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The prehistory of fire in Australasia

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Introduction

Understanding what controls the occurrence and severity of wildfire is crucial to managing fire regimes in changing climates. The direct observational record (e.g. from satellite or remote-sensed data) is very short (~10 years) and although historical records of fire from eastern Australia extend back to the mid-20th century, they only encompass a limited range of climate variability. Furthermore, the interpretation of historical records can be complicated, for example, by the recent and widespread human modification of the landscape. Extending the record back in time, to periods when the climate was radically different from present or when large changes in regional climate occurred over a few decades, is therefore important to improving our understanding of the interactions between climate, vegetation and fire regimes.

Palaeoenvironmental analyses use natural 'archives' that build up sequentially to 'reconstruct' aspects of the past and so allow consideration of environmental change. These 'archives' include the sediments found in lakes, swamps and ocean basins, as well as ice-cores and tree rings. Waterlogged, and hence anoxic, sediments accumulate in some depositional environments such as swamps and lakes and incorporate and preserve micro- and macroscopic fossils. Depth within the sediment column is also assumed to be a function of time when dating techniques, such as radiocarbon dating, are used. Analyses of undisturbed, accumulating sediments provides insight into the processes and trends over time.

Palaeoenvironmental reconstruction can be based on any indicator that is sensitive to the environmental condition or process under examination. The abundance of charcoal in sediments is widely used as an indicator of past fire activity (Clark 1982, 1983; MacDonald *et al*. 1991; Patterson *et al*. 1987) and has been interpreted as a 'fire history' at more than 900 sites across the globe (Global Palaeofires Database 2009). Charcoal is an inorganic carbon compound produced from the incomplete combustion of organic material (Clark 1984; Patterson *et al*. 1987). The resistant chemical nature of charcoal means that it is relatively easy to quantify in sediments using a variety of techniques (Conedera *et al*. 2009).

The analysis of charcoal from sedimentary records provides insight into the past occurrence of fire over much longer time spans than is available from other sources, such as the historic or instrumental record. This longer temporal perspective on fire is not just limited to answering historic questions (what happened when?) but allows current trends in fire activity to be placed in context (are modern fire regimes unusual?) and provides scenarios that are applicable to characterising fire activity resulting from future climate change (e.g. what happens to fire frequency during hotter or drier climates?). Quantitative techniques aimed at calculating the frequency of fire events from the sedimentary record have also provided baseline information, such as the 'normal' fire interval, which can be useful for natural resource management, emergency services and planning (e.g. Whitlock *et al*. 2003).

Reviews of the literature exploring the theories of charcoal production and its subsequent taphonomy can be found in Tolonen (1986), Patterson *et al*. (1987), Whitlock and Larsen (2001), Whitlock and Bartlein (2004) and most recently in Conedera *et al*. (2009). The quantification of charcoal provides information on past fire activity within a spatial scale dependent on the size of the charcoal fraction quantified (Mooney and Tinner 2011). It is generally assumed that charcoal abundance is related to the amount of biomass burnt. However, charcoal records cannot be used to provide quantitative estimates of the amount of biomass burnt; rather they are generally interpreted in terms of relative changes (see e.g. Haberle *et al*. 2001; Carcaillet *et al*. 2002; Power *et al*. 2008; Marlon *et al*. 2009; Daniau *et al*. 2010).

The 'traditional' approach to the reconstruction of past fire activity is an extension of palynology, which seeks to reconstruct vegetation and the reasons for vegetation change. Due to their size, charcoal particles quantified in pollen preparations using microscopy reflect fire from all scales, up to and including a regional or extra-regional source area (Clark 1988; Tinner *et al*. 1998). Larger charcoal particles (typically >100 µm in length) travel much shorter distances than the charcoal encountered on a typical pollen slide and so reflect fire at a local scale (Whitlock and Millspaugh 1996; Clark *et al*. 1998). This size fraction of charcoal particles can be isolated with wet sieving and easily quantified under a stereo-microscope. Nevertheless, it has been shown that both microscopic and macroscopic records produce comparable results in terms of broad-scale regional histories of fire (Tinner *et al*. 2006; Conedara *et al*. 2009).

Although there have been several papers dealing with the fire history of Australasia (see e.g. Singh *et al*. 1981; Haberle *et al*. 2001; Kershaw *et al*. 2002; Lynch *et al*. 2007), the most comprehensive survey of the charcoal records to date is provided by Kershaw *et al*. (2002), who provided a synthesis and review of 70 Australian charcoal records. Two of these records stretch back to the Tertiary and reveal increasing fire activity through the Cenozoic as environmental conditions became drier and more variable. These records are important because they suggest that fire was already a component of landscape dynamics in Australia before the onset of the dramatic climatic fluctuations characteristic of the Quaternary. Only nine sites reviewed in Kershaw *et al*. (2002) encompass the last glacial/interglacial cycle (covering around 80 000 years), but they reveal an apparently strong relationship between climate and fire.

Most of the sites (58) in the Kershaw *et al*. (2002) synthesis are of Holocene age. These sites show four marked periods of fire activity: an interval of low fire activity from 7–5 ka; an increase in fire across all biomes except wet forests after 5 ka; a further increase after 2 ka; and maximum fire activity post-dating the arrival of European settlers. The 'late European' period (as defined by Kershaw *et al*. 2002 Figure 1.6) was conspicuous because it was characterised by lower levels of burning than any other time in the Holocene.

The relative contribution of people to changes in the Australian fire regime has been a contentious issue. Kershaw *et al*. (2002) argued that, although it is difficult to disentangle the relative contributions of climate and people, the climate-controlled trajectory of increasing fire during the late Pleistocene appears to have been enhanced by people. They were also of the opinion that a balance had been achieved between the activities of people and vegetation change by the Holocene, and hence all major changes in vegetation and burning during this period were consistent with inferred patterns of climate change.

Changes in biomass burning can be brought about in many ways, through changes in fire weather, the incidence of ignitions, the availability and dryness of fuel, and through changes in fuel type. Each of these factors is influenced by climate, but climate also determines the nature of the vegetation, which in turn affects fuel characteristics and availability. Human activities can affect fire regimes through changing the incidence of ignitions, but perhaps more importantly by affecting the nature of the vegetation, the degree of landscape fragmentation and fuel loads. Disentangling these various influences can be problematic (Kershaw *et al*. 2002), particularly at individual sites or when the records are short, and much of the interpretation of individual palaeo-fire records has been based on the perceived coincidence in time of changes in potential driving factors. We have adopted a somewhat different approach, by synthesising charcoal records at a continental scale and looking for recurring patterns in the fire response to external forcings.

The Australasian charcoal database

We have analysed charcoal records from Australasia, here broadly defined to include tropical south-eastern Asia, New Guinea, New Zealand and the islands of the western Pacific (20°N–50°S, 100°E to 177°W). Sites from South-East Asia and New Guinea were included to set the sparse records from tropical Australia in a broader context. Sites from the western Pacific were included to assess the potential role of El Niño-Southern Oscillation (ENSO) on fire regimes (see e.g. Lynch *et al* . 2007) and partly because this region has a different settlement history from Australia (e.g. Stevenson and Hope 2005; Sutton *et al*. 2008; Wilmshurst *et al*. 2008).

Our analysis is based on a synthesis of 224 sites (Figure 1.1), derived from the Global Palaeofire Working Group (GPWG 2010) charcoal database (Version 2.5: Mooney *et al*. 2011). The charcoal records from Australasia were obtained using a variety of techniques although the quantification of charcoal on pollen slides is the most common method, accounting for approximately 80% of the records. This conforms to the generalisation by Patterson *et al*. (1987) who noted that charcoal is most commonly isolated or concentrated from sediments and then quantified using various optical techniques. Approximately 44 sites in the Australasian dataset (~20%) are macroscopic charcoal records obtained by wet sieving: these account for a higher proportion of the more recent and more continuous sequences, which mirrors trends in North America and Europe (e.g. Carcaillet *et al*. 2001; Whitlock and Bartlein 2004; Conedera *et al*. 2009). Digestion methods, which determine the mass of elemental carbon (e.g. Winkler 1985; Bird and Cali 1998; Kurth *et al*. 2006) have not been applied to Australian palaeoenvironmental sequences.

