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Late Quaternary fire regimes of Australasia

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ABSTRACT

We have compiled 223 sedimentary charcoal records from Australasia in order to examine the temporal and spatial variability of fire regimes during the Late Quaternary. While some of these records cover more than a full glacial cycle, here we focus on the last 70,000 years when the number of individual records in the compilation allows more robust conclusions. On orbital time scales, fire in Australasia predominantly reflects climate, with colder periods characterized by less and warmer intervals by more biomass burning. The composite record for the region also shows considerable millennial-scale variability during the last glacial interval (73.5–14.7 ka). Within the limits of the dating uncertainties of individual records, the variability shown by the composite charcoal record is more similar to the form, number and timing of Dansgaard–Oeschger cycles as observed in Greenland ice cores than to the variability expressed in the Antarctic ice-core record. The composite charcoal record suggests increased biomass burning in the Australasian region during Greenland Interstadials and reduced burning during Greenland Stadials. Millennial-scale variability is characteristic of the composite record of the sub-tropical high pressure belt during the past 21 ka, but the tropics show a somewhat simpler pattern of variability with major peaks in biomass burning around 15 ka and 8 ka. There is no distinct change in fire regime corresponding to the arrival of humans in Australia at 50 ± 10 ka and no correlation between archaeological evidence of increased human activity during the past 40 ka and the history of biomass burning. However, changes in biomass burning in the last 200 years may have been exacerbated or influenced by humans.

1. Introduction

Australia includes some of the most fire-prone landscapes on Earth (Williams et al., 2001; Bradstock et al., 2002; Bond and Keeley, 2005; Russell-Smith et al., 2007). Fire has major impacts on the native flora and fauna, on landscape stability and on biogeochemical cycling. Characteristics of the land cover and the release of gases and particulates (CO_2, CO, CH_4, N_2O, BVOCs, black carbon) during bushfires affects air quality, atmospheric composition and hence radiation budgets, and thus changes in fire regimes through time could have important feedbacks to climate (Ramanathan and Carmichael, 2008; Bowman et al., 2009; Arneth et al., 2010).

Australian vegetation has developed a variety of responses and morphological and reproductive adaptations to fire, including the

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widespread use of resprouting (Purdie, 1977; Gill et al., 1981; Enright et al., 1998), suggesting that fire has played an important role over evolutionary timescales. Many species require regular fire in order to persist, and this is particularly evident in humid but intermittently drought-prone environments where eucalypts dominate the vegetation. Other taxa, including species typical of the more consistently humid east and northeastern coast, are fire sensitive (Gill et al., 1981; Bradstock et al., 2002). In the tropical communities of Australasia, fire often depends on drought and fires in the recent past have been linked to human activity and El Niño events (van der Werf et al., 2008a; Lynch et al., 2007). Fire is thus a fundamental issue in many Australasian landscapes and influences community composition, the location of boundaries between communities and vegetation dynamics through time.

There are persistent questions about the role of humans in the long-term history of fire and vegetation in the Australasian region. It has been argued that the frequent use of fire by Aboriginal people, to manipulate the availability of resources (Jones, 1969; Nicholson, 1981), resulted in vegetation change and other environmental impacts in the late Pleistocene (e.g. Singh et al., 1981; Flannery, 1994; Miller et al., 2005). Ideas about pre-historic fire continue to influence debates concerning natural resource management, with suggestions that Aboriginal-like fire management (i.e. frequent and low intensity fires) could prevent conflagrations in the modern setting (e.g. Select Committee on the Recent Australian Bushfires, 2003).

One of the earliest examinations of long-term changes in fire regimes and their impact on the development of Australian vegetation was provided by Gill et al. (1981). More recently, Kershaw et al. (2002) have summarised the long-term history of fire in Australia, although their discussion of the last 10,000 years focused solely on southeastern Australia and they used a qualitative assessment of *ca*. 60 sites. Nevertheless, Kershaw et al. (2002) indicated that there were differences in the timing of peak Holocene fires associated with different biomes across this region. The interval of least fire across all biomes was during the mid-Holocene (7000-5000 yr BP) and the maximum registration of fire occurred during the early European period. Lynch et al. (2007), which is the most recent review of Australian palaeofire regimes but only covers a small number of iconic records, also identifies the mid-Holocene as a time of low fire activity and argues that higher levels of biomass burning are associated with the onset or intensification of the El Niño-Southern Oscillation (ENSO) after *ca*. 4 ka. Lynch et al. (2007) conclude that the longer-term record of fire in Australia shows a gradual increase coincident with the purported long-term aridification of the continent (Hesse et al., 2004).

There have been several site specific or regional studies examining the interactions between vegetation and fire during the late Quaternary in Australia (e.g. Black et al., 2007). However, it is only recently, and largely through the efforts of the Global Palaeofire Working Group (GPWG: Power et al., 2008; Power et al., 2010), that sufficient data have become available at a continental scale to make it possible to apply robust statistical techniques to the analysis of past fire regimes. Power et al. (2008), based on an analysis of 355 charcoal records, showed that fire regimes globally reflected long-term climate changes. This compilation included only 48 records from "Australia", (defined as mainland Australia, New Zealand and Pacific Islands west of 180oE, but not New Guinea which was included in southeast Asia). This preliminary analysis revealed distinctive patterns in fire regimes over the last 21 ka, with relatively little change during the late glacial and through the transition to the Holocene, and charcoal peaks around 15-16 ka, 11-10 ka and 4.5–2.5 ka (Power et al., 2008). Here, we present a more extensive synthesis of the charcoal data based on 223 sites from Australasia and analyse these data to determine how fire regimes have changed over centennial to multi-millennial timescales.

1.1. Methods: source and treatment of charcoal records

Our focus in this paper is on the Australasian region, which we define here to include tropical southeastern Asia, New Guinea, New Zealand and the islands of the western Pacific. Southeastern Asia and New Guinea were included to set the records from tropical Australia in a broader context. The inclusion of sites from the western Pacific (including New Zealand) is partly motivated by the need to address the potential role of changes in the ENSO on Australasian climates (McGlone et al., 1992; Shulmeister and Lees, 1995; Lynch et al., 2007) and partly because this region has a different settlement history from Australia (see e.g. Stevenson and Hope, 2005). We extracted 196 sedimentary charcoal records from this broadly defined Australasian region (20° N - 50° S and 100° E to 177° W) from Version 2 of the Global Charcoal Database (GCD-V2; Daniau et al., in preparation,) compiled by the Global Palaeofire Working Group (GPWG: http://gpwg.org/). These data were supplemented by an additional 27 sites, chosen to increase the number of long records and to improve the spatial coverage (GCD-V2.5).

This GPWG database contains sedimentary charcoal records from both marine and terrestrial sites. It includes descriptive data (metadata) about both the sites and the methods used, and detailed information on site chronology (including information on the number of radiometric dates and, for records extending back beyond the limits of radiocarbon dating, details of correlative tie-points used to erect the chronology). All radiocarbon dates have been calibrated and the age model for each site is expressed in calendar years. Where multiple records (e.g. macro and micro-charcoal records) are available at the same site, all are included in the database.

Since charcoal records are obtained using many different techniques and expressed using a large range of metrics, the data were standardized to facilitate comparisons between sites and through time (see Power et al., 2008 for a full description). The protocol involved three steps. First, non-influx data (e.g. concentration expressed as particles/cm^3; charcoal-to-pollen ratios) were transformed to influx values (i.e. particles/cm^2/yr), or quantities proportional to influx, by dividing the values by sample deposition times. Second, a Box-Cox transformation was used to homogenize 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 experimentation has demonstrated that the choice of the base period does not affect the results significantly (Marlon et al., 2008; Power et al., 2010) and standardization does not alter the overall pattern of variability or "signal" in a record. For mapping purposes, the z-scores were divided into five approximately equal groups: z-score $> +0.8$ (strong positive anomalies compared to the long-term average over the base period), z-score between $+0.4$ and $+0.8$ (positive anomalies), z-score between $+0.4$ and -0.4 (weak positive or negative anomalies), z-score between -0.4 and -0.8 (negative anomalies), and z-score < -0.8 (strong negative anomalies).

In order to summarise the broadest-scale history of changing fire regimes through time, we constructed composite charcoal records for Australasia as a whole and for various sub-regions. Composite curves were then obtained by fitting a locally weighted regression (or "lowess" curve) to the pooled transformed and rescaled data (see e.g. Marlon et al., 2008), using a fixed window width and a tricube weight function with one "robustness iteration". We used window half-widths of 100, 200, 400, or 2000 years, to emphasize different scales of temporal variability from centennial variability in the past 2000 years through to multi-millennial variability in the longer records. The windows were selected to avoid both over-smoothing, as would result from selecting a large window width, and under-smoothing, leading to a composite curve that was susceptible to the influence of individual data points. These choices do not affect the results and conclusions.

Confidence intervals for each composite curve were generated by a bootstrap re-sampling (with replacement) of individual sites over 1000 replications. This approach differs from the usual bootstrap method in which individual observations are sampled with replacement, and emphasizes the uncertainty in regression curves that arises from the inclusion or exclusion of the whole record from individual sites. Bootstrap 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. Our approach is therefore conservative because it permits the identification of "signals" that may arise solely from the influence of a particular 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. Minor fluctuations in the composite curve during times when the bootstrap confidence intervals are wide are most probably meaningless. Most of the time, the composite curve lies in the middle of the bootstrap confidence intervals; when one confidence limit deviates from the composite curve more than the other confidence limit, this too indicates that some individual records depart markedly from the typical pattern of the time. Again, the interpretation of minor fluctuations in the composite curve under these circumstances is not likely to be meaningful. We also provide a measure of the number of observations contributing to the composite curve (at each of the target points that define the curve) as an additional measure of the confidence to be placed in the reconstructions. Since this takes into account the number of points at each site that fall within the window width for a particular target point, these curves are not a simple measure of the number of sites.

2. Results

2.1. Characteristics of the data set

The geographic coverage of sites in the dataset is uneven (Fig. 1), with few sites from central and western Australia, but with eastern and especially southeastern, Australia well represented. While there are some sites from tropical mainland Australia, most of the tropical records are from Papua New Guinea, with a small number of additional sites scattered throughout southeastern Asia as well as several tropical Pacific Islands (Fig. 1a). The coverage of records from New Zealand is reasonable, although it does not approach the known number of records for the island. The paucity of data from northern, western and central Australia is partially a reflection of absence of suitable sites for the preservation of sedimentary charcoal records (Pickett et al., 2004) and partially a function of the comparative difficulty of working in these regions. Nevertheless, the data do provide a reasonably comprehensive coverage of the range of climates and vegetation types found in Australasia (Fig. 1a–e).

