C dates, 3 AMS, 1 POCO	High	Magri, 1999; Magri, 2008	yes
846	Valle di Castiglione	41.725	12.75972	44	terrestrial	H, E	2,3,4	7 ¹⁴ C dates	High	Alessio et al., 1986;	yes

										Magri, 2008	
849	MD04-2845	45.35	-5.2167	-4175	marine	E	part of 2,3,4	17 AMS dates, IRD, POCO	High	Daniau et al., 2009	yes
859	Caledonia fen	-37.3333	146.7333	1280	terrestrial	E, H	2,3,4	9 AMS, 3 OSL	High	Kershaw et al., 2007	yes
876	Wyelangta	-38.6472	143.4614	450	terrestrial	E, H	2,3,4	1 POCO	Low	McKenzie and Kershaw, 2000	yes
899	Lake Selina	-41.8833	145.6	516	terrestrial	E, H	2,3,4	1 ¹⁴ C, 2 AMS, 2 tie-points based on correlation using $\delta^{18}O$ and paleomag.	Low	Colhoun et al., 1999	yes
900	Lombok Ridge Core G6-4	-10.7833	118.0667	-3510	marine	E, H	2,3,4	1 AMS, 3 tie-points based on MIS boundary	Low	Wang et al., 1999	yes
901	Lake Coomboo	-25.2195	153.1959	90	terrestrial	E, H	2,3,4	3 ¹⁴ C	Low	Longmore, 1997	yes
902	Lynchs Crater	-17.3667	145.7	760	terrestrial	H	2,3,4	2 ¹⁴ C	High	Kershaw, 1986	yes
1021	Lake Surprise	-38.0612	141.9223	93	terrestrial	H	1,2, part of 3	5 ¹⁴ C dates	Low	Builth et al., 2008	yes
1057	Tyrendarra Swamp	-38.1986	141.7626	13	terrestrial	H	Part of 3	3 AMS dates	High	Builth et al., 2008	yes
1101-1102	Lake Wangoom LW87 (Top Core and	-38.35	142.6	100	terrestrial	E, H	2,3,4	1 AMS, 2 UTh	Low	Harle et al., 2002	yes

	bottom core)										
1104	ODP 1078C	-11.9167	13.4	-426	marine	H	2, part of 3	13 AMS	Low	Dupont et al., 2008	yes
1105	MD97-2140	2.0667	142.2667	-2547	marine	E, H	2,3,4	4 tie-points based on MIS boundary, correlation with radiocarbon dated from other close core	Low	Thevenon et al., 2004	no

Fig. 1: Location of marine and terrestrial sites with charcoal records covering all or part of the last glacial. The sites are classified according to sampling resolution: high (> 1 sample per ka) indicated by black circles, or low (< 1 sample per ka) indicated by open circles (See Table 1).

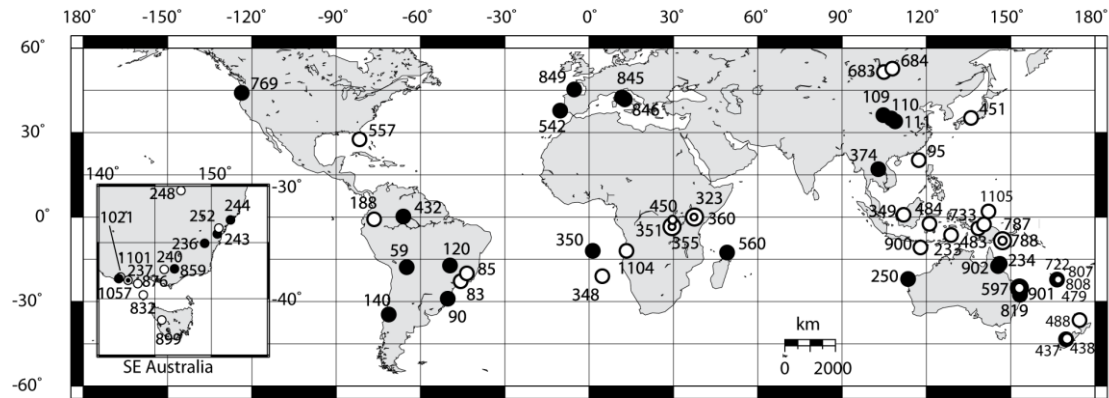


Fig. 2: Box plots of transformed (see text) charcoal data for (a) global, (b) northern hemisphere, and (c) southern hemisphere during the glacial compared to the Eemian and Holocene, and for MIS 4, 3 and 2. The lower and upper limits of the box represent the 25th and 75th percentiles, respectively.