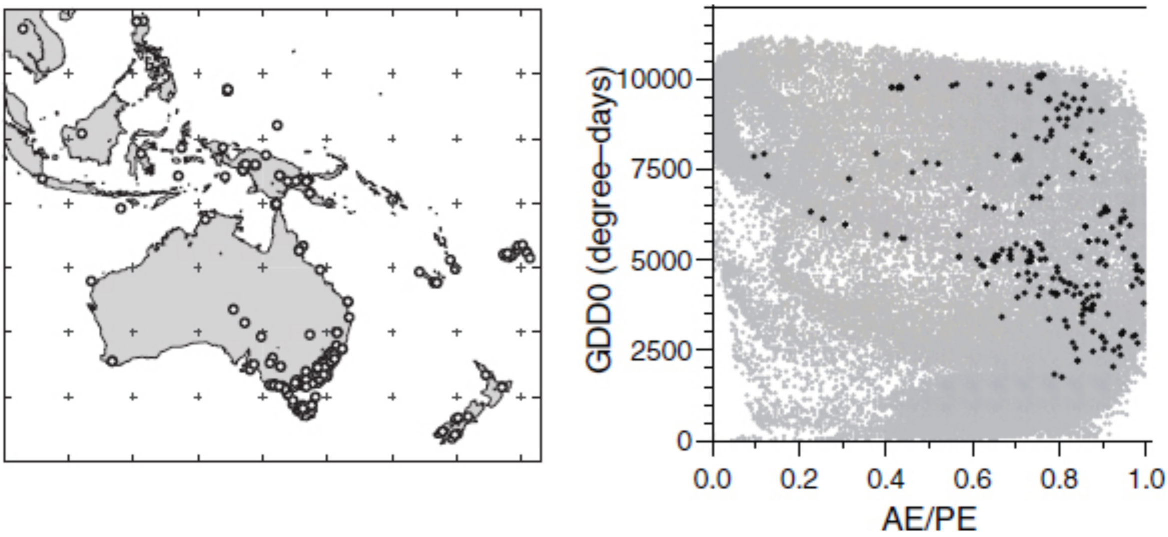

Figure 1.1. Distribution of sedimentary charcoal sites in the Australasian region (20°N–50°S and 100°E to 177°W) in geographic space and climate space. The climate space used here is the global climate space from Figure 1.2. The charcoal records are derived from version 2.5 of the Global Palaeofire Working Group (GPWG) global charcoal database (Mooney *et al*. 2011).

Because charcoal is quantified using many different techniques and expressed using a large range of metrics, the data were standardised to facilitate comparisons between sites and through time (see Power *et al*. 2008, 2010). Standardisation was achieved using three steps. First, non-influx data (e.g. charcoal concentration expressed as particles/cm^3; charcoal-to-pollen ratios) were transformed to influx values (e.g. particles/cm^2/year), or quantities proportional to influx, by dividing the values by sample deposition times. Poor chronological control is a limitation for a number of sites included in the synthesis: this results in coarse age models (and estimates of deposition time) and hence charcoal accumulation is probably best considered at some sites as estimated influx values.

The second step in the standardisation process was the application of a Box-Cox transformation to homogenise the inter-site variance by transforming individual charcoal records toward normality. Finally, the data were rescaled using a common base period (0.2–21 ka) to yield z-scores, so that all sites have a common mean and variance. Previous analyses have shown that the choice of the base period does not affect the results significantly (Marlon *et al*. 2008; Power *et al*. 2010) and standardisation does not alter the overall pattern of variability or 'signal' in a record.

The z-scores can be plotted as maps, to examine the spatial patterns of changes in fire regime during particular intervals or composited to produce time series showing the evolution of biomass burning through time in a particular region. Composite curves for specific regions have been produced by fitting a locally weighted regression (or 'lowess' curve) to the pooled standardised (transformed and rescaled) data using a fixed window width and a tricube weight function with one 'robustness iteration'. We used window widths of various lengths to emphasise different scales of temporal variability: short windows (e. g. of 100 or 200 years) were used to resolve centennial variability in the past 2000 years and much longer windows (e.g. 2000 years) were used for multi-millennial variability over longer time periods.

Confidence intervals for each composite curve were generated using bootstrap re-sampling (with replacement) of individual sites with 1000 replications. Confidence limits for each target point were taken as the 2.5th and 97.5th percentiles of the 1000 fitted values for that target point. This conservative approach identifies the uncertainty in regression curves that arises from the inclusion or exclusion of the entire record from an individual site: when the bootstrap confidence intervals are wide, this indicates greater uncertainty in the composite curve and greater sensitivity of that curve to the addition/subtraction of an individual record. Furthermore, when one confidence limit deviates from the composite curve more than the other confidence limit, this indicates that some individual records depart markedly from the typical pattern of the time. Minor fluctuations in the composite curve during periods when the bootstrap confidence intervals are wide, or when the composite curve is not located in the middle of the confidence limits, are probably meaningless. The number of observations contributing to the composite curve at each of the target points that define the curve is also provided, and serves as an additional measure of the confidence to be placed in the reconstructions.

The strength of combining multiple records, as here, is that it allows a robust assessment of the influence of individual records, and hence of the chronological uncertainties of each record, on the shape of the composite curve. When interpreting individual records, the quality of the age model is of prime importance: here, by focusing on statistically robust features of the composite record, we are able to draw plausible inferences about the controls on these features. This approach is analogous to the production of composite-anomaly maps in synoptic climatology (e.g. Yarnal *et al*. 2001; Shinker *et al*. 2006) or multi-model ensembles to diagnose robust features of climate-change simulations (e.g. Cai and Cowan 2006; Braconnot *et al*. 2007; Meehl *et al*. 2007).

Changes in fire activity through time

The charcoal data for all 224 sites in the data set is graphically summarised in a Hovmöller diagram of z-scores by latitude versus time (Figure 1.2). Hovmöller diagrams (Hovmöller 1949) are two-dimensional plots showing how the value of some attribute varies in space-time; one axis refers to time and the other to spatial location. They are particularly useful for displaying large amounts of data in a meaningful and understandable form and are thus used routinely in meteorology to map such things as climate anomalies (see e.g. von Storch *et al*. 1998; Woolnough *et al*. 2001).

Figure 1.2 clearly depicts the latitudinal and chronological range of the data and it is clear that the dataset displays a high degree of temporal variability and between-site heterogeneity. Table 1.1 shows the number of records in our analysis during the Marine Isotope Stages of the last 130 000 years. Marine Isotope Stages are defined on the basis of changes in the oxygen isotope ($^{18}O/^{16}O$) ratios of foraminifera shells from marine sediments, which result from changes in ocean geochemistry concomitant with changes in global ice volume and thus, broadly speaking, correspond to warmer (interglacial or interstadial) and colder (glacial or stadial) periods. Oxygen isotope stratigraphy from marine records provides a global 'signature' and hence is widely used as a chronological tool (Sanchez Goñi and Harrison 2010).