The temporal coverage of the data set is also uneven (Table 1). There are 18 records that extend back beyond 70 ka (Fig. 1c). These sites are not spatially clustered, but rather occur from just north of the equator to the South Island of New Zealand, and therefore experienced very different climate and fire regimes from one another. There are 19 sites that provide a record from the beginning of Marine Isotope Stage (MIS) 3 (here defined as 59.4 ka, following Sanchez Goñi and Harrison, 2010) and 40 sites that are recording by the end of MIS 3 (here defined as 27.8 ka, following Sanchez Goñi and Harrison, 2010). Over half of these records are sampled at an average resolution of more than 1 sample per ka (Table 1). The number of records and their spatial distribution during MIS 3 makes Australasia one of the best-documented regions of the world for this period (see Daniau et al., 2010). The number of charcoal records increases steadily during MIS 2 (27.8–14.7 ka), such that by the end of MIS 2 there are approximately 70 sites and by 10 ka there are about 110 sites recording fire. Sampling resolution for sites covering the Holocene varies from ca 1 sample per ka to 1 sample per decade: the majority of the short records (i.e. those covering the last 1-2 ka) have been sampled at decadal resolution, while the majority of the late Holocene records have centennial resolution.

The charcoal records from Australasia were obtained using several different techniques (Table 1). The quantification of charcoal fragments on pollen slides dominates the original records (80.4% of the sites where method was explicitly recorded). However, 44 of the records (19.6% of the sites where method was explicitly recorded) are macroscopic charcoal records obtained by wet sieving. Macroscopic, sieved, charcoal records are more likely to provide a record of local fires and provide more continuous temporal sequences than charcoal records obtained through other means. Nevertheless, although differences between the macro- and micro-charcoal records at individual sites have been used to infer the spatial scale of fire events (Power et al., 2010), both microscopic and macroscopic records produce comparable results in terms of broadscale regional histories of fire (Tinner et al., 2006; Conedera et al., 2009).

2.2. Temporal trends in biomass burning

There are relatively few sites with a record prior to MIS 4 and the composite curve for the interval prior to 70 ka is too noisy to interpret. We have therefore chosen to focus our analyses on the interval after 70 ka. The composite record from Australasia (Fig. 2, purple curve) shows a steady increase in biomass burning during the later part of MIS 4 (73.5–59.4 ka). Biomass burning remains generally high through MIS 3, but decreases at the beginning of MIS 2. Biomass burning levels are generally high through the Holocene. This pattern, of higher levels of biomass burning during MIS 3 than MIS 4, the reduction of biomass burning at the beginning of MIS 2, and the return to higher levels of biomass burning during the Holocene, is consistent with the global pattern of less fire during cold stadial or glacial stages, and increased fire during warmer interstadials and interglacials (Power et al., 2008; Daniau et al., 2010). Although there are marked changes in fire activity during MIS 3 (see below), there is no fundamental shift in the composite charcoal record that could be associated with the colonization of Australia by Aboriginal people (50 ± 10 ka: Bird et al., 2004).

Superimposed on these general trends, the composite record shows considerable millennial-scale variability in biomass burning (Fig. 2, black curve). This is most marked in MIS 3, but can also be seen during the later part of MIS 4, with multiple peaks in biomass burning in the 400-year smoothed record between 70 and 20 ka (Fig. 2, black curve). Some of these charcoal peaks appear to occur around the time of the Dansgaard–Oeschger (D–O) warming events as described in Greenland ice-core records (Fig. 3). Other peaks in the charcoal record are not apparently coincident with D–O warming events, but this may be a function of the considerable uncertainty associated with the chronologies of individual charcoal records that are beyond the limit of radiocarbon dating. Several of the records contributing to the composite curve have been tuned to match assumed limits of the transition between the last interglacial and the glacial (MIS 5/4 boundary) and the MIS 3/2 boundary on the basis of changes in pollen stratigraphy, and this procedure could certainly add uncertainties of several thousand years to the assumed age model. In terms of the number and shape of the peaks, and in terms of the timing during the post-30 ka interval (when radiocarbon dating is likely to be more reliable), the

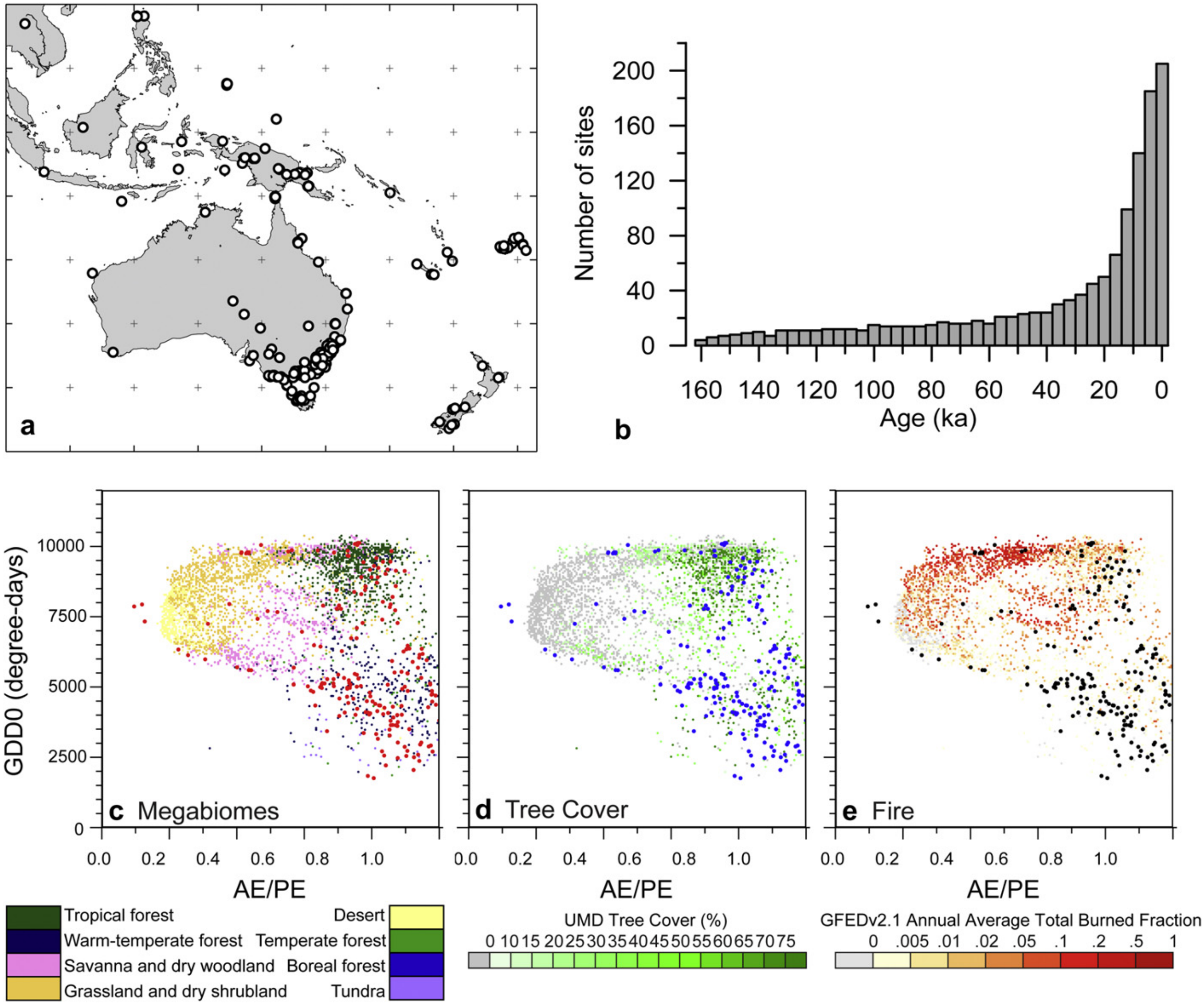

Fig. 1. Distribution of sites in (a) geographic space, (b) through time, (c, d) bioclimate and vegetation space and (e) fire regime space. In (b), a site is counted if it contributes data to a 4000-year wide bin. Bioclimate space is defined here by growing degree days above 0° (GDD0: an index of summer warmth) and the ratio of actual to equilibrium evapotranspiration (AE/PE: an index of effective moisture) calculated using climate data from the CRU historical climatology data set (New et al., 2002). The biome at each site (c) is based on a modern day simulation with the BIOME4 model (Kaplan et al., 2003) and has been grouped into major vegetation types (megabiomes) following the classification in Harrison and Prentice (2003). The tree cover (d) is derived from remotely-sensed vegetation data (De Fries et al., 2000). Fire regimes (e) are represented by burnt area (annual average total burnt fraction) from the GFED v2.1 data set (van der Werf et al., 2006).

form of the composite charcoal curve is more similar to the pattern of millennial-scale variability shown in the NGRIP record than to the pattern shown by the EPICA deuterium record (Fig. 3), which shows a smaller number of lower amplitude “Antarctic Isotope Maxima” over the same interval (EPICA Community Members, 2006). Furthermore, this pattern is consistent with a global analysis of charcoal records that span the interval from 80 ka to 10 ka (Daniau et al., 2010) that show a consistent (across the D–O events) increase in biomass burning accompanying D–O warming events.

There are relatively low levels of biomass burning between ca 24 to 18 ka (i.e. around the time of the Last Glacial Maximum, LGM), but fire increases thereafter. The post-glacial increase of biomass burning (Fig. 4a) is expressed differently in the Intertropical Convergence Zone belt (ITCZ, 20° N to 20° S, 100° E to 177° W) and the sub-tropical high pressure belt (STH, 25–45° S, 100° E to 177° W). In the tropics (Fig. 4b), biomass burning increases gradually after the LGM through to the early Holocene, but with two broad peaks of increased fire centered around 15 ka and 8 ka. There is a sharp decrease in biomass burning culminating at 6.5 ka. There is a rapid recovery after this and the rest of the Holocene appears relatively complacent until the last few hundred years. The records from the STH belt (Fig. 4c) show a different pattern, with no marked trends during the deglacial period, but there is a tendency towards increased biomass burning during the first part of the Holocene culminating around 6 ka and a decline in fire during the late Holocene. The STH record also shows centennial- to millennial-scale variability throughout the last 21 ka. Again, the most pronounced changes in STH fire regimes occur during the last few hundred years.

The composite Australasian record of biomass burning over the past two millennia (Fig. 5a) is remarkably flat except for the pronounced increase in fire in the past 200 years. Marlon et al. (2008) showed that global biomass burning gradually declined over most of the last two millennia in response to long-term

Table 1
Sites with charcoal records from the Australasian region.