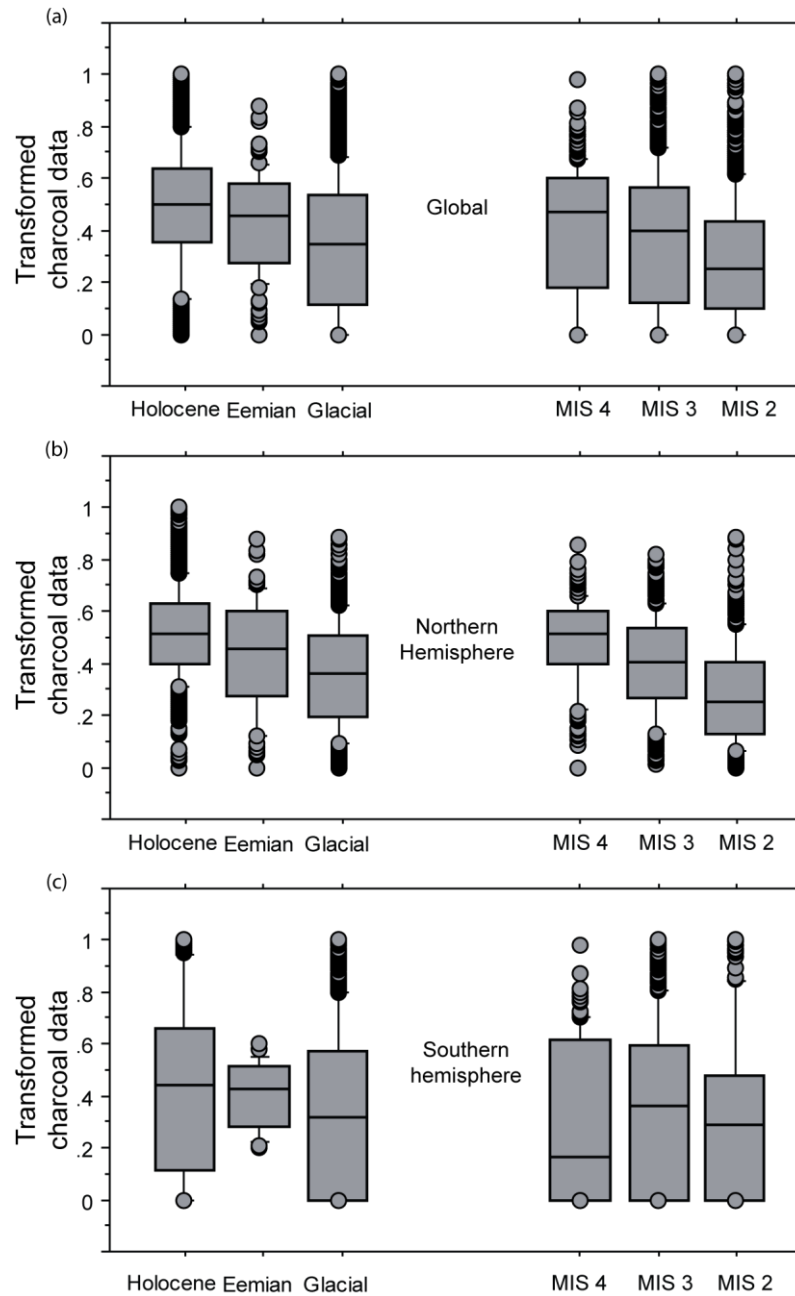


Fig. 3: Changes in transformed charcoal concentration/abundance for the 30 high-resolution records covering the last glacial, arranged from north (top) to south (bottom). The sites are designated by their site number (see Table 1) and continent (North America: NAM; South America: SAM; Europe: EUR; Australasia: AUST; Africa: AFR; Asia: ASIA). Grey bands indicate MIS 2 and 4. The chronological position of the D-O warming events, using the GICC05 chronology (Wolff et al., this volume), are shown by red lines and the number of the corresponding GIs is shown.