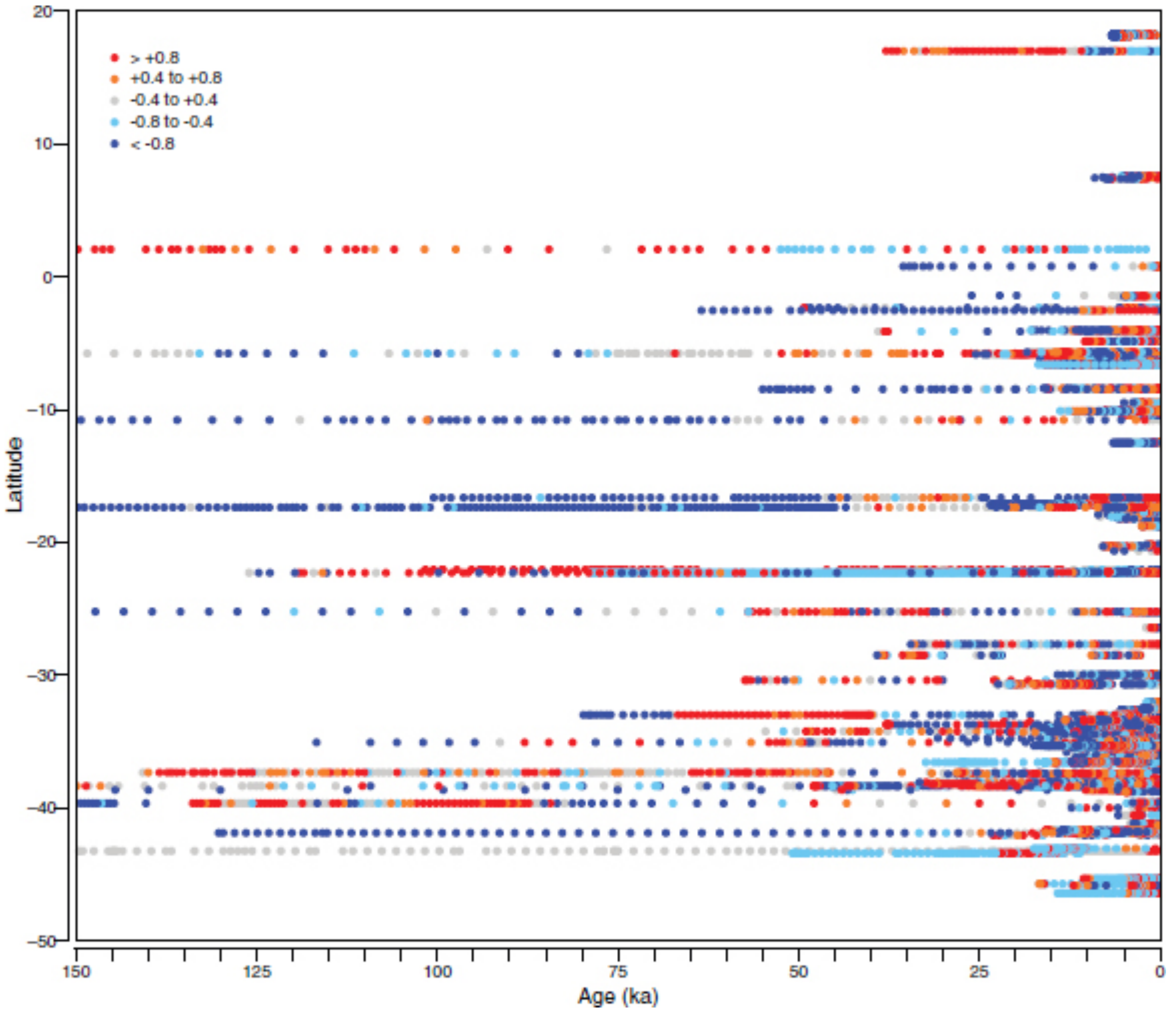

Figure 1.2. Hovmöller plot of changes in biomass burning by latitude in the Australasian region (20°N–50°S and 100°E to 177°W) over the past 150 ka based on data from V2.5 of the Global Palaeofire Working Group (GPWG) global charcoal database (Mooney *et al*. 2011) The individual charcoal records are standardised and rescaled using a common base period (0.2–21 ka) to yield z-scores, so that all sites have a common mean and variance.

Table 1.1. Definition of Marine Isotope Stages used in this chapter. The stage boundaries are from Sanchez Goñi and Harrison (2010). The number of charcoal sites recording at the beginning and end of each stage is also shown.

Marine Isotope Stage	Age (years before present)	Number of sites (at start/at end)
MIS 5	130 000–73 500	12/15
MIS 4	73 500–59 400	15/20
MIS 3	59 400–27 800	20/38
MIS 2	27 800–14 700	28/70

MIS 1	14 700–present	70/124

Although there are only 12 records in our dataset that extend back to the last interglacial (circa 130–73.5 ka), these records span the latitudinal range from the tropics to the far south of the region (Figure 1.2, Table 1.1). The number of sites rises steadily through time, such there are 15 sites from the beginning of the last glacial (73.5 ka, following Sanchez Goñi and Harrison 2010), increasing to 20 sites at the beginning of Marine Isotope Stage (MIS) 3 (59.4 ka), 38 sites by the end of MIS 3 (27.8 ka), circa 70 sites by the end of MIS 2 (14.7 ka) and more than 140 sites by 10 ka. Australasia is one of the best documented regions of the world during the glacial interval (see Daniau *et al*. 2010) and contributes more than a quarter of the Holocene records in the GPWG global charcoal database (Mooney *et al*. 2011).

The finer scale structure in these records can be examined through the construction of composite curves. The composite curve covering the glacial (Figure 1.3) suggests a broad tendency for higher levels of biomass burning during the interstadial conditions of MIS 3 (59.4–27.8 ka) than during either of the bracketing colder stadials, MIS 4 (73.5–59.4 ka) and MIS 2 (27.8–14.7 ka). Lake-level evidence from south-eastern Australia shows that MIS 3 was characterised by cool and relatively moist conditions (e.g. Bowler 1981, 1986; Wasson and Clark 1988). Conditions become increasingly drier and colder from about 30 ka onwards (Allan and Lindsay 1998) and a corresponding decrease in biomass burning is evident (Figure 1.3).

However, the most striking feature of the Australasian composite curve in Figure 1.3 is the millennial-scale variability. The timing, number and shape of the peaks in this composite resembles the pattern of millennial-scale variability found in Greenland ice-core data (Wolff *et al*. 2010), as opposed to the smaller number of lower amplitude ‘Antarctic Isotope Maxima’ registered over the same interval (EPICA Community Members 2006; Jouzel *et al*. 2007). This implies that biomass burning across Australasia increases during Dansgaard-Oeschger (D-O) warming events and decreases during the subsequent cooling phases (i.e. the Greenland stadials). This correspondence is consistent with a recent global analysis of charcoal records that span the interval from 80 ka to 10 ka (Daniau *et al*. 2010), suggesting that biomass burning increases across D-O warming events. Harrison and Sanchez Goñi (2010) and Mooney *et al*. (2011) have argued that immediate atmospheric teleconnections (changes in atmospheric circulation) in response to changes in atmospheric composition link terrestrial climates in Northern and Southern Hemispheres on D-O timescales.

Although the longer-term and millennial-scales changes in biomass burning across Australasia during the glacial appear to reflect a global signal of temperature changes, this is less apparently the case during the transition from the Last Glacial Maximum (LGM) to the current interglacial (Figure 1.4). The composite charcoal record for the last 22 000 years shows a gradual increase from the low levels of biomass burning at the LGM, with a more rapid increase synchronous with the warming shown in the EPICA ice core record from about 17 000 years ago. There is a sharp decline in biomass burning at the beginning of the Antarctic Cold Reversal (ACR, 14.7 ka), and the longer term trend shows fire remaining relatively low until the beginning of the Holocene, after which a slight trend towards increasing fire is apparent.

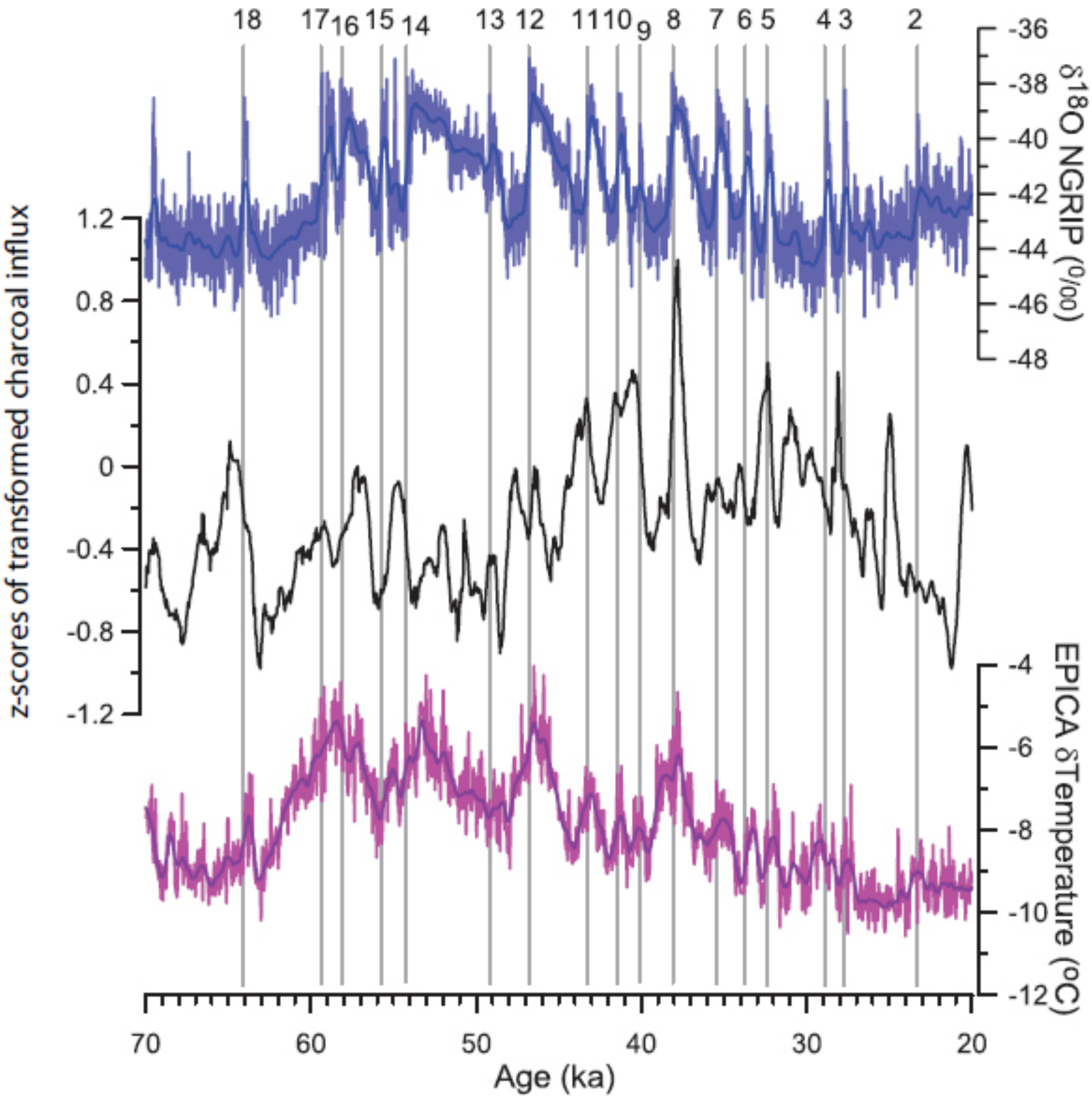

Figure 1.3. Reconstruction of biomass burning over the period from 70 to 20 ka for the Australasian region (20°N to 50°S; 100°E to 177°W). Biomass burning is represented by a composite curve with a smoothing window of 400 years (bold black curve) to emphasise centennial- to millennial-scale variability. The composite biomass-burning curve is compared with indices of polar temperature: the deuterium record from Antarctica (EPICA) and δ18O from Greenland (NGRIP). The ice core data were obtained from the World Data Center for Palaeoclimatology hosted by NOAA/NGDC (http://www.ncdc.noaa.gov/paleo/paleo.html).