Site Name	Latitude (°)	Longitude (°)	Elevation (m)	Site type	Charcoal methods	Record length (age ka)	Resolution (samples/ka)	References
Aguai Ramata	−6.5639	145.211	1950	lacustrine	macro, sieved	15.84	32.88	Haberle, 2007
Aire Crossing	−38.6467	143.4797	180	terrestrial	micro, pollen slide	10.33	3.97	McKenzie and Kershaw, 2004
Ajax Hill	−46.4192	169.2922	680	mire	macro, sieved	14.21	4.01	McGlone, 2009
Ajkwa 1	−4.867	136.968	1.3	coastal	micro, pollen slide	6.16	4.38	Ellison, 2005
Ajkwa 2	−4.867	136.968	1.3	coastal	micro, pollen slide	10.43	5.85	Ellison, 2005
Ajkwa 3	−4.867	136.968	1.3	coastal	micro, pollen slide	2.96	13.83	Ellison, 2005
Ajkwa 4	−4.867	136.968	1.3	coastal	micro, pollen slide	5.55	4.86	Ellison, 2005
Ajkwa 5	−4.867	136.968	1.3	coastal	micro, pollen slide	5.55	7.38	Ellison, 2005
Allom Lake	−25.2333	153.1667	100	lacustrine	macro, sieved	57.00	2.14	Donders et al., 2006
Anggi Lake	−1.403	133.902	1945	lacustrine	micro, pollen slide	26.08	0.35	Hope, 2007a
Anumon Swamp	−20.163	169.828	45	mire	micro, pollen slide	1.87	4.27	Hope, 1996b
Argan Swamp	−10.05	142.06	3	coastal	macro, sieved	5.52	23.74	Rowe, 2006a
Aru	−5.917	134.2	1	terrestrial	micro, pollen slide	6.89	2.61	Hope and Aplin, 2005
Badu 15	−10.06	142.09	20	terrestrial	macro, sieved	4.19	7.63	Rowe, 2006a, b, 2007
Banda Sea Core SHI-9014	−5.76667	126.9667	−3163	marine	micro, pollen slide	173.31	0.54	van der Kaars et al., 2000
Bar20	−10.1	142.12	18	coastal	macro, sieved	2.83	25.42	Rowe, 2006a, 2007
Bega Swamp	−36.5167	149.5	1080	mire	micro, pollen slide	12.25	39.92	Polach and Singh, 1980
Black Swamp	−30.0206	151.4828	1450	mire	micro, pollen slide	10.62	6.21	Dodson et al., 1986
Blue Lake Kosciuzko	−36.141	148.437	1950	lacustrine	macro, sieved	8.18	60.43	Raine, 1974
Blundells Flat	−35.19	148.49	762	mire	macro, sieved	2.69	90.70	Hope et al., 2006b
Bobundara Swamp	−36.25	150.06	75	coastal	micro, pollen slide	8.01	9.99	Hope et al., 2006a
Boggy Swamp	−29.9611	151.4986	1160	mire	micro, pollen slide	12.49	6.73	Dodson et al., 1986
Boigu Gawat Core 1	−10.1	142.14	10	coastal	macro, sieved	4.57	10.94	Rowe, 2006a, 2007
Boigu Gawat Core 2	−10.1	142.14	10	coastal	macro, sieved	13.82	3.62	Rowe, 2007
Bolin Billabong combined core	−37.7667	145.0667	50	fluvial	micro, pollen slide	0.87	46.91	Leahy et al., 2005
Bonatoa	−18.07	178.53	4	mire	micro, pollen slide	5.41	5.18	Hope, 1996a; Hope et al., 1999
Bondi Lake Centre Core	−36.8061	149.9378	22	coastal	micro, pollen slide	1.16	34.40	Dodson et al., 1993
Bondi Lake South Core	−36.8061	149.9378	22	coastal	micro, pollen slide	2.12	13.19	Dodson et al., 1993
Boulder Flat	−37.45	148.92	25	terrestrial	micro, pollen slide	27.22	1.25	Kenyon, 1989
Breadalbane	−34.8	149.5167	694	lacustrine	micro, pollen slide	6.50	2.62	Dodson, 1986
Bridgewater Lake Core B	−38.3	141.3833	20	lacustrine	micro, pollen slide	4.86	7.62	Head, 1988
Brooks Ridge Fen	−36.1583	148.5944	1450	mire	micro, pollen slide	1.65	8.48	Mooney et al., 1997
Burraga Swamp	−32.0281	151.4334	1462	mire	micro, pollen slide	1.87	17.16	Dodson et al., 1994c
Burralow Creek Swamp	−33.5333	150.6333	320	mire	micro, pollen slide	1.05	12.34	Chalson, 1991
Butchers Swamp	−29.9722	151.4522	1230	mire	micro, pollen slide	14.21	2.25	Dodson et al., 1986
Buxton	−37.4483	145.6919	235	mire	micro, pollen slide	7.20	5.83	McKenzie, 2002
Byenup Lagoon Site 1	−34.4667	116.7333	175	lacustrine	micro, pollen slide	4.90	3.67	Dodson and Lu., 2000
Byenup Lagoon Site 2	−34.4667	116.7167	175	lacustrine	micro, pollen slide	13.20	1.36	Dodson and Lu., 2000
Caledonia Fen	−37.3333	146.7333	1280	mire	micro, pollen slide	140.72	1.21	Kershaw et al., 2007
Cameron's Lagoon	−41.967	146.683	1045	mire	micro, pollen slide	7.83	2.68	Thomas and Hope, 1994
Chapple Vale Swamp	−38.6261	143.323	30	mire	micro, pollen slide	7.16	9.64	McKenzie and Kershaw, 1997
Clarks Junction	−45.7181	170.1139	560	lacustrine	macro, sieved	16.80	2.20	McGlone, 2001
Club Lake	−36.4	148.3166	1980	lacustrine	micro, pollen slide	0.84	64.06	Dodson et al., 1994a
Cobrico Swamp	−38.3	143.0333	140	mire	micro, pollen slide	0.64	15.71	Dodson et al., 1994b
Core Fr10/95-GC-17	−22.0458	113.5018	−1093	marine	micro, pollen slide	101.64	0.96	Van der Kaars and De Deckker, 2002
Cotter Source Bog margin	−35.9667	148.8167	1755	mire	micro, pollen slide	10.68	1.68	Jones, 1990
Cotter Source Bog center	−35.9667	148.8167	1755	mire	micro, pollen slide	6.93	2.31	Jones, 1990; Hope and Clark, 2008
Cuddie Springs	−30.3795	147.3117	127	lacustrine	micro, pollen slide	57.45	0.54	Field et al., 2002
Dalhousie Springs	−26.4167	135.52	150	spring	micro, pollen slide	1.96	9.71	Boyd, 1990
Den Plain 3	−41.3	146.2	350	lacustrine	micro, pollen slide	2.38	5.04	Moss et al., 2007
Doge Doge	−17.9066	177.2776	8	fluvial	micro, pollen slide	8.58	1.86	Hope et al., 2009
Dove Lake	−41.66	145.96	934	lacustrine	micro, pollen slide	13.50	5.70	Dyson, 1995
Dublin Bog	−41.727	146.233	710	mire	macro, sieved	15.90	5.79	Colhoun et al., 1991
Egg Lagoon	−39.65	143.95	20	mire	micro, pollen slide	149.38	0.74	D'Costa et al., 1993
Evoran Pond	−18.7613	169.0118	194	lacustrine	micro, pollen slide	2.56	10.92	G. Hope, unpublished data

Eweburn Bog	−45.32	167.8089	320	mire	micro, pollen slide	10.61	5.75	Ogden et al., 1998
Fred South Swamp	−38.1368	141.7835	27	mire	micro, pollen slide	18.03	1.16	Builth et al., 2008
Gallahers Swamp	−34.29	150.43	535	mire	micro, pollen slide	22.21	0.90	Hope, 2005a
Galway Tarn	−43.4083	169.8733	130	lacustrine	macro, sieved	50.86	2.36	M. Vandergoes, unpublished data
Ginini Flats	−35.31	148.46	1590	mire	micro, pollen slide	3.45	4.35	Hope et al., 2005
Glendhu Bog	−45.8381	169.7264	600	mire	macro, sieved	12.00	5.17	McGlone and Wilmshurst, 1999
Goochs Swamp	−33.45	150.26	960	mire	macro, sieved	14.06	5.12	Black and Mooney, 2006
Greens Bush	−38.4333	144.9333	160	coastal	micro, pollen slide	5.77	4.85	Jenkins, 1992
Grey Pole Swamp Broughton Island	−32.6149	152.3173	9	mire	macro, sieved	5.38	46.44	Horn, 2005
Griffith Swamp	−33.2833	151	20	mire	macro, sieved	6.43	42.33	Mooney et al., 2007
Haeapugua	−5.8333	142.7833	1650	marsh	micro, pollen slide	25.50	3.22	Haberle and Ledru, 2001
Henty Bridge	−41.9922	145.4736	115	spring	micro, pollen slide	29.42	0.61	Colhoun, 1985
Hogan's Billabong	−36.025	146.7153	140	lacustrine	micro, pollen slide	1.92	18.23	Reid et al., 2007
Hogayaku	−3.983	137.383	3580	lacustrine	micro, pollen slide	5.37	10.43	Hope, 2007b
Hopwoods Lagoon	−33.217	150.9933	38	lacustrine	macro, sieved	0.66	60.94	Smeulders, 1999
Howes Waterhole Swamp	−33.0167	150.6667	280	mire	macro, sieved	2.37	75.95	Mason, 2004
Ijomba	−4.0333	137.2167	3630	mire	micro, pollen slide	17.91	2.46	Haberle and Ledru, 2001
Ingar Swamp	−33.7699	150.4563	584	mire	micro, pollen slide	8.01	2.62	Chalson, 1991
Jibbon Lagoon	−34.0833	151.15	65	mire	macro, sieved	1.58	48.75	Mooney et al., 2001
Kaipo	−38.4	177.1	555	mire	micro, pollen slide	17.51	7.77	Newnham and Lowe, 2000; Hajdas et al., 2006
Katoomba Swamp	−33.7173	150.3189	950	mire	micro, pollen slide	7.16	3.63	Chalson, 1991
Kelela Swamp	−4.0207	138.9125	1650	mire	micro, pollen slide	11.67	3.17	Haberle et al., 1991
Kettlehole Bog	−43.0546	171.7862	600	mire	micro, pollen slide	17.45	5.16	McGlone et al., 2004
Killalea Lagoon	−34.6003	150.8678	22	coastal	micro, pollen slide	1.93	12.44	Dodson et al., 1993
Kings Tableland Swamp	−33.7333	150.4833	780	mire	micro, pollen slide	18.39	1.36	Chalson, 1991
Kings Tableland Swamp (short core)	−33.7333	150.4833	780	mire	macro, sieved	1.06	94.54	Chalson, 1991; Black, 2001
Kings Waterhole	−33.0167	150.6667	280	mire	macro, sieved	6.30	17.15	Black, 2001
Kohuora	−36.57	174.52	73	lacustrine	micro, pollen slide	32.43	4.87	Newnham et al., 2007a
Kosipe A	−8.4667	147.2	1960	marsh	micro, pollen slide	33.36	0.87	Hope, 2009
Kosipe C	−8.4667	147.2	1960	marsh	micro, pollen slide	55.00	0.80	Hope, 2009
Koumac	−20.65	164.283	2	coastal	micro, pollen slide	6.38	0.94	Hope et al., 1999
Kurnell Fen	−34.01	151.1	15	coastal	micro, pollen slide	9.19	3.59	Martin, 1994
Kurnell Swamp	−34.0333	151.2167	2	mire	micro, pollen slide	2.00	5.50	Martin, 1994
Lac Suprin	−22.18	166.59	230	lacustrine	micro, pollen slide	33.17	1.42	Hope and Pask, 1998
Lake Baraba Thirlmere Lakes	−34.2342	150.5397	305	lacustrine	macro, sieved	54.60	1.70	Black et al., 2006
Lake Condah	−38.0667	141.8333	60	lacustrine	micro, pollen slide	11.18	2.95	Builth et al., 2008
Lake Coomboo	−25.2195	153.1959	90	lacustrine	micro, pollen slide	202.31	0.31	Longmore, 1997
Lake Couridjah Thirlmere Lakes	−34.2322	150.542	310	lacustrine	macro, sieved	15.42	1.95	Clark, 1997
Lake Curlip	−37.8333	148.565	2	lacustrine	micro, pollen slide	0.32	99.96	Ladd, 1978
Lake Euramoo	−17.1599	145.6286	718	lacustrine	macro, sieved	23.48	32.71	Haberle, 2005
Lake Eyre (Core LE 82-2)	−28.5	137.25	−15	lacustrine	micro, pollen slide	39.18	1.51	Gillespie et al., 1991; Luly, 2001
Lake Flannigan King Island	−39.6	143.95	40	mire	micro, pollen slide	4.05	4.69	D'Costa, 1997
Lake Frome	−30.68	139.78	40	lacustrine	micro, pollen slide	22.68	5.07	Singh and Luly, 1991; Luly and Jacobsen, 2000; Luly, 2001
Lake George	−35.0656	149.4181	673	lacustrine	micro, pollen slide	116.71	0.58	Singh et al., 1981; Singh and Geissler, 1985
Lake Habbema	−4.1167	138.7	3120	lacustrine	micro, pollen slide	11.59	3.80	Haberle et al., 2001
Lake Hordern	−38.7833	143.4667	3	lacustrine	micro, pollen slide	4.80	4.59	Head and Stuart, 1980
Lake Hordorli	−2.533	140.55	680	mire	micro, pollen slide	63.45	0.95	Hope and Tulip, 1994
Lake Johnston	−41.8666	145.55	900	lacustrine	micro, pollen slide	11.72	4.52	Dodson et al., 1998; Anker et al., 2001
Lake Majo	−1.46667	127.4833	140	lacustrine	micro, pollen slide	5.80	4.14	Haberle and Ledru, 2001
Lake Mountain	−37.469	145.875	1450	mire	micro, pollen slide	8.66	2.42	McKenzie, 1997
Lake Selina	−41.8833	145.6	516	lacustrine	micro, pollen slide	130.24	0.62	Colhoun et al., 1999
Lake Surprise	−38.0612	141.9223	93	lacustrine	micro, pollen slide	33.30	2.34	Builth et al., 2008
Lake Tyrrell1	−35.31	142.78	42	lacustrine	micro, pollen slide	17.46	2.98	Longmore et al., 1986; Luly et al., 1986; Luly, 1993, 1998