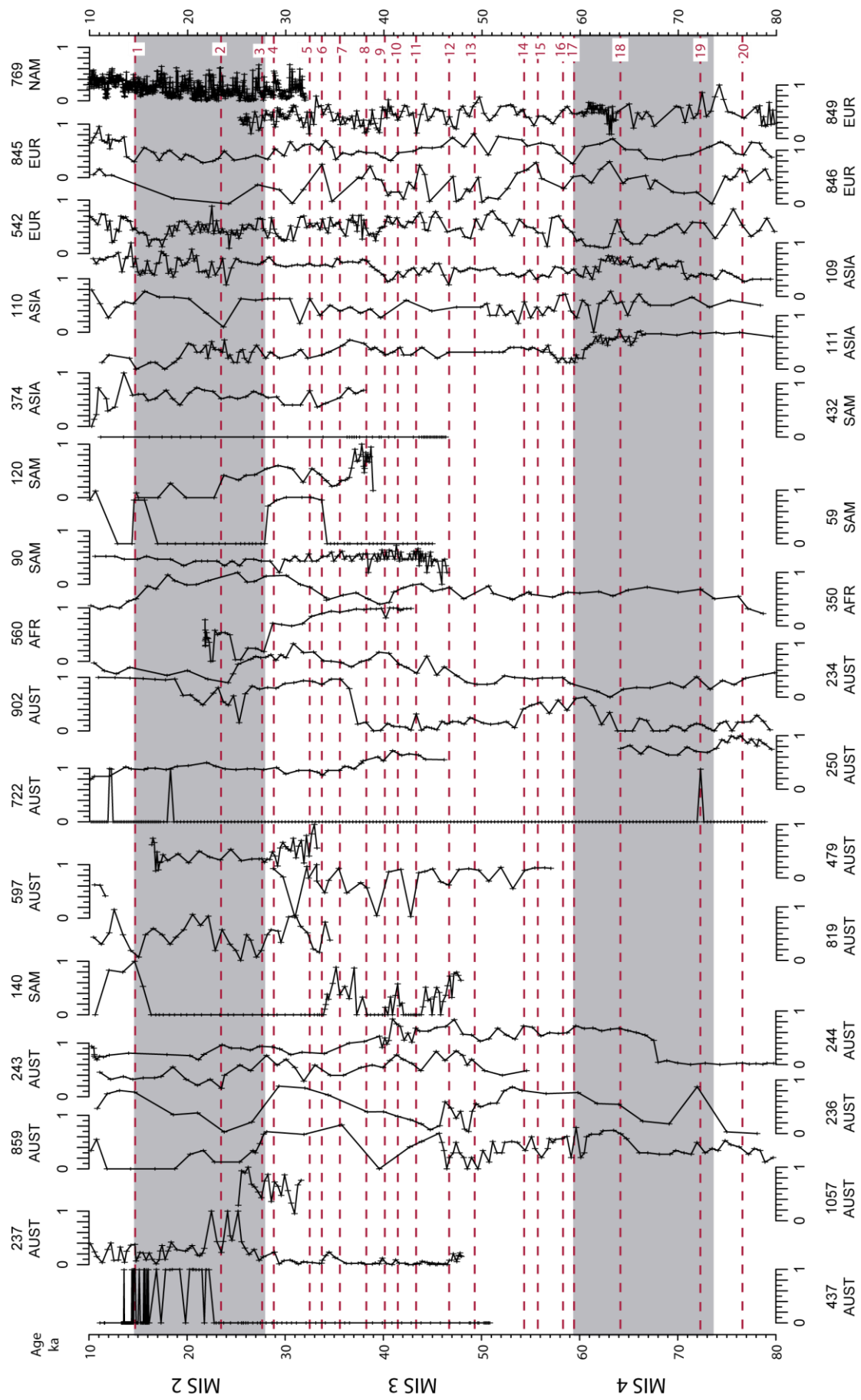


Fig. 4: Reconstruction of global biomass burning through the glacial based on the 30 high-resolution sites. Charcoal data are summarized using a lowess curve (middle) with a 400-yr half-window width (red), and to represent the long-term trend, a 2000-year window width (gray). The number of individual charcoal records that contribute to the summary curves is plotted at the bottom. For comparison with the charcoal curve, the 20-yr sampling resolution NGRIP record (age yr b1950) is shown, along with a 400-yr smoothed curve (blue).