The record of biomass burning over the last 2000 years (Figure 1.5) is relatively flat, with an apparent increase centred on about AD 400, less fire at about AD 550–700 and then a marked increase beginning at about AD 1500, which culminates in a maximum at about AD 1900. The Australasian records do not show the temperature-driven decrease in biomass burning during the two millennia before the industrial revolution evident at a global scale (Marlon *et al*. 2008), nor do they appear to register the Little Ice Age, which is seen in the global record of biomass burning as a distinct trough (Marlon *et al*. 2008). Biomass burning is relatively high across Australia during this period and in particular during the peak of cold temperatures in some areas of the Northern Hemisphere from 1600 to 1800 AD (Jansen *et al*. 2007). Another widely discussed climatic anomaly of the last millennia – the Medieval Warm Period, which affected some areas of the globe in the period between AD 950 and 1250 (Mann *et al*. 2009) – is also not an obvious feature of the Australasian record of biomass burning (Figure 1.5).

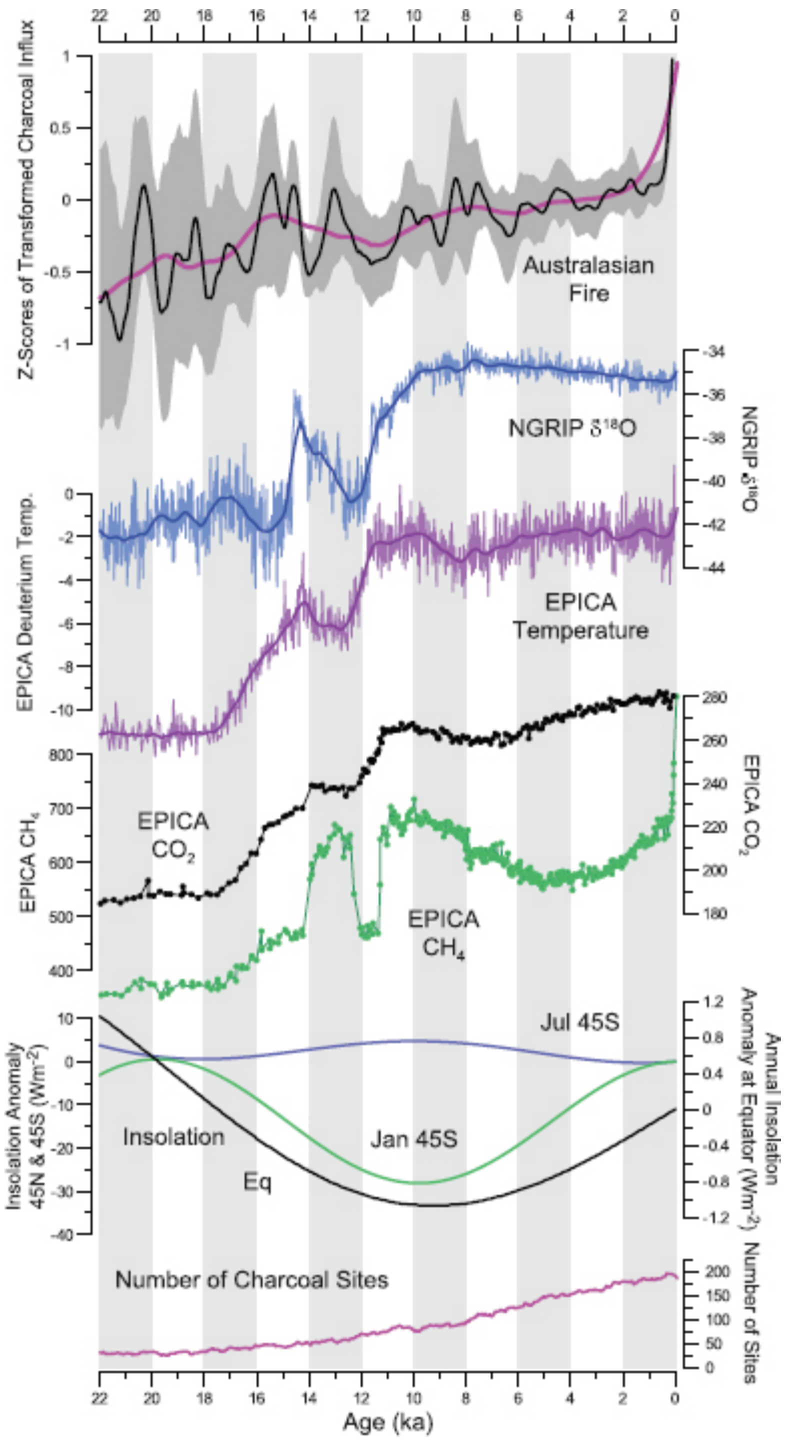

Figure 1.4. Reconstruction of biomass burning over the period from 22 ka to present for the Australasian region (20°N–50°S; 100°E to 177°W). The composite charcoal record has been smoothed using a window of 2000 years (red curve) to emphasise the long-term trends and 400 years (bold black curve) to emphasise centennial- to millennial-scale variability. The bootstrapped confidence intervals are based on the 400-year smoothing of the curve. The number of sites contributing to the 400-year composite curve is shown at the bottom of the figure. The composite biomass burning curves are compared with changes in January and July insolation at 45°S and annual insolation at the equator, CO_2 and CH_4 (from the EPICA ice core: Lüthi *et al*. 2008; Loulergue *et al*. 2008), and records of polar temperature changes from Antarctica and Greenland (deuterium from EPICA, δ18O from NGRIP). The ice-core data were obtained from the World Data Center for Palaeoclimatology, NOAA/NGDC (http://www.ncdc.noaa.gov/paleo/paleo.html).

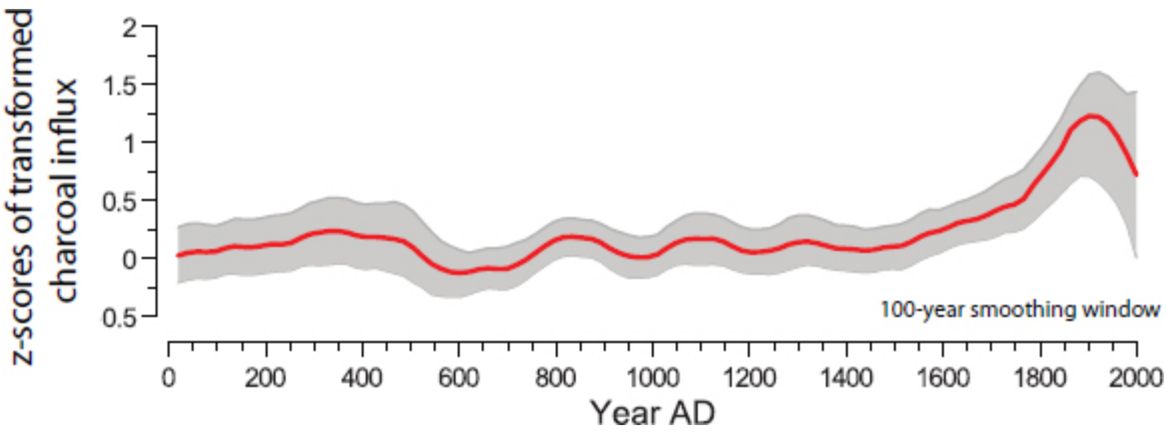

Figure 1.5. Reconstruction of biomass burning for the last 2000 years for Australasia (20°N–50°S, 100°E–177°W). The curves have been smoothed using a window of 100 years (bold red curve).