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Table 1 (*continued*)

Site Name	Latitude (°)	Longitude (°)	Elevation (m)	Site type	Charcoal methods	Record length (age ka)	Resolution (samples/ka)	References
Lake Tyrrell2	−35.31	142.78	42	lacustrine	micro, pollen slide	11.69	12.07	Longmore et al., 1986; Luly et al., 1986; Luly, 1993, 1998
Lake Wangoom LW87 core	−38.35	142.6	100	lacustrine	micro, pollen slide	196.60	0.52	Harle et al., 2002
Laravita	−8.3909	147.352	3570	mire	micro, pollen slide	16.14	1.92	Hope, 2009
Lashmars Lagoon	−35.8	138.0667	2	lacustrine	micro, pollen slide	12.95	5.41	Clark, 1983
Laukutu Swamp	−9.4794	160.0854	20	mire	micro, pollen slide	3.92	3.32	Haberle, 1996
Loch Sport Swamp	−37.9666	147.6833	2	mire	micro, pollen slide	9.26	3.13	Hooley et al., 1980
Lombok Ridge Core G6-4	−10.7833	118.0667	−3510	marine	micro, pollen slide	375.20	0.36	Wang et al., 1999
Long Swamp	−38.0833	141.0833	2	mire	micro, pollen slide	6.70	4.92	Head, 1988
Lynchs Crater	−17.3667	145.7	760	lacustrine	macro, sieved	234.75	1.11	Kershaw et al., 2007
Lynchs Crater (holocene core)	−17.3667	145.7	760	lacustrine	micro, pollen slide	7.39	3.52	Kershaw, 1983
Mago Island	−17.44	179.157	2	mire	micro, pollen slide	7.84	2.04	Hope et al., 2009
Main Lake Tower Hill	−38.3167	142.3667	20	lacustrine	micro, pollen slide	12.91	4.34	D'Costa et al., 1989
Maluyo Swamp	18.18	121.58	5	terrestrial	macro, sieved	6.26	95.79	J. Stevenson, unpublished data
McKenzie Road Bog	−38.4333	146.7667	50	mire	micro, pollen slide	0.17	301.17	Robertson, 1986
MD97-2140	2.0667	142.2667	−2547	marine	micro, imaging	364.65	0.32	Thevenon et al., 2004
Mela Swamp	−9.4794	160.0854	20	mire	micro, pollen slide	5.01	2.60	Haberle, 1996
Micalong Swamp	−35.3333	148.5167	1100	mire	micro, pollen slide	15.95	3.45	Kemp, 1993
Middle Patriarch Swamp	−39.998	148.181	19	lacustrine	micro, pollen slide	12.28	1.55	Ladd et al., 1992
Mill Creek	−33.4043	151.0303	4	terrestrial	micro, pollen slide	10.54	3.51	Devoy et al., 1994
Mountain Lagoon	−33.5	150.5167	604	lacustrine	micro, pollen slide	23.82	0.21	Robbie, 1998
Muellers Rock	−35.39	148.5	1102	mire	micro, pollen slide	10.57	1.99	Worthy et al., 2005
Mulloon	−35.4417	149.5567	799	mire	micro, pollen slide	3.95	5.07	G. Hope, unpublished data
Nadrau	−17.75	177.88	680	terrestrial	micro, pollen slide	2.25	13.78	Hope et al., 2009
Native Companion Lagoon	−27.6754	153.4107	20	coastal	micro, pollen slide	34.50	1.68	Petherick et al., 2008
Navatu	−18.07	178.53	4	mire	micro, pollen slide	8.30	2.41	Hope et al., 2009
Nekkeng	7.45	134.52	9	terrestrial	micro, pollen slide	9.06	0.77	Athens and Ward, 2005
Neon	−8.4725	147.3099	2875	mire	micro, pollen slide	13.00	1.54	Hope, 2009
Newall Creek	−42.07	145.44	140	terrestrial	micro, pollen slide	23.09	1.78	Van de Geer et al., 1989
Newnes Swamp	−33.3825	150.2222	1060	mire	micro, pollen slide	13.15	1.37	Chalson, 1991
Ngardmau	7.608	134.57	10	terrestrial	micro, pollen slide	5.30	3.02	Athens and Ward, 2005
Ngerchau	7.63	134.52	9	terrestrial	micro, pollen slide	4.80	2.50	Athens and Ward, 2005
Ngerdok 2	7.52	134.603	25	lacustrine	micro, pollen slide	3.67	8.45	Athens and Ward, 2005
Ngerkell	7.605292	134.6259	10	terrestrial	micro, pollen slide	3.97	2.02	Athens and Ward, 2005
Nong Pa Kho	17.01	102.93	180	mire	micro, pollen slide	37.92	1.87	Penny and Kealhofer, 2004
Noreikora Swamp	−6.3333	145.8333	1750	marsh	micro, pollen slide	6.87	3.49	Haberle and Ledru, 2001
North Torbreck	−37.4814	146.9472	564	fluvial	micro, pollen slide	13.48	1.56	McKenzie, 2002
Northwest Crater Tower Hill	−38.3167	142.3667	20	lacustrine	macro, sieved	25.85	3.13	D'Costa et al., 1989
Notts Swamp	−33.8098	150.4076	682	mire	micro, pollen slide	7.94	2.14	Chalson, 1991
Nursery Swamp	−35.41	148.58	1092	mire	micro, pollen slide	11.28	2.13	Rogers and Hope, 2006
Oaks Creek	−37.5856	146.1667	610	mire	micro, pollen slide	6.50	4.00	McKenzie, 2002
ODP Site 820	−16.6333	146.3	−280	marine	micro, pollen slide	100.50	1.10	Moss and Kershaw, 2000
Okarito Pakihi	−43.2417	170.2167	70	mire	macro, sieved	150.12	1.10	Vandergoes et al., 2005; Newnham et al., 2007b
Olbed 1	7.5	134.54	20	terrestrial	micro, pollen slide	7.98	2.63	Athens and Ward, 2005
Paoay Lake LP3	18.2	120.54	15	lacustrine	macro, sieved	6.70	87.11	Stevenson et al., 2009
Paoay Lake LP4	18.12	120.54	15	lacustrine	macro, sieved	6.49	35.58	Stevenson et al., 2009
Pemerak Swamp	0.7888	112.05	40	mire	micro, pollen slide	35.54	0.82	Anshari et al., 2001
Penrith Lakes	−33.7139	150.6774	18	lacustrine	micro, pollen slide	37.88	1.74	Chalson, 1991
Pine Camp	−34.75	141.13	21.4	lacustrine	micro, pollen slide	30.76	1.56	Cupper, 2005, 2006
Plum Swamp	−22.26	166.61	40	mire	micro, pollen slide	23.05	3.38	Stevenson, 1998
Poets Hill	−41.883	145.559	600	lacustrine	micro, pollen slide	14.16	2.33	Colhoun, 1992
Poley Creek	−37.4078	145.2189	630	mire	micro, pollen slide	19.24	1.77	Pittock, 1989
Powelltown	−37.8667	145.7031	168	mire	micro, pollen slide	7.27	3.30	McKenzie, 2002
Quambie Lagoon	−12.5	131.17	20	lacustrine	macro, sieved	6.54	17.59	J. Stevenson, unpublished data
Queens Swamp Core QS 3	−33.9002	150.5916	665	mire	macro, sieved	10.67	22.21	S. Mooney, unpublished data