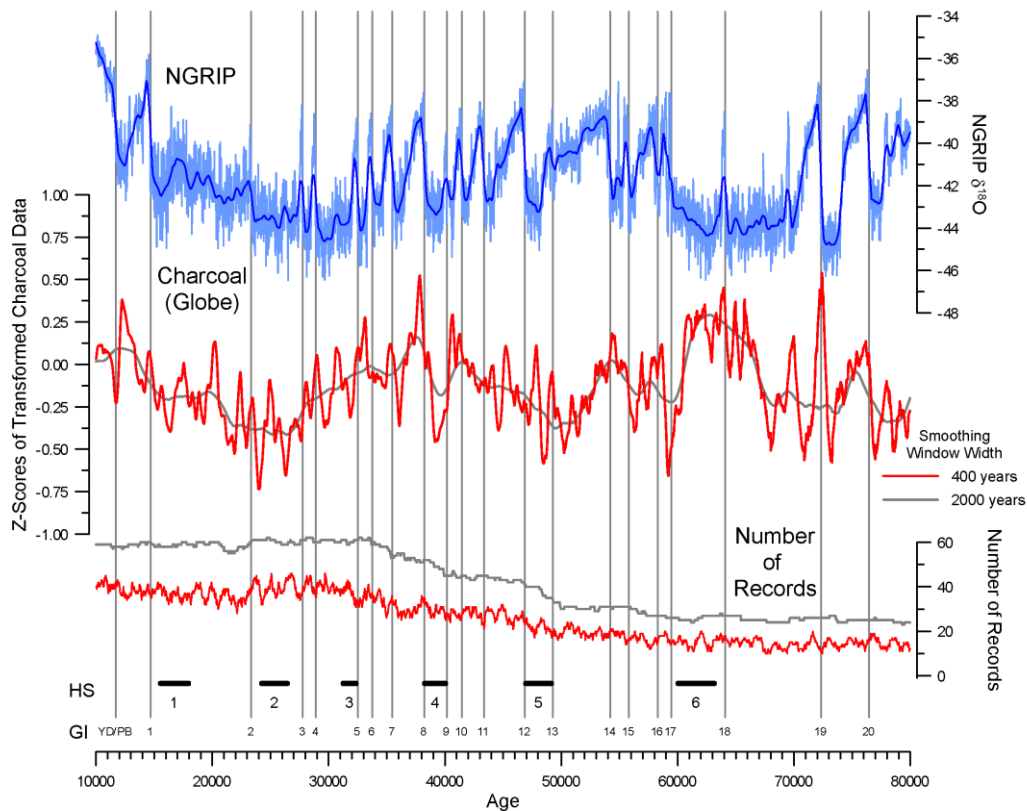


Table 2: Start of D-O warming events, most rapid cooling and Heinrich stadials following Sánchez-Goñi and Harrison (this volume), Wolff et al.(this volume) and this study. Ages are given in ka b1950.

Start of which GI	Start of GIs	Start of most rapid cooling	Start of HS
1	14.642	14.150	
2	23.290	23.250	HS 1 = 18.0
3	27.730	27.590	HS 2 = 26.5
4	28.850	28.630	
5	32.450	32.130	HS 3 = 32.7
6	33.690	33.410	
7	35.430	34.750	
8	38.170	37.450	
9	40.110	39.970	HS 4 = 40.2
10	41.410	40.990	
11	43.290	42.670	
12	46.810	44.350	
13	49.230	48.630	HS 5 = 50.0
14	54.170	n/a	
15	55.750	55.410	
16	58.230	56.650	
17	59.390	58.590	
18	64.045	63.850	HS 6 = 63.2
19	72.280	70.970	
20	76.400	74.270	

Fig. 5: Superposed epoch analysis (SEA) composites illustrating changes in biomass burning associated with D-O warming event (A and B), D-O cooling event (C and D) and Heinrich Stadials (E and F) with x-axis corresponding to time (before event on the right and after event on the left). The confidence intervals are given by Monte-Carlo simulation using the timing of the 20 D-O warming events, 19 D-O cooling events and the 6 Heinrich Stadials. The NGRIP $\delta^{18}\text{O}$ SEA composite record for the D-O cycles and Heinrich events is shown for comparison.