A distinct increase in Australasian biomass burning occurs at about 1800 AD (Figure 1.5), which culminates in a peak at 1900 AD and a gradual decline during the 20th century. The peak in biomass burning at 1900 AD is the largest of the entire record, irrespective of the temporal length or of the smoothing window used. A change in biomass burning from the late 18th century is not unexpected in some regions of Australia, given the impact of European settlement and the land clearing associated with agricultural development. However, this change in land use is not synchronous across the entire Australasian region.

In their global synthesis, Marlon *et al*. (2008) described changes similar to those evident over the last few centuries in the Australasian region (Figure 1.5). This included a sharp increase in biomass burning from AD 1750, which peaked at AD 1870, after which global biomass burning declined. The decline in biomass burning during the 20th century has been confirmed by ice-core measurements of carbon monoxide concentration and stable isotopic composition, and specifically attributed to a decline in Southern Hemisphere biomass burning (Wang *et al*. 2010). Marlon *et al*. (2008) suggested that the decrease in biomass burning during the 20th century, apparent in their analyses at both global and continental scales, was a result of fire suppression through landscape fragmentation consequent on agricultural expansion. There has been no quantitative analysis to determine whether landscape changes were sufficiently large to explain the observed 20th century decline in the composite biomass burning curve for Australasia.

Spatial patterns of change in biomass burning

Although consistent conclusions about biomass burning through time emerge from the composite curves described above, there are differences between the records from individual sites (as shown by the bootstrap uncertainties on these curves). Regional coherency is most apparent during intervals of more extreme climates. During more benign climates, some spatial variability in biomass burning appears to relate to differing responses to climate changes.

As an example: the LGM at circa 21 ka was characterised by low biomass burning in the tropics and over most of Australia, with only a few sites in south-eastern Australia and a single site on the South Island of New Zealand showing higher-than-average z-scores for this period (Figure 1.6). A similarly homogenous pattern of reduced burning across extra-tropical Australasia is apparent during the Antarctic Cold Reversal (ACR) (see Mooney *et al*. 2011). Power *et al*. (2008) and Harrison *et al*. (2010) have argued that the reduction in biomass burning during extreme cold, dry intervals such as the LGM or the ACR reflects reduction in vegetation productivity and hence a reduction of fuel loads.

Pollen-based reconstructions of Australian vegetation during the LGM (see e.g. Harrison and Dodson 1993; Dodson 1994; Hope 1994; Kershaw 1995, 1998; Harle 1997; Pickett *et al*. 2004), although differing in detail, all indicate expanded areas of semi-arid and open vegetation during the LGM. Geomorphic evidence, such as the expansion of areas of mobile sand dunes (Hesse *et al*. 2004), also suggests a reduction in tree cover over large parts of the Australian continent. Thus, it is plausible to explain the low biomass burning during this interval as a reflection of vegetation changes that resulted in a reduction in fuel availability.

In contrast, the regional patterns of biomass burning are very different by the mid-Holocene (circa 6 ka, Figure 1.6). Although most sites in the tropics, northern Australia, the Pacific Islands and New Zealand are characterised by low biomass burning, south-eastern Australia is marked by high biomass burning. Within each of these regions, however, there are sites that show an opposite trend. To some extent, this heterogeneity can be explained by the more modest changes in climate during the Holocene. Reconstructions of mid-Holocene vegetation changes, for example, show only rather small changes compared with today and have been interpreted as evidence for comparatively muted regional climate changes (Markgraf *et al*. 1992; Harrison and Dodson 1993; Pickett *et al*. 2004).

In the case of south-eastern Australia, the large number of sites makes finer scale mapping possible (Figure 1.7). This reveals a spatial coherency in the changes of biomass burning at a sub-regional scale. Thus, throughout the Holocene there is an opposition in the temperate latitudes between sites in southernmost south-eastern Australia (including Tasmania) and southeastern New South Wales (including the interior). Pickett *et al*. (2004) showed that vegetation changes in the western part of south-eastern Australia during the mid-Holocene indicated drier conditions and contrasted with records of wetter conditions from the Snowy Mountains and east of the Dividing Range. The fire records suggest that this opposition, which is consistent with the synoptic rainfall patterns associated with the southern annular mode (Meneghini *et al*. 2007), occurred throughout the Holocene. Nevertheless, between-site variability is high, suggesting that local factors are important in mediating changes in fire regimes.

Controls on biomass burning

Our conceptual framework for interpreting the palaeofire record is based on analyses of the modern relationships between the various factors influencing fire and burnt area. At both a global and at continental scales (Figure 1.8), climate (temperature and moisture) is a major control on biomass burning. Fire is prominent in warm climates (here represented by GDD0, the accumulated temperature sum during the growing season) with intermediate levels of moisture availability (here represented by the ratio of actual to equilibrium evapotranspiration, AE/PE). In wet climates (AE/PE >0.8), although vegetation productivity (here represented by net primary productivity, NPP) is high, the vegetation and litter are too wet to combust. In dry climates (AE/PE <0.2) and cold climates (GDD0 <5000º/days), vegetation productivity is reduced and hence fuel load becomes limiting (see also van der Werf *et al*. 2006; Krawchuk and Moritz 2009). Vegetation distribution (here represented by major vegetation types, megabiomes as defined by Harrison and Prentice 2003) is also determined by climate, and the greatest area burnt occurs in open forest, savannas and dry woodlands, and xerophytic grasslands. Natural ignition sources (e.g. lightning) are present across the range of climates and do not appear to influence burnt area. Although the area of climate space represented in Australasia (and Australia) is limited, the pattern of relationships between area burnt, climate, vegetation productivity and ignitions is identical to that observed globally.

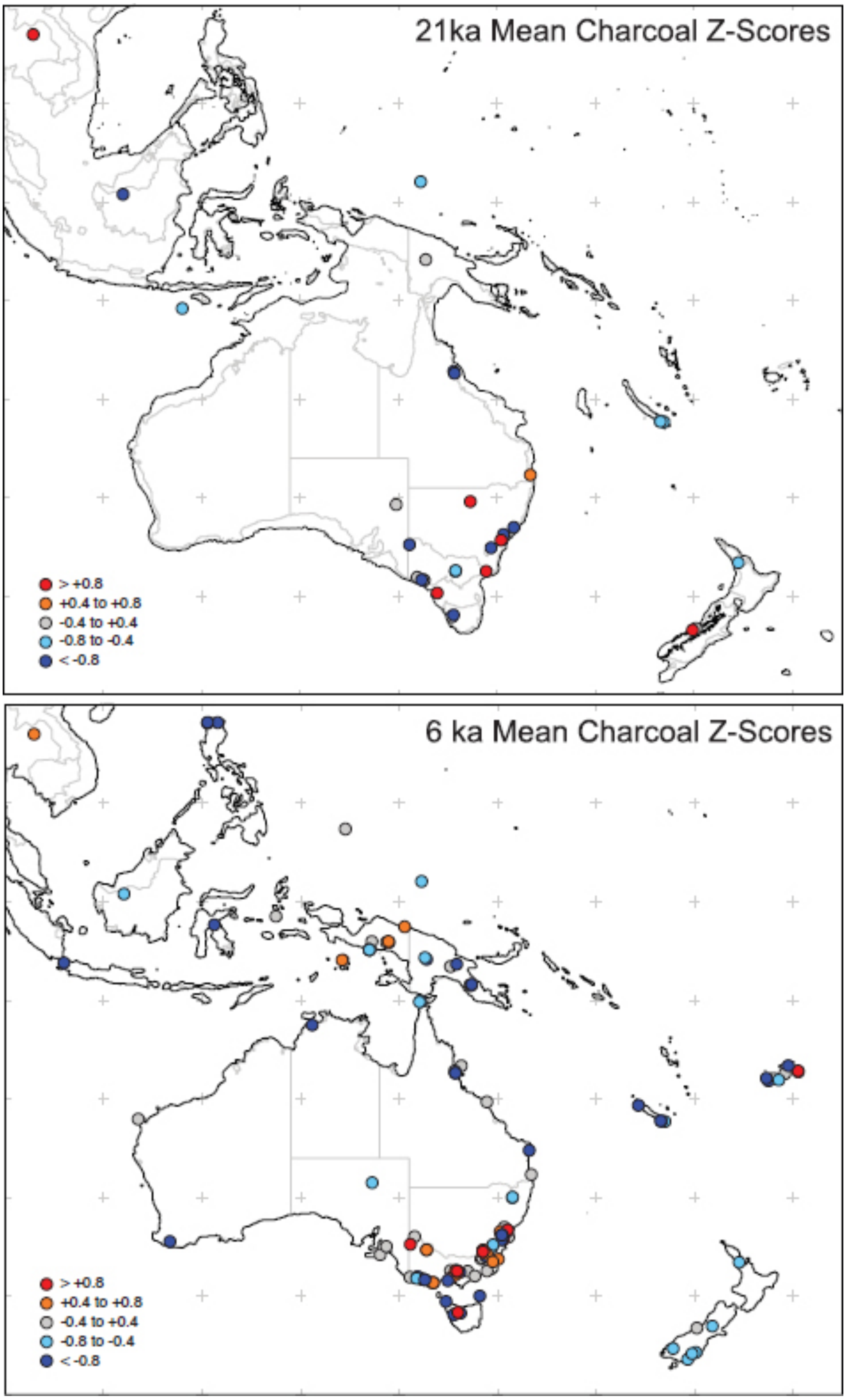

Figure 1.6. Reconstructions of the geography of changes in fire regimes, as expressed by mean z-scores, at 21+0.5 ka and 6+0.5 ka. The expanded continental area at the Last Glacial Maximum (LGM) reflects the change in global sea-level caused by the expanded ice sheets. The LGM (21 ka) and mid–Holocene (6 ka) time periods are targets for model evaluation exercises within the Palaeoclimate Modelling Intercomparison Project (PMIP) and are recommended simulations for the 5th Assessment Report (AR5) of the Intergovernmental Panel on Climate Change (IPCC).