Rawa Danau	−6.1833	105.9667	100	lacustrine	micro, pollen slide	16.59	1.87	Haberle and Ledru, 2001
Redhead Lagoon	−32.9944	151.7208	65	lacustrine	micro, pollen slide	79.79	1.25	Williams, 2005
Rennix Gap	−36.22	148.3	1570	terrestrial	micro, pollen slide	11.94	3.02	G. Hope, unpublished data
Ringarooma Humus Site 1	−41.3	147.6167	885	terrestrial	micro, pollen slide	0.20	110.00	Dodson et al., 1998
Ringarooma Humus site 2	−41.3	147.6167	885	terrestrial	micro, pollen slide	0.32	53.13	Dodson et al., 1998
Ringarooma River site I	−41.3	147.6167	885	terrestrial	micro, pollen slide	0.39	56.24	Dodson et al., 1998
Ringarooma River Site II	−41.3	147.6167	885	terrestrial	micro, pollen slide	0.28	61.40	Dodson et al., 1998
Rock Arch Swamp	−34.3	150.39	575	mire	micro, pollen slide	10.87	1.20	Hope, 2005a
Rotten Swamp (high-res study)	−35.7	148.8833	1445	mire	micro, pollen slide	6.31	56.74	Hope and Clark, 2008
Rotten Swamp Core 4	−35.7	148.8833	1445	mire	micro, pollen slide	11.41	1.58	Clark, 1986
Ryans Swamp	−35.09	150.39	8	mire	micro, pollen slide	4.45	6.29	Radclyffe, 1993
Saint Louis Lac	−22.23	166.55	5	mire	micro, pollen slide	7.19	8.63	Stevenson, 2004
Sapphire Swamp	−30.0341	151.5604	1260	mire	micro, pollen slide	0.35	62.86	Dodson et al., 1986
Sari	−16.63	179.5	67	mire	micro, pollen slide	6.53	3.52	Hope et al., 2009
Snobs Creek	−37.558	145.928	775	mire	micro, pollen slide	15.32	2.02	McKenzie, 1997
Snowy Flats	−35.54	148.47	1618	mire	micro, pollen slide	9.18	2.72	Hope et al., 2005
Soleve	−17.25	−179.49	2	mire	micro, pollen slide	7.16	2.93	Clark and Hope, 1997
Solomons Jewel Lake	−41.8	146.2667	1185	lacustrine	micro, pollen slide	4.85	10.10	Dodson, 2001
Sondambile	−6.3453	147.1116	2850	lacustrine	macro, sieved	1.34	243.12	Haberle et al., 2005
South Salvation Creek Swamp	−33.6326	151.2587	132	mire	micro, pollen slide	4.78	4.19	Kodela and Dodson, 1988
Stockyard Swamp, Hunter Island	−40.55	144.75	65	mire	micro, pollen slide	5.83	2.40	Hope, 1999
Storm Creek	−37.45	145.8	1177	Mire	micro, pollen slide	29.33	1.40	McKenzie, 1997
Sundown Swamp	−41.1667	144.6667	10	mire	micro, pollen slide	4.15	2.65	Hope, 1999
Supulah Hill	−4.1167	138.9667	1580	mire	micro, pollen slide	39.00	0.95	Haberle et al., 1991; Hope, 1998
Tadpole Swamp	−38.1317	145.2756	60	mire	micro, pollen slide	10.24	1.85	Aitken and Kershaw, 1993
Tagamaucia	−16.49	−179.56	820	lacustrine	micro, pollen slide	16.36	3.79	Hope, 1996a
Talita Kupai	−10.1	142.12	33	coastal	macro, sieved	2.41	61.72	Rowe, 2006a
Tea Tree Swamp Core DRA	−37.2167	148.8333	900	mire	micro, pollen slide	0.03	490.20	Gell et al., 1993
Tea Tree Swamp Core DRE	−37.2167	148.8333	900	mire	micro, pollen slide	0.52	29.12	Gell et al., 1993
Tea Tree Swamp Core DRN-A	−37.2167	148.8333	900	mire	micro, pollen slide	0.54	126.80	Gell et al., 1993
Tiam Point	−10.12	142.18	3	coastal	micro, pollen slide	7.70	6.62	Rowe, 2006a, 2007
Tiger Snake Swamp	−38.1313	145.2758	60	mire	micro, pollen slide	8.41	2.73	Aitken and Kershaw, 1993
Tom Burns (D-section Core)	−37.3833	145.8167	1075	Mire	micro, pollen slide	12.07	2.57	McKenzie, 1997
Tom Burns (Missen Core)	−37.3833	145.8167	1075	Mire	micro, pollen slide	31.56	0.48	McKenzie, 1997
Tom Gregory Swamp	−35.38	148.49	1024	mire	micro, pollen slide	12.39	1.29	Hope, 2005b
Tugupugua	−5.6691	142.6108	2300	mire	micro, pollen slide	18.42	3.42	Haberle and Ledru, 2001
Tyrendarra Swamp	−38.1986	141.7626	13	marsh	micro, pollen slide	31.63	1.04	Builth et al., 2008
Voli Voli	−18.16	177.485	2	mire	micro, pollen slide	5.87	7.15	Dickinson et al., 1998
Vunimoli	−18.22	177.88	251	mire	micro, pollen slide	4.75	2.11	Hope et al., 1996, 1999
Waikaremoana	−38.45	177.02	582	lacustrine	micro, pollen slide	1.65	9.11	Newnham et al., 1998
Waitabu	−18.23	−178.781	43	other	micro, pollen slide	1.71	18.16	Latham et al., 1983
Wanda	−2.33	121.23	440	mire	micro, pollen slide	49.30	1.24	Hope, 2001
Wanum	−6.6365	146.7986	35	lacustrine	macro, sieved	3.89	72.92	Haberle et al., 2005
Warrananga	−33.97	141.56	22	lacustrine	micro, pollen slide	10.96	4.93	Cupper, 2005, 2006
Warrimoo Swamp	−33.7226	150.6162	195	mire	micro, pollen slide	4.63	6.27	Chalson, 1991
Waruid	−10.4	142.09	5	coastal	macro, sieved	7.20	27.10	Rowe, 2006a, 2007
Wet Lagoon	−34.47	149.25	700	mire	micro, pollen slide	2.30	12.17	Dodson, 1986
Whitehaven Swamp	−20.3	148.9	45	mire	micro, pollen slide	8.04	6.60	Genever et al., 2003
Wildes Meadow Swamp	−34.6208	150.5111	670	mire	micro, pollen slide	3.62	3.04	Kodela, 1996
Wilson Bog	−34.97	138.69	425	mire	macro, sieved	7.18	13.23	Buckman et al., 2009
Wingecarribee Swamp W2	−34.5666	150.5166	685	mire	macro, sieved	3.50	124.68	de Montford, 2008
Worimi Swamp	−32.5166	152.3333	8	mire	macro, sieved	2.46	36.53	Mooney and Maltby, 2006
Wyelangta	−38.6472	143.4614	450	fluvial	micro, pollen slide	223.93	0.23	McKenzie and Kershaw, 2000
Xere Wapo B	−22.29	166.97	220	lacustrine	micro, pollen slide	126.04	0.52	Stevenson and Hope, 2005
Xere Wapo C	−22.29	166.97	220	lacustrine	macro, sieved	78.80	3.55	Stevenson and Hope, 2005
Xere Wapo D	−22.29	166.97	220	lacustrine	macro, sieved	80.00	3.26	Stevenson and Hope, 2005
Yacata	−17.156	−179.51	2	mire	micro, pollen slide	6.50	4.77	Hope et al., 2009
Yano	7.38	134.54	40	terrestrial	micro, pollen slide	7.54	3.72	Athens and Ward, 2005

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Table 1 (*continued*)

Site Name	Latitude (°)	Longitude (°)	Elevation (m)	Site type	Charcoal methods	Record length (age ka)	Resolution (samples/ka)	References
Yaouk Swamp	−35.49	148.5	1100	lacustrine	micro, pollen slide	12.01	2.50	J. Keaney and G. Hope, unpublished data
Yawi Ti	−6.6147	143.8863	1150	mire	macro, sieved	16.87	6.88	Haberle, 2007
Zurath Islet	−10.16	142.06	3	coastal	macro, sieved	11.83	5.41	Rowe, 2006a, 2007

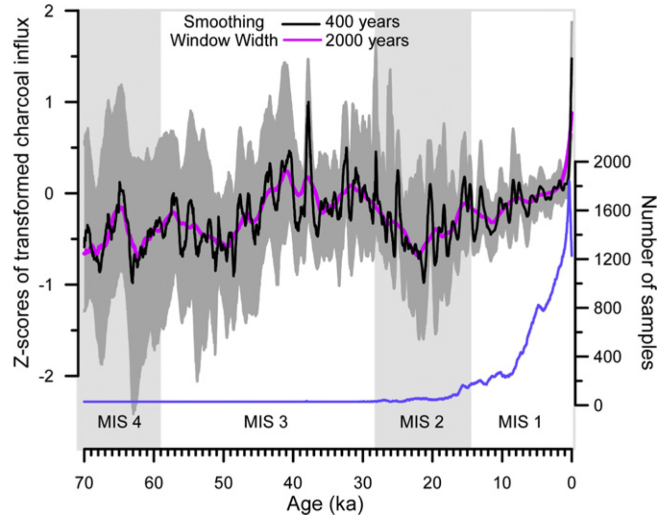

Fig. 2. Reconstruction of biomass burning over the period from 70 ka to present for Australasia as a whole (20°N - 50°S; 100°E to 177°W). The curves have been smoothed using a window of 2000 years (purple curve) to emphasize the long-term trends and a window of 400 years (bold black curve) to emphasize the millennial-scale variability. The number of observations contributing to the record within each 400-year window is also shown (blue curve). The limit for Marine Isotope Stage (MIS) 1 is defined as 14.7 ka to present, MIS 2 (27.8–14.7 ka), MIS 3 (59.4–27.8 ka), and MIS 4 (73.5–59.4 ka) (Sánchez Goñi and Harrison, 2010). (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article).

cooling culminating in the Little Ice Age. This is seen in the composite record from tropical Australasia (Fig. 5b) but not in the record from temperate Australasia (Fig. 5c). There is a pronounced increase in burning after ca 1800 A.D. in both the ITCZ (Fig. 5b) and

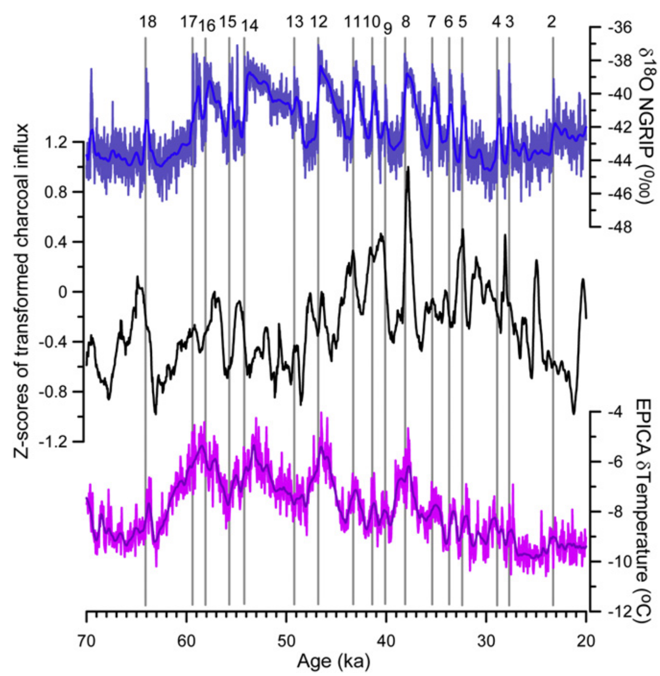

Fig. 3. Millennial-scale variability in biomass burning during the glacial (70-20 ka) compared to temperature indices from the NGRIP and EPICA ice-core records. Charcoal data are summarized using a lowess curve with a 400-year half-window width (black). For comparison with the charcoal curve, the 20-year sampling resolution NGRIP record (agecal. yr b1950) and EPICA are shown, along with a 400-year smoothed curve (bold lines in blue and purple, respectively). The numbered vertical lines mark the Dansgaard–Oeschger warming events. 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). (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article).

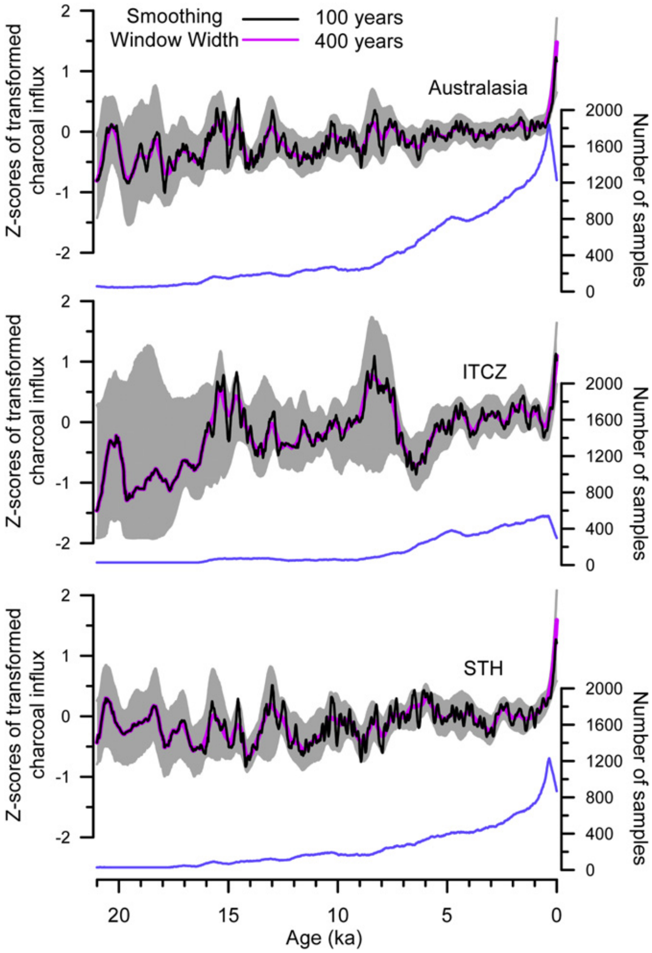

Fig. 4. Reconstruction of biomass burning for the last 21 ka for Australasia as a whole (20°N-50°S, 100°E-177°W), the belt corresponding to the modern Inter-tropical Convergence Zone (ITCZ: 20°N-20°S, 100°E to 177W, broadly the tropical region) and the modern sub-tropical high pressure belt (STH: 25°S-45°S, 100°E to 177°W, broadly temperate Australasia). The curves have been smoothed using a window of 400 years (purple curve) and a window of 100 years (bold black curve) to emphasize the centennial-scale variability. The bootstrapped confidence intervals are based on a 400-year smoothing of the curves. The number of observations contributing to the record within each 400-year window is also shown (blue curves). (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article).

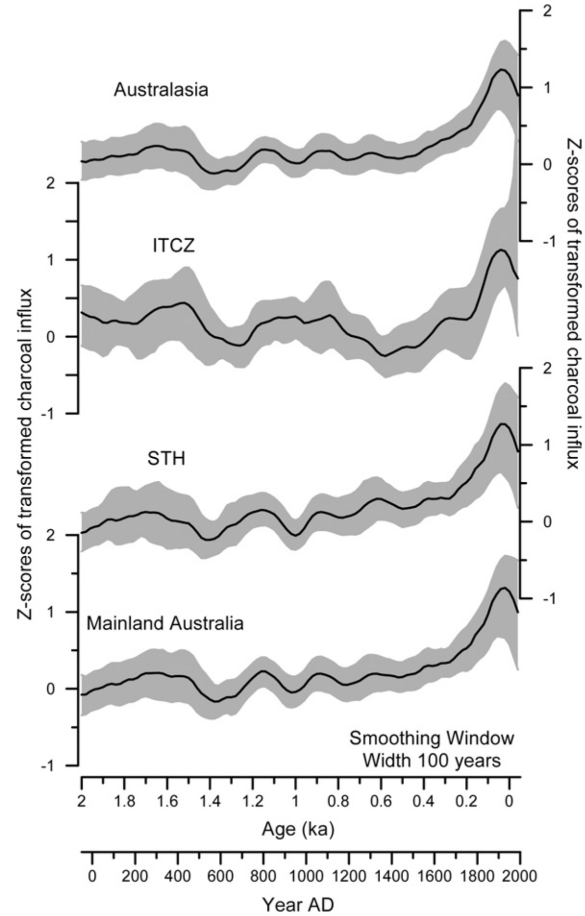

Fig. 5. Reconstruction of biomass burning for the last 2 ka for Australasia as a whole (20°N-50°S, 100°E-177°W), the belt corresponding to the modern Inter-tropical Convergence Zone (ITCZ: 20°N-20°S, 100°E to 177W, broadly the tropical region), the modern sub-tropical high pressure belt (STH: 25°S-45°S, 100°E to 177°W, broadly temperate Australasia), and for all sites on the Australian mainland. The curves have been smoothed using a window of 100 years (bold black curve).

STH regions (Fig. 5c), and this increase is also seen in the composite record of the Australian mainland (Fig. 5d). The upturn is delayed compared to the timing of the increase in global biomass burning which occurred ca 1750 A.D (Marlon et al., 2008). Although European colonization of Australia may have been responsible for the upturn seen on the mainland, and in the composite regional curve, it is more difficult to invoke this explanation for the record from tropical Australasia (Fig. 5b) which suggests that any anthropogenic influence was exacerbated by changes in climate and vegetation productivity associated with the post-industrial increase in atmospheric CO_2 concentration. The curves for all of the regions show reduced fire during the last *ca.* 50 years, and this is despite the widespread use of prescribed burning as a means of fire control in much of Australia. The observed reduction in biomass burning in Australasia occurs a few decades after the global reduction in biomass burning which Marlon et al. (2008) attribute to increased landscape fragmentation and fire suppression. There are a number of reasons why the last fifty or so years of the record might be less tightly constrained, including issues with sampling, problems with dating and the potential impact of human activities on catchment processes and sedimentation (Gale, 2009). Nevertheless, there are over 140 sites contributing to the record during the 20th century and 168 sites recording the last two centuries, and thus the recent downturn in fire appears to be a robust feature of the record. Furthermore, Marlon et al. (2008) demonstrate (in their supporting information) that the rapid increase in biomass burning followed by an abrupt decrease cannot be explained by simple sedimentation rate variations at the top of cores.

2.3. Spatial patterns in biomass burning since the Last Glacial Maximum

Most of the sites in our data set are concentrated in eastern, particularly southeastern Australia (Fig. 1a). Nevertheless, there are spatial patterns in the changes in biomass burning that are worth exploring. Here (Fig. 6), we present maps of the average z-scores for key intervals since the LGM to illustrate some of the regional patterns of change: positive z-scores indicate more biomass burning and negative z-scores less biomass burning than the long-term average for the base period (21–0.2 ka).

The LGM (Fig. 6a) was characterized by low biomass burning in the tropics and over most of Australia. A few sites in southeastern Australia and one site on the South Island of New Zealand show higher-than-average z-scores. There is considerable millennial-

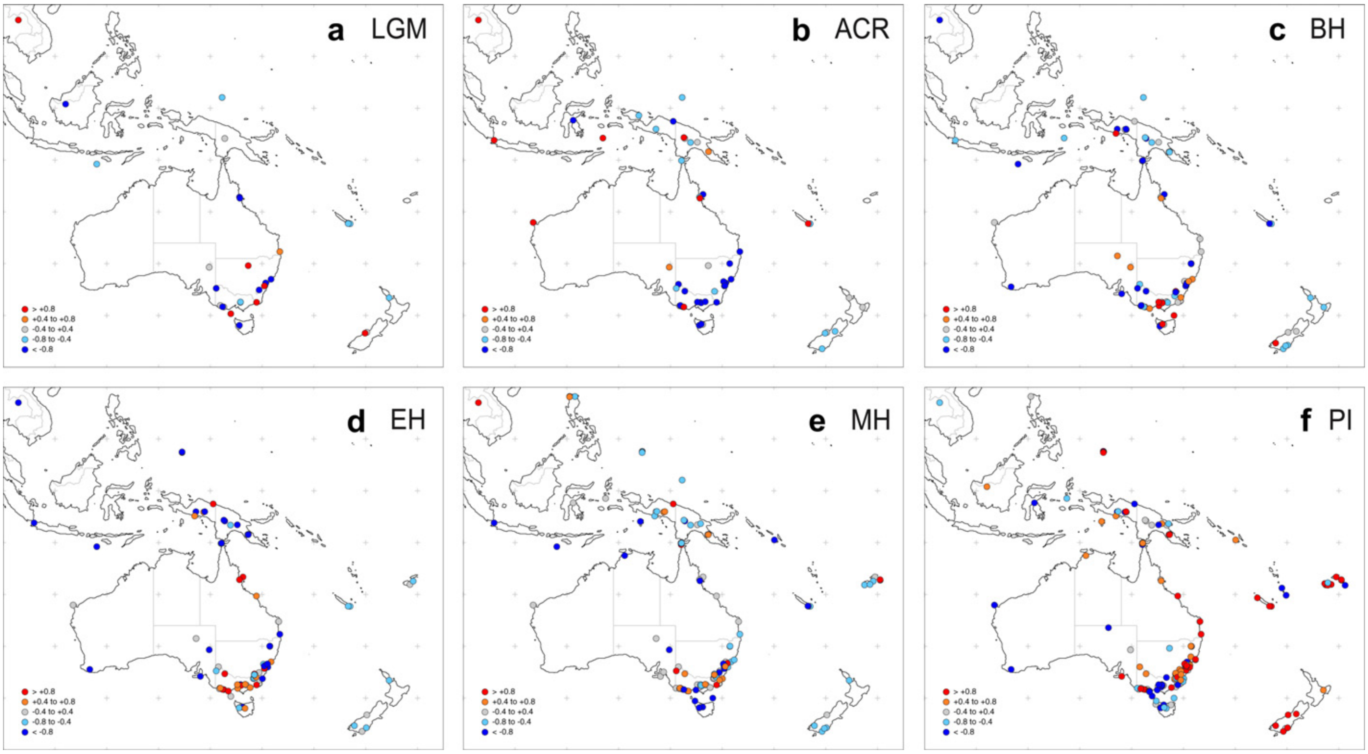

Fig. 6. Reconstructions of the geography of changes in fire regimes, as expressed by z-score anomalies from the long-term mean over the base period between 21-0.2 ka, at key time including (a) the Last Glacial Maximum (LGM, ca 21 ka), (b) the Antarctic Cold Reversal, (ACR, 14.2–11.8 ka) (c) the beginning of the Holocene (BH, 10.5–9.5 ka), (d) the early Holocene (EH, 8.5–7.5 ka) (e) the mid-Holocene, (MH, 6-5 ka), (f) the pre-industrial period (PI, 0.7–0.2 ka).

scale variability superimposed on the long-term increase in biomass burning after the LGM (Fig. 4) and this is reflected in the mapped patterns, with individual sites showing more/less fire in succeeding 1000-yr intervals (not shown). However, a more coherent spatial pattern is established during the Antarctic Cold Reversal (ACR) (Fig. 6b), with low biomass burning in southeastern Australia and New Zealand contrasting with high biomass burning at many sites in the tropics. In contrast, the beginning of the Holocene is marked by high biomass burning in southeastern Australia and low biomass burning in the tropics (Fig. 6c).