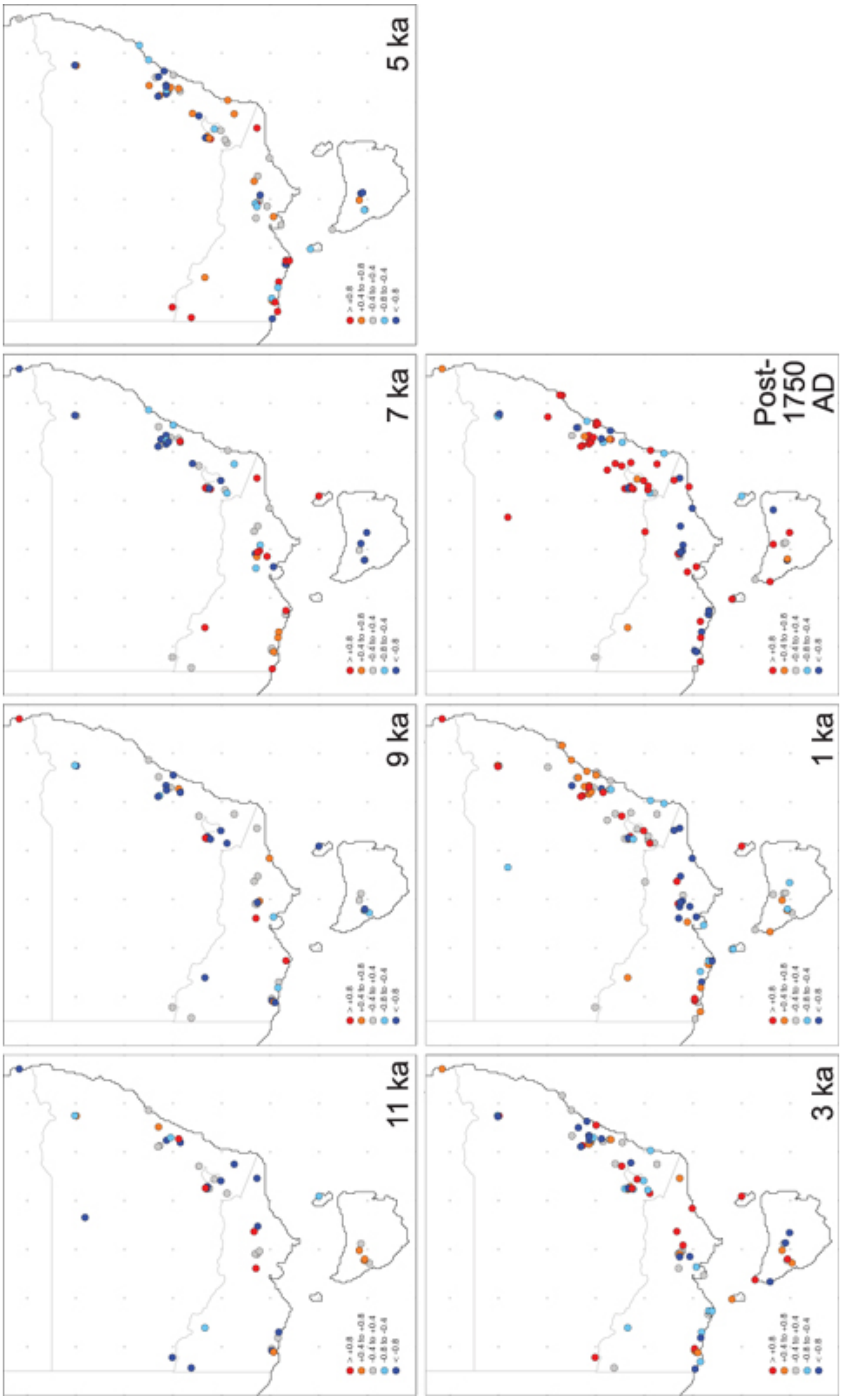

Figure 1.7. Reconstructions of the geography of changes in fire regimes in intervals between 11 to 1 ka compared with the post-1750 AD interval.

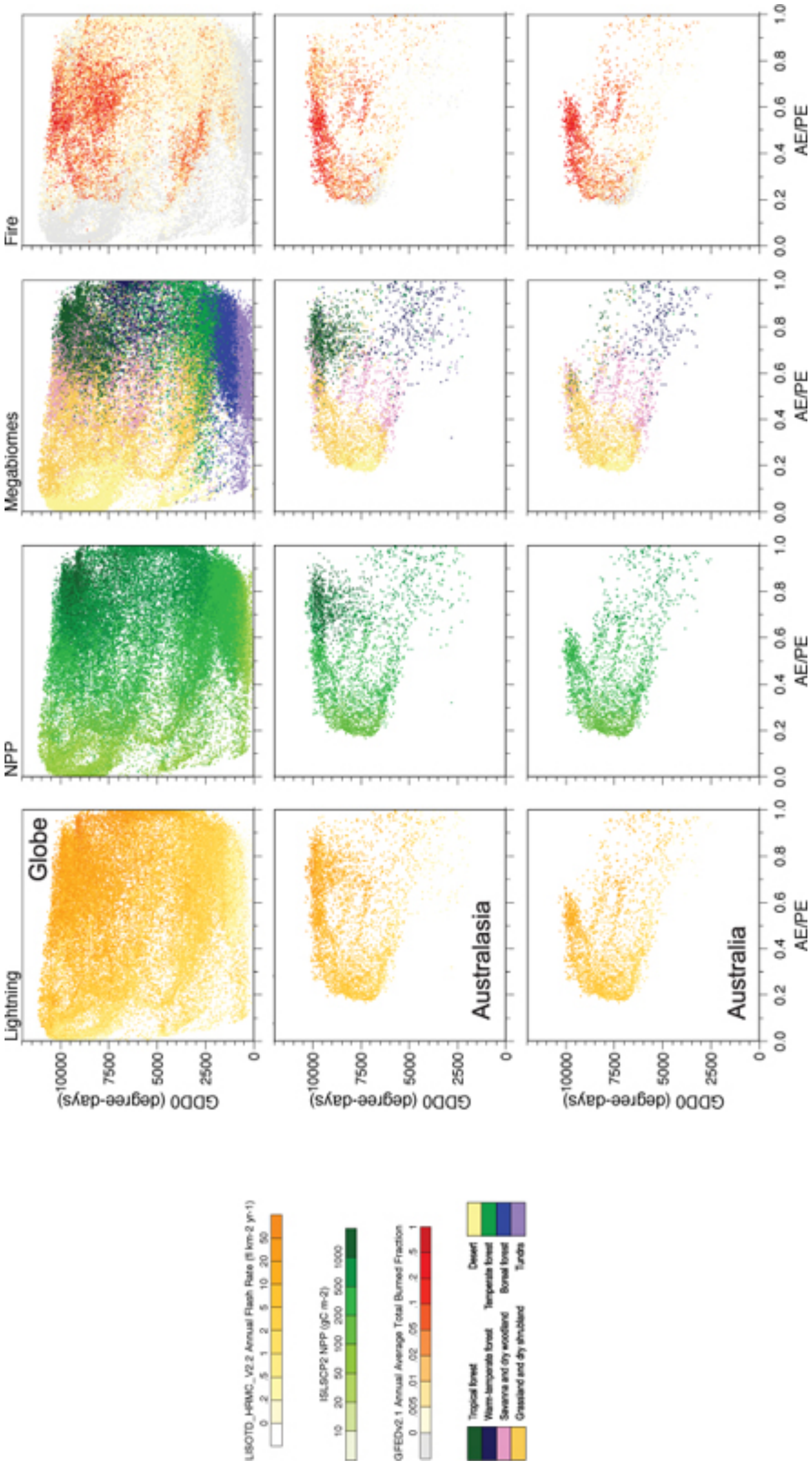

Figure 1.8. Relationships between climate, lightning, vegetation and fire at a global scale (top panels), within Australasia (middle panels) and within Australia (bottom panels) based on satellite burnt–fraction data (GFED v2.1) (van der Werf *et al*. 2006). Climate is here described by variables related to the controls on plant growth: temperature and moisture. Values of actual to equilibrium evapotranspiration (AE/PE) and growing season warmth as expressed by growing degree days above a 0°C baseline (GDD0) are used herein, calculated using the CRU CL 2.0 dataset (New *et al*. 2002). Lightning data are derived from V2.2 the LIS/OTD HRMC dataset (http://gcmd.nasa.gov/records/GCMD_lohrmc.html). Model-based estimates of net primary productivity (NPP) are from Cramer *et al*. (1999). The distribution of major vegetation types is based on a modern day simulation with the BIOME4 model (Kaplan *et al*. 2003). Here, for simplicity, the biomes are grouped into major vegetation types (megabiomes) following Harrison and Prentice (2003).

Although climate changes operate at a regional scale, this analysis helps to explain how similar climate changes can produce different effects on fire regimes at a local scale. For example, increasing precipitation in fuel-limited regions, such as dry grasslands and xerophytic shrublands, will lead to an increase in vegetation productivity and hence an increase in fire. Similar increases in precipitation in regions with more productive vegetation can lead to less fire because fuel becomes wetter. Thus, it is not surprising that during the Holocene, for example, the palaeo-fire record shows more heterogeneity than the underlying climate drivers. Similarly, the tendency for greater spatial coherency in fire regimes changes during intervals typified by very large changes in climate, such as the LGM and the ACR, arises because such large climate changes lead to large shifts in vegetation type and productivity.

Mooney *et al*. (2011) have proposed that the millennial-scale variability in fire regimes across Australasia during the glacial period is associated with the large and rapid climate changes associated with the D-O cycles. The magnitude of the D-O warming is of the order of 10–15°C in Greenland and marine records from the southern extra-tropics indicate significant temperature shifts (e.g. Pahnke and Zahn 2005; Kaiser and Lamy 2010). It is therefore not surprising that changes of this magnitude could produce regionally coherent changes in fire regimes, including in Australasia.

The Australasian palaeo-fire record, with low levels of burning during globally cold intervals and increased burning during warmer intervals, and particularly during intervals of rapid warming associated with D-O events, demonstrates the importance of temperature changes in driving shifts in biomass burning. This insight is important, because it implies that biomass burning will increase with global warming. Analyses of short-term changes in fire regimes (e.g. Westerling *et al*. 2006; van der Werf *et al*. 2006) have tended to emphasise the importance of interannual variability in precipitation as a control on biomass burning. This is not inconsistent with our interpretation of the palaeo-fire record given that, at a large scale, changes in temperature affect the vigour of the hydrological cycle (Hartmann and Larson 2002). Given the much larger uncertainties associated with the prediction of future changes in precipitation, it is vital that the influence of temperature changes on biomass burning is not obscured.

Modern observations of fire in Australia show distinct spatial differences in both the amount and timing of biomass burning (Figure 1.9). Notably, these observations show patterns considerably different from the typical maps of fire seasons (e. g. Luke and McArthur 1978). Northern Australia is characterised by more fire overall and the occurrence of fire in winter; western and south-eastern Australia are characterised by less fire overall and the fire is seasonally concentrated in late spring and summer. These patterns are related to atmospheric circulation controls on climate, with fire occurring after the end of the monsoon rains in northern Australia and at the end of the winter rainy season further south (in both cases, when productivity is at a maximum and environmental aridity increasing). The palaeo-fire record shows that changes in the tropics and in extra-tropical regions of Australia are not synchronous, because changes in the strength of the monsoon and westerly circulations are not necessarily coupled over long timescales (Rojas and Moreno 2010; Mooney *et al*. 2011). The transition between summer-dominated and winter-dominated fire regimes today, however, parallels the regional opposition in vegetation and fire changes shown during the Holocene (Figure 1.7). However, and as shown by the charcoal records, the boundaries between different types of fire regime are not fixed through time.

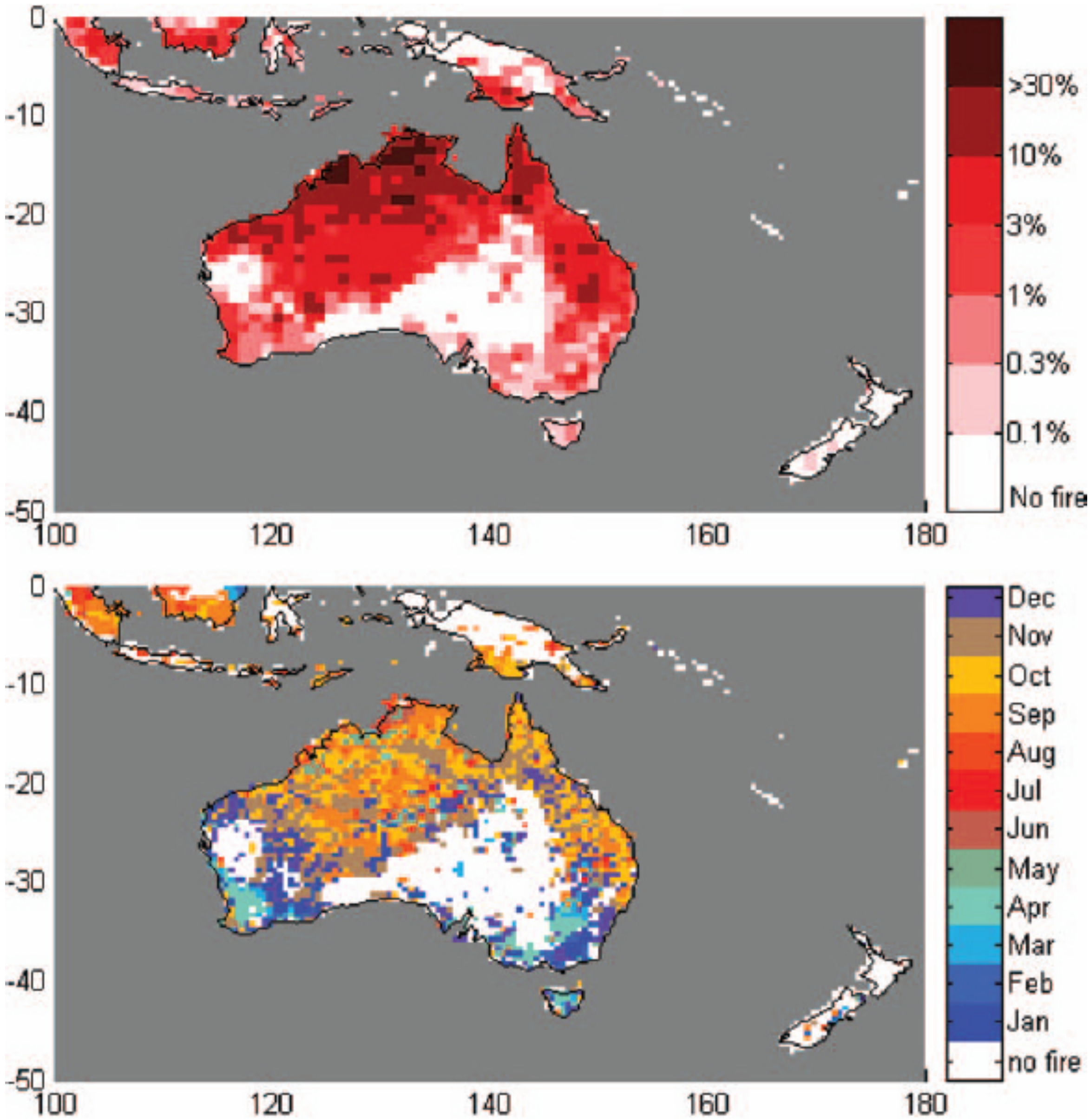

Figure 1.9. Spatial patterns in fire regimes across Australasia (20°N-50°S and 100°E to 177°W) under current climate conditions. (a) Annual burnt fraction, and (b) observed month of maximum burnt area for regions where the fractional burnt area is greater than >0.1%, both averaged over 1997–2006, from the satellite-derived GFED v2.1 product (van der Werf *et al*. 2006).

Fire and people

A major focus of palaeo-fire research in Australasia has been the influence of human activity on the dynamic between fire and vegetation (see e.g. Turney *et al*. 2001; Kershaw *et al*. 2002; Black and Mooney 2006; Lynch *et al*. 2007). Peaks in charcoal occurring at about the time of Aboriginal settlement of Australia (50±10 ka: Bird *et al*. 2004) have been interpreted as evidence of a significant impact of human activities on fire regimes and, through this, on the vegetation at individual sites (e.g. Singh *et al*. 1981; Kershaw 1986; Kershaw *et al*. 2007). The use of 'firestick farming' by Aboriginal people has been associated with the progressive modification of the Australian vegetation (e.g. Flannery 1994), and even the aridification of climate after circa 50 000 years ago (Miller *et al*. 2005). Furthermore, historical narratives have often emphasised a contrast between fire promotion in the pre-European and fire suppression in the post-European period.