The most marked spatial feature of the Holocene record is the opposition in the temperate latitudes between sites in southernmost southeastern Australia (including Tasmania) and southeastern NSW (including the interior). In the early Holocene, here illustrated by the interval centered on 8 ka (Fig. 6d), the southern region is characterized by high biomass burning and the region further north by low biomass burning. By the mid-Holocene, here illustrated by the interval centered on 5.5 ka (Fig. 6e), sites in Tasmania and the southernmost tip of the continent show less biomass burning and many sites further north show increased biomass burning. The pattern has reversed again by the late Holocene (not shown). The pre-industral era (0.2–0.7 ka, Fig. 6f) is characterized by high biomass burning along the east coast of Australia, in New Zealand and over much of the tropics. However, Tasmania and the southernmost part of southeastern Australia, and the limited number of sites from western Australia, show less biomass burning than average.

Although there are robust, large-scale patterns in changes in biomass burning through time, the changes within any one region through time are complex; adjacent sites can show differences even in the sign of the change at particular times. This may, in part, be a function of the quality of the data and of individual age models. It is also in the nature of wildfires, which are spatially disjunct, influenced by topography and the changing nature of the vegetation itself, and conditioned by timing of previous fires.

2.4. Fire and humans during the late Pleistocene

Smith et al. (2008) have used a data set of 971 radiocarbon ages from 286 archaeological sites in arid Australia (AustArch1) to illustrate the overall trend in archaeological site incidence by summing the probability density functions of individual radiocarbon ages, which they suggest can be interpreted as a first approximation of human activity and population history through time. Smith et al. (2008) stress the preliminary nature of this record, but nevertheless the compilation allows us to explore the relationship between an approximate measure of human activity and the charcoal-derived record of fire from Australia (Fig. 7). Although there are other compilations of archaeological information from Australia (e.g. Lourandos and David, 1998; Lourandos and David, 2002; Turney and Hobbs, 2006), the AustArch1 data set is the most comprehensive and the raw data are available in a format which allows reanalysis.

There are relatively few archaeological dates, and little structure in the probability density curve, prior to *ca*. 20 ka. In contrast, there are large changes in fire on millennial timescales between 40-20 ka. The lack of congruence between the archaeological and fire records during this period suggests that the changes in fire activity do not reflect changes in human activity. Smith et al. (2008) identified six intervals of increased "human activity" during the past 20 ka, including a major increase in the late Holocene. Some of the reconstructed peaks in human activity correspond to peaks in fire (e.g. 7-8 ka). However, other peaks in human activity correspond to troughs in biomass burning (e.g. 5–4.5 ka) and there are peaks in biomass burning that have no correspondence with changes in human activity (e.g. at 13 ka and 9.5 ka). Most importantly, the

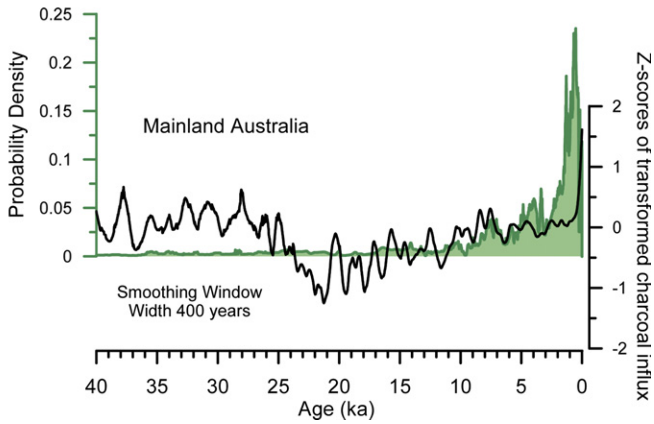

Fig. 7. Comparison of the composite charcoal curve for sites from the Australian mainland over the last 40 ka (black curve) compared to probability density estimates of human populations based on radiocarbon-dated archaeological records (green infilled curve). Population data from the AustArch Database (see Smith et al., 2008 for fuller description of the methodology). (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article).

major inferred increase in population during the late Holocene ('intensification') is not accompanied by an increase in fire. Thus, this comparison does not support the hypothesis that changes in post-glacial fires in Australia were caused by humans.

3. Discussion and conclusions

We have presented an analysis of the long-term changes in fire regimes across Australasia based on a comprehensive synthesis of over 200 charcoal records from the region. Despite the supposed importance of fire for Australasian vegetation (Bowman, 2000; Bradstock et al., 2002), the long-standing arguments about the role of Aboriginal colonisation on fire regimes (e.g. Singh et al., 1981; Horton, 2000; Flannery, 1994; Kershaw et al., 2002; Black and Mooney, 2006), and the uncertainty about the impact of climate change on fire regimes and fire-related disasters (Chapman, 1999; Williams et al., 2001; Cary, 2002; Bushfire Co-operative Research Centre, 2006; Russell-Smith et al., 2007), there have been relatively few analyses of the palaeo-record of fire at a continental scale.

Previous reviews (see e.g. Singh et al., 1981; Kershaw et al., 2002; Lynch et al., 2007) have been based on comparatively few sites and the assumption that iconic records are characteristic of much broader geographic regions. Although there are obvious spatial gaps in the data set used in this analysis, and some uncertainty associated with pooling records with age models of very different quality, we have been able to document robust changes in fire regime through time for which there are physically plausible explanations. The analysis presented here is preliminary in nature: much more could be done to explore the existing data set. Nevertheless, by demonstrating the potential of large-scale syntheses of charcoal data to shed light on the environmental history of the continent we hope to encourage the collection of additional, high-quality data from the Australasian region.

On glacial-interglacial timescales, changes in fire regimes closely follow global temperature: biomass burning is reduced during cold glacial or stadial intervals (MIS 4, MIS 2) and increased during warmer interglacial or interstadial intervals (MIS 3, the Holocene). This finding is consistent with global analyses (e.g. Power et al., 2008; Daniau et al., 2010) and presumably reflects the strong control of vegetation productivity on the availability of fuel (Harrison et al., 2010). Although we have confined our detailed analyses to the past 70 ka, we see no evidence for the expression of long-term aridification of Australia (see e.g. Lynch et al., 2007) in the charcoal records of fire regimes. The length of this interval of increasing aridity cited in the literature is somewhat vague, but we see no evidence for a long-term trend in biomass burning superimposed on the glacial-interglacial pattern.

The charcoal records show considerable millennial-scale variability during the glacial. Given the limited radiometric age control on these records, and the widespread use of correlation to assumed stage/event boundaries, it is unclear whether individual peaks in charcoal correlate with specific D-O warming events registered in the Greenland ice core and reflected in marine records from the North Atlantic. Nevertheless, the shape of the charcoal curves is more reminiscent of the temperature changes in Greenland than those recorded in Antarctica (Fig. 3) and, in conjunction with the number of events registered and the approximate temporal correlations of these events, suggests that Australasian fire regimes have co-varied with D-O cycles: with increased fire during warming events and Greenland Interstadials and reduced fire during Greenland Stadials. Millennial-scale variability has been identified in several individual records from Australasia (see e.g. Sikes et al., 2002; Turney et al., 2004; Vandergoes et al., 2005; Kershaw et al., 2007), although with the caveat that the dating was insufficient to determine whether this variability was in or out of phase with the D-O cycles. However, Muller et al. (2008) have stated explicitly that warmer intervals of the D-O cycles are associated with dry conditions at Lynch's Crater.

EPICA Community Members (2006) have demonstrated that the temperature records from Greenland and Antarctica are out of phase, and this bi-polar temperature seesaw is attributed to the shutdown or slowdown of the thermohaline circulation (see Kageyama et al., 2010). However, the bi-polar seesaw hypothesis does not imply that the opposition in the direction of the change in temperature shown by the ice-core records is a pan-hemispheric phenomenon. In fact, decomposition of climate variability during the deglaciation (Shakun and Carlson, 2010), based on 104 high-resolution palaeoclimate records, has shown that approximately 60% of the climate signal is common between the two hemispheres, while only 11% of the signal is associated with the bi-polar seesaw. Shakun and Carlson (2010) attributed the hemispheric synchroneity in climate to the influence of CO_2, although presumably this argument would apply to all of the greenhouse gases. The greenhouse gas synchronisation of hemispheric climate changes likely operated during the whole of the glacial, where D-O variability was accompanied by changes in CO_2 of the order of 15–20 ppm and CH_4 values of the order of about 100 ppm, or about 20% and 50% of their glacial-interglacial range, respectively (Ahn and Brook, 2008). There are other mechanisms that could extend a northern-hemisphere climate signal into the southern hemisphere, through changes in atmospheric circulation that accompany THC shutdown, as shown by freshwater-forcing experiments (Stouffer et al., 2006: Muller et al., 2008; Kageyama et al., 2009, 2010). Thus, the observation that Australasian fire regimes display some coherence with millennial-scale northern-hemisphere climate variability is mechanistically credible (Clark et al., 2002; Muller et al., 2008; Daniau et al., 2010). Furthermore, Southern Hemisphere vegetation records show millennial-scale variability (see e.g. Hessler et al., 2010; Harrison and Sanchez Goni, 2010), although the vegetation response appears to be more muted than shown in European records (Fletcher et al., 2010) it is nevertheless in phase with D-O events (Harrison and Sanchez Goni, 2010).

The pre-40 ka composite biomass-burning curve for Australasia is based on 25 individual records and this value rises to 38 sites by 30 ka. There is clearly a need for more long records, sampling a wider geographic range of regional climates. Charcoal records from marine cores are providing an increasingly detailed view of changes in fire regimes during the glacial (see e.g. Beaufort et al., 2003; Thevenon et al., 2004; Daniau et al., 2007, 2009) with the added advantage that these records can be linked, through isotope stratigraphy, directly to changes in sea-surface temperatures and to changes recorded in ice cores. Marine records could be more widely exploited for charcoal analysis in the Australasian region.

Low biomass burning was characteristic of the LGM. Vegetation reconstructions show an expansion of xerophytic vegetation in southern Australia and of tropical deciduous broadleaf forest and woodland in the tropics (both mainland Australia, Papua and the islands of southeastern Asia) at the LGM (Pickett et al., 2004). Although the number of records included in this synthesis is limited, particularly for northern Australia, the interpretation is consistent with earlier reconstructions (see e.g. Markgraf et al., 1992; Harrison and Dodson, 1993) showing expansion of drought-tolerant vegetation during the LGM. Although superficially these changes could be interpreted as an expansion of more fire-prone vegetation, they have been interpreted as indicating colder and drier conditions. Indeed, Williams et al. (2009) argue for a reduction in precipitation of ca 30–50% in the tropics and a significant reduction in winter precipitation across the temperate zone, accompanied by reductions in temperature of ca 3–8 °C. The observed reduction in biomass burning must reflect a significant reduction in fuel availability under cold, dry conditions, sufficient to offset the increase in aridity and also the stronger winds that have been adduced from dune evidence (e.g. Hesse et al., 2004).