The composite record of fire during the last glacial is characterised by marked variability in biomass burning on millennial timescales (Figure 1.3). There is an increase in biomass burning around 50 ka, corresponding to D-O 13 (49.28 ka: Wolff *et al*. 2010), but similar increases appear to be associated with D-O events both before and after this time. The composite charcoal data provides no support for a fundamental shift in the fire regime around 50±10 ka. Indeed, analysis of a number of individual records (e.g. Stevenson and Hope 2005) have already shown that changes in vegetation at around this time are independent of either changes in fire or of human activity.

Mooney *et al*. (2011) have compared the charcoal record of the last 40 kyr with an index of human activity based on archaeological site density (Smith *et al*. 2008): they found no correlations between intervals of increased human activity and intervals of high biomass burning. This includes no apparent association with purported changes in Aboriginal socio-economic relationships ('intensification') after the mid-to-late Holocene (Lourandos 1980) and biomass burning. Thus, the palaeo-fire record provides little support for large-scale impacts of human activity on Australasian fire regimes. This is consistent with the analysis of modern observations, which suggests that ignition is not the limiting factor on fire.

Palaeo-records of fire: conclusions and implications

We began this review by indicating that the palaeoenvironmental record of fire could be used to answer three types of question: What happened when? Are modern fire regimes unusual? And what happens during warmer climates? Comparison of the composite palaeo-fire records and reconstructions of changes in fire regimes on historical and observational timescales (Figure 1.10) show that a high degree of variability is characteristic of Australasian fire regimes on all timescales from multi-millennial to annual. Despite this variability, the palaeo-fire records show consistent responses to climate change, with cold intervals characterised by less fire and warm intervals by more fire. We can also conclude that the larger the underlying climate change, the more likely that the response in fire will be registered at a continental, rather than local, scale. The records also show that the response of fire to climate change is rapid. There is, for example, no discernable lag (with the dating resolution of the records, circa 50–100 years) between rapid climate changes during the last glacial period and the response of fire to these changes. These palaeoenvironmental records indicates that, all other things being equal, predicted increases in temperature during the 21st century across Australia (mean annual increases of >2°C with medium emission scenarios by 2070: CSIRO and BoM 2007; Meehl *et al*. 2007) will lead to a rapid increase in biomass burning.

The palaeo-fire record suggests that the recent past has been unusual: firstly because of the very large increase in biomass burning during the 19th century but also because of the reduction in biomass burning during the 20th century. A similar pattern was identified, though on the basis of a smaller number of records, by Kershaw *et al*. (2002). This perspective is radically different from historical narratives (e.g. Pyne 1991) that have emphasised deliberate fire suppression as characteristic of the European colonisation of Australia. Mooney *et al*. (2011) have shown that the increase in biomass burning after about 1800 AD is registered in tropical regions of Australasia, as well as on the mainland, and have suggested that post-industrial changes in climate and atmospheric composition may have been more important than human activities in generating increased fire.

The 20th century decrease in biomass burning, which is registered in many parts of the world (Marlon *et al*. 2008; Wang *et al*. 2010), is obviously not consistent with a climate driver. Marlon *et al*. (2008) suggested that this decrease was an unforeseen consequence of landscape fragmentation resulting from agricultural expansion. Although the downturn in biomass burning is a composite signal, and therefore not necessarily seen in all regions, the idea that landscape fragmentation leads to reduced fire raises fundamental issues for fire management in the future because it implies that any decrease in landscape fragmentation (e.g. through reforestation for carbon sequestration: Williams *et al*. 2004) could further exacerbate climate-induced increases in fire.

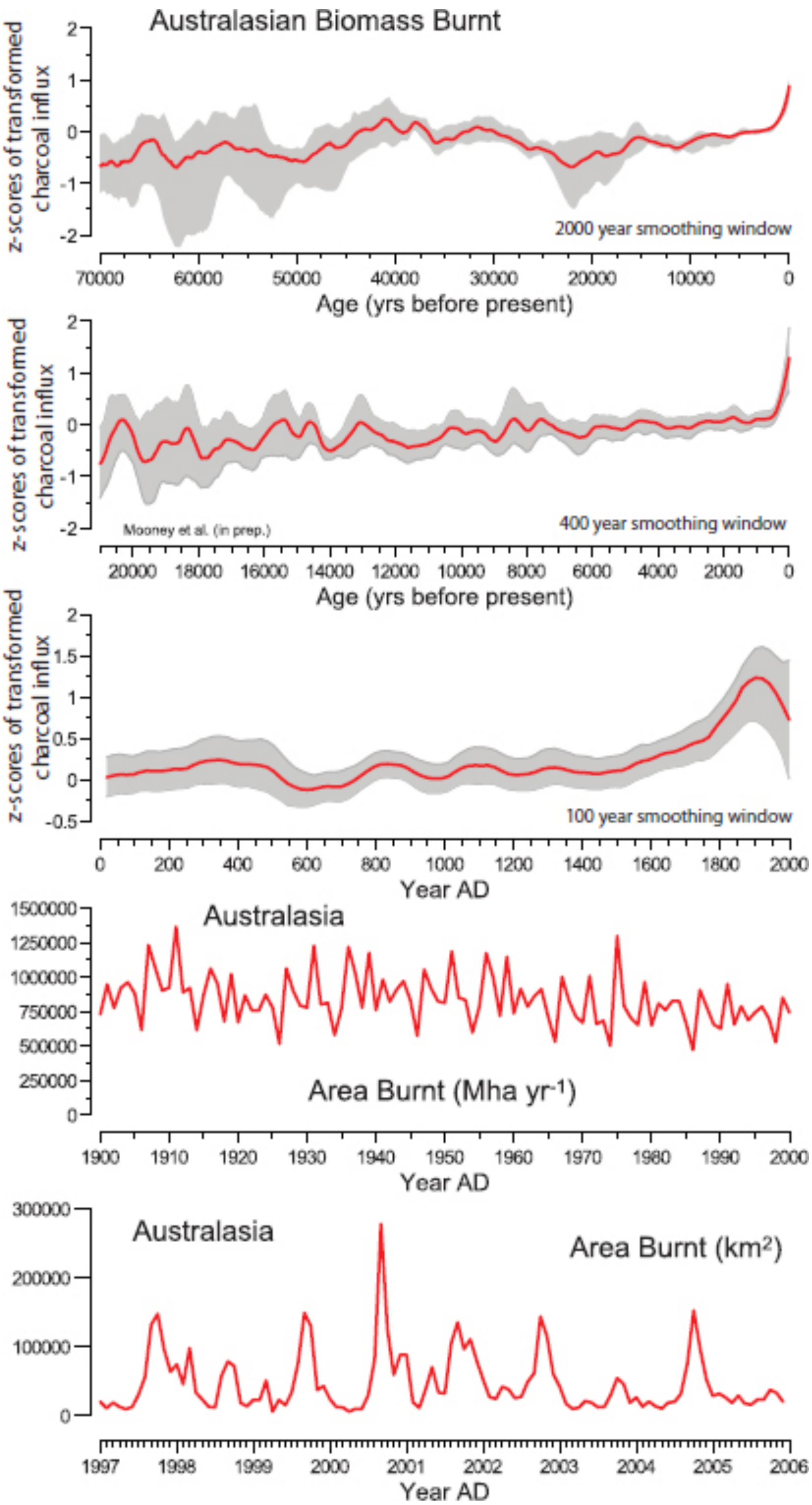

Figure 1.10. Variability in fire for Australasia (20°N–50°S and 100°E to 177°W) over different timescales: estimates of biomass burnt, based on a synthesis of 224 sedimentary charcoal records from Australasia (Mooney *et al*. 2011), (panels from top to bottom) during the last glacial (70 000–0 years before present), since the Last Glacial Maximum (22 000–0 years before present), and over the past two millennia (0–2000 years AD). The lower two panels are recent estimates of fire in Australasia: the first is a modelled estimate of the total area burnt (Mha) in Australasia by Mouillot and Field (2005) and the bottom panel depicts the area burnt from 1997 to 2006 derived from satellite-based remote sensing (GFED v2.1, van der Werf *et al*. 2006).

Through extending the length of the observational record, and particularly into times when climate and environmental conditions have been radically different from today, the palaeo-fire record offers new perspectives on environmental management. The palaeo-fire record from Australasia, though far from complete in terms of spatial and temporal coverage, nevertheless is a rich and as yet barely exploited source of information about the responses to climate and environmental change.

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