Previous syntheses of palaeoenvironmental data from Australia have identified the late glacial as an interval of aridity, more pronounced than during the LGM and certainly than during the Holocene (see e.g. Harrison and Dodson, 1993; Kershaw et al., 2003; Williams et al., 2009). The charcoal records show considerable spatial and temporal variability during the deglaciation. There is no evidence for dry conditions persisting for several thousand years, as postulated by earlier syntheses. However, the charcoal records do appear to show a reduction in biomass burning during the ACR. The ACR has been identified in palaeoenvironmental records from southeastern Australia and New Zealand (see e.g. Barrows et al., 2007; Calvo et al., 2007; Williams et al., 2009), although the recognition that this climate event could have had an impact on fire regimes at a continental scale is new.

Vegetation reconstructions (see e.g. Harrison and Dodson, 1993; Pickett et al., 2004) indicate that the mid-Holocene vegetation patterns of Australia were very similar to those typical of the pre-industrial (and pre-European) era. Despite the lack of vegetation change, Lynch et al. (2007) indicate that the interval between 7-5 ka (which they call the “climatic optimum”) was one of reduced biomass burning in southeastern Australia. They do not specify which records this assertion is based on, and their conclusion is certainly not supported by the records presented here: most sites in southeastern Australia show considerably higher-than-average levels of fire at 6 ka, although sites in the interior (and the limited number of sites in western and northern Australia) show less-than-average biomass burning.

We have identified a tendency for sites in the southernmost part of southeastern Australia to show an opposite pattern of change in biomass burning during the Holocene from those further north and along the east coast. This spatial differentiation was also identified by Pickett et al. (2004) in their reconstruction of mid-Holocene vegetation shifts. They argued that sites in the far south show a shift towards more moisture-stressed vegetation in the mid-Holocene, while sites in the Snowy Mountains, on the Southern Tablelands and east of the Great Dividing Range have more moisture-demanding vegetation in the mid-Holocene than today. These shifts in vegetation are consistent with the reconstructed shifts in biomass burning, with the former region characterized by more fire and the latter by less fire during the mid-Holocene.

Lynch et al. (2007) have argued that the pronounced increased in biomass burning after *ca.* 4 ka shown in the charcoal record from Lake Euramoo (Haberle, 2005), and assumed to represent a regional signal, reflects an increases in climate variability associated with ENSO. There is considerable evidence for an increase in ENSO strength and/or variability after the mid-Holocene (see e.g. Rodbell et al., 1999; Tudhope et al., 2001; Andrus et al., 2002; Koutavas et al., 2002) but the onset of increased variability has been variously placed at 6.5 ka (Black et al., 2008), 5 ka (McGlone et al., 1992), 4 ka (Shulmeister and Lees, 1995) and 3 ka (Gagan et al., 2004) and reanalysis of the Lake Pallcacocha record (Rodbell et al., 1999) suggests that variations on the ENSO time scale persisted throughout the Holocene (Rodó and Rodriguez-Arias, 2004). Neither the composite nor regional charcoal records (Fig. 4) show a clear relationship between changes in ENSO (e.g. Moy et al., 2002) and patterns of biomass burning. This suggests the need for a re-evaluation of the relationship between biomass burning and changes in ENSO, over both late Holocene and longer timescales, taking into account that the sign of the relationship between biomass burning and ENSO may be different depending on the nature of the vegetation.

Lack of data has bedevilled previous interpretations of both the fire history and, more generally, the palaeoenvironmental history of Australasia. Reconstructed changes at individual sites have often been used to draw quasi-continental scale inferences about past climate and environmental changes. The fact that many of the conclusions drawn from such iconic sites are not supported by the large-scale data synthesis of charcoal records presented here argues for an urgent need to re-assess our understanding of the late Quaternary history of Australia.

Chronology has also been a contentious issue for the interpretation of palaeofire records from Australasia. Alternative age models have been proposed for several of the longer records included in this synthesis (see for e.g. Lynch's Crater: Kershaw, 1986; Turney et al., 2001, 2004; Kershaw et al., 2007; Muller et al., 2008; Lake George: Singh et al., 1981; Singh and Geissler, 1985; Fitzsimmons and Barrows, 2010). In the interpretation of individual records, the quality of the age model is of prime importance and reliance on assumed correlations with global events or stratigraphic boundaries (e.g. the MIS boundaries) precludes detailed analysis of the relationships between observed changes in fire and the global controls on regional climate. The strength of the technique used here of compositing multiple records 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. By focusing on statistically robust features of the composite record, we are able to draw plausible inferences about the controls on these features. Clearly, improvement of the individual age models and increased reliance on radiometric dating would substantially enhance the amount of detail that could be extracted from the composite records. Nevertheless, our analyses show that there is considerable temporal structure in the fire records and it is highly unlikely that this structure could be generated by chance, particularly since the reconstructed changes in biomass burning through time are consistent with current understanding of past climate changes.

The simplest interpretation of the charcoal records is that climate, and climate-modulated changes in vegetation productivity and distribution, control fire regimes on centennial to multi-millennial timescales. The observed changes in biomass burning can be plausibly explained by current understanding of the broadscale changes in climate over the last ca. 70 ka. However, there are many changes in the fire record for which we still lack a robust mechanistic explanation. As previous work has shown (e.g.

Power et al., 2008), simplistic explanations of fire records in terms of single climate (or environmental) drivers are likely to be wrong. Similar changes in climate can produce changes in biomass burning of opposite sign depending on the state of the vegetation (van der Werf et al., 2008b; Harrison et al., 2010). For example, an increase in precipitation will tend to suppress fires in most forests but could lead to an increase in fire in areas where fuel is limited because it will support more plant growth. Coeval changes in temperature can offset the impact of changes in precipitation on vegetation productivity. Atmospheric CO_2 concentrations influence vegetation productivity and distribution even without a change in climate (Harrison and Prentice, 2003; Prentice and Harrison, 2009). Finally, fire regimes are affected by the considerable heterogeneity of local climates. Sites in close geographic proximity may nevertheless have different climates, particularly with respect to rainfall seasonality. Shafer et al. (2005), for example, have shown that large-scale changes in atmospheric circulation lead to opposite changes in precipitation between sites in the same region depending on whether they experience a summer or winter-rainfall maximum and this in turn affects the response of the fire regime to these changes in atmospheric circulation (Millspaugh et al., 2004). Given all these competing influences, it is unlikely that interpretations of the charcoal record in terms of climate will be unequivocal. While speculations as to the causes of specific changes in fire regimes are interesting as a source of hypotheses, we would argue strongly for the combined use of observational data and modelling (see e.g. Webb et al., 1998; Harrison et al., 2003) to interpret the observed changes in fire regimes through time.

The influence of humans in modifying natural fire regimes has been a major feature of the interpretation of charcoal records from the Australasian region. The apparent coincidence of increases in sedimentary charcoal at iconic sites (Lake George, Lynch's Crater, Darwin Crater) with the arrival of Aboriginal people in Australia, has been attributed to anthropogenic fire (e.g. see Turney et al., 2004). This has led to circular arguments about the relationship between fire and people, including assertions that people arrived on the continent before the last glacial (Jackson, 1999; Singh et al., 1981). Changes in vegetation cover driven by anthropogenic modification of fire regimes have been explicitly invoked as a mechanism for causing aridification of Australia over the past 50-60 ka and for megafaunal extinctions (Miller et al., 2005). This causal association of anthropogenic fire with changes in climate, vegetation or fauna has remained seductive, despite several lines of contrary evidence. This evidence includes major changes in Late Quaternary charcoal records and hence fire prior to the arrival of humans (e.g. ca 130 ka: Singh et al., 1981; Dodson et al., 2005); the fact that a broadly-synchronous transition to more xerophytic vegetation has been found in New Caledonia, a region which was not settled by humans until ca 3000 years ago (Stevenson and Hope, 2005); and modeling evidence that the purported changes in vegetation cover were insufficient to cause a sustained change in Australasian climate (Pitman and Hesse, 2007). We have found no evidence of a change in fire regimes at a continental scale at the time of Aboriginal colonisation of Australia (50 ± 10 ka).

Changes in Aboriginal socio-economic relationships (Lourandos, 1980, 1983) have been invoked as causing changes in fire regimes in the mid-to-late Holocene. Lourandos (1980, 1983) suggested that the intensification of land use by Aboriginal groups, including a shift from more nomadic to more sedentary populations, was responsible for an increase in fire from *ca.* 5 ka onwards. Lynch et al. (2007) have also argued for an increase in fire associated with people, although they suggest that this occurred somewhat later after ca 3 ka. Again, we see no unequivocal relationship between inferred populations and/or the intensity of Aboriginal occupation of the continent and fire regimes during the past 21 ka, nor do the charcoal records show a marked increase in fire during the late Holocene. The charcoal records support the idea that fire regimes were controlled by changing climate and climate-induced changes in vegetation, though it is possible that these natural changes had an impact on Aboriginal populations through control of resources.

It is possible that our failure to identify a distinctive human fingerprint on fire activity at a continental or regional scale, either at the purported time of Aboriginal settlement of the Australian continent or over the past 21 ka, reflects the use of sedimentary charcoal records or of analytical techniques that emphasize the composite signal of many individual records. It could be argued, for example, that sedimentary charcoal is an unreliable indicator of the small-scale or low intensity fires characteristic of Aboriginal burning, particularly if the material burnt was primarily non-woody. However, studies from other areas (e.g. Brown et al., 2005) show that predominantly grassland fires are recorded by sedimentary charcoal. Furthermore, if sedimentary charcoal records fail to capture smaller-scale fires, then the use of individual charcoal records to identify human impact on the landscape is also suspect. In the absence of independent archaeological evidence, the interpretation of local charcoal signals in terms of human impact is at best equivocal as to causation even when accompanied by changes in vegetation.

It is more plausible to argue that the compositing of individual charcoal records at regional or continental scales is more likely to emphasize changes that are congruent with climate or environmental changes that operate at similar spatial scales and less likely to identify human impacts if these changes were time transgressive across a region or highly localized. The timing of Aboriginal settlement of Australia is subject to large uncertainty, and there is no clear agreement about whether settlement was time transgressive or not (see discussion in Bird et al., 2004). Furthermore, the debate about Aboriginal use of fire has focused on whether it resulted in widespread transformation of the Australian vegetation (see e.g. Flannery, 1994; Horton, 2000; Miller et al., 2005). Thus, while we cannot dismiss the idea that some individual records may contain an overprint of human influence on the local fire regime, the evidence presented here clearly demonstrates that the dominant control of fire activity is climate or climate-modulated changes in vegetation cover.

In conclusion, compilation and analysis of charcoal records from Australasia have allowed us to document changes in fire regimes over the past 70 ka and thus to examine (and resolve) some persistent controversies about the relationship between fire, climate and humans. We cannot, as yet, explain all of the features of the fire record nor have we exhausted the potential of the current database. The ongoing improvement of the database, through e.g. extension of the geographical coverage, higher resolution sampling, and improvement of age models, will allow further questions to be addressed. Community-based regional synthesis of charcoal and vegetation records, particularly if this can be combined with carefully designed model experiments, will continue to yield new insights into the behaviour of fire in response to climate and other environmental factors on palaeo-timescales.

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