Quaternary Research 55, 97–104 (2001)
doi:10.1006/qres.2000.2188, available online at http://www.idealibrary.com on IDEAL®

Correlations among Charcoal Records of Fires from the Past 16,000 Years in Indonesia, Papua New Guinea, and Central and South America

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Received December 15, 1999

Microscopic charcoal preserved in lake and swamp sediments from 10 sites in Indonesia and Papua New Guinea and from 5 sites in Central and South America have been used to reconstruct long-term fire histories for these two regions. Comparison of these records demonstrates that fire is promoted during periods of rapid climate change and high climate variability, regardless of the presence or absence of humans. Broad synchrony of changes in corrected charcoal values in each region supports an atmospheric transmission of the climate signal via the dominant large-scale atmospheric circulation systems (Walker Circulation) that appears to have persisted since 16,000 cal yr B.P. Altered climate boundary conditions under the influence of changing El Niño-related variability, insolation, sea level, and sea surface temperature all influenced the strength of this connection. Correlation of biomass burning records between the regions tends to increase in the Holocene. The main period of inverse correlation occurs during the Younger Dryas Stade, when extratropical climate most affected the tropics. The strongest correlation between the two regions postdates 5000 cal yr B.P., when El Niño-related variability intensified. Fluctuations in tropical biomass burning are at least partly controlled by orbital forcing (precession), although extratropical climate influences and human activity are also important.

Key Words: **fire; charcoal; Walker Circulation; El Niño; rain forest.**

INTRODUCTION

Despite the importance of fire and drought to agricultural production, forest management, biodiversity, and the global carbon cycle, there is little information on the long-term persistence or frequency of these events and the consequences for biotic communities. An evaluation of the paleoecological record of past fire events provides the best way of determining the driving forces behind changes in the frequency and intensity of these events through time. Records of ancient microscopic charcoal show that fires have repeatedly occurred, even in tropical forests, since at least the late Pleistocene, resulting in disturbance of forest ecosystems that in some cases have led to long-term depletion of rain forest communities (Kershaw *et al.*, 1997).

The extensive destruction of rain forest by fires across southeastern Asia and northern South America during the severe 1997–1998 El Niño event clearly demonstrated the susceptibility of large areas of rain forest to fire under conditions of extreme drought (Goldammer, 1999). Climatic conditions conducive to combustion in tropical rain forests are brought about by periodic failure of the tropical atmospheric circulation system to supply moisture to the land surface. In the region of Indonesia and Papua New Guinea, the severity of the 1997–1998 El Niño event was brought about by an anomalous eastward displacement of the zone of convection over the maritime continent to the central Pacific and a subsequent failure of the Southern Hemisphere Summer Monsoon (Webster *et al.*, 1998). In the Central and South American region convection over land was displaced westward of its normal position over Amazonia to northern Peru, and an enhanced subtropical jet stream blocked the southern polar front systems from bringing moisture to the tropics (Martin *et al.*, 1993).

The broad synchrony existing between these two regions is determined by large-scale atmospheric circulation systems, including (1) zones of convection and subsidence over each region, forming the Walker Circulation, and (2) redistribution of heat from the tropics to the poles via Haldley Circulation. The average zone of convection, or meteorological equator, over the continents is strongly dependent on the influence of land–sea distributions, land–sea temperature gradients, and cloud cover on large-scale atmospheric circulation (Philander *et al.*, 1996). While the seasonal movement of the meteorological equator determines the wet and dry seasons north and south of the

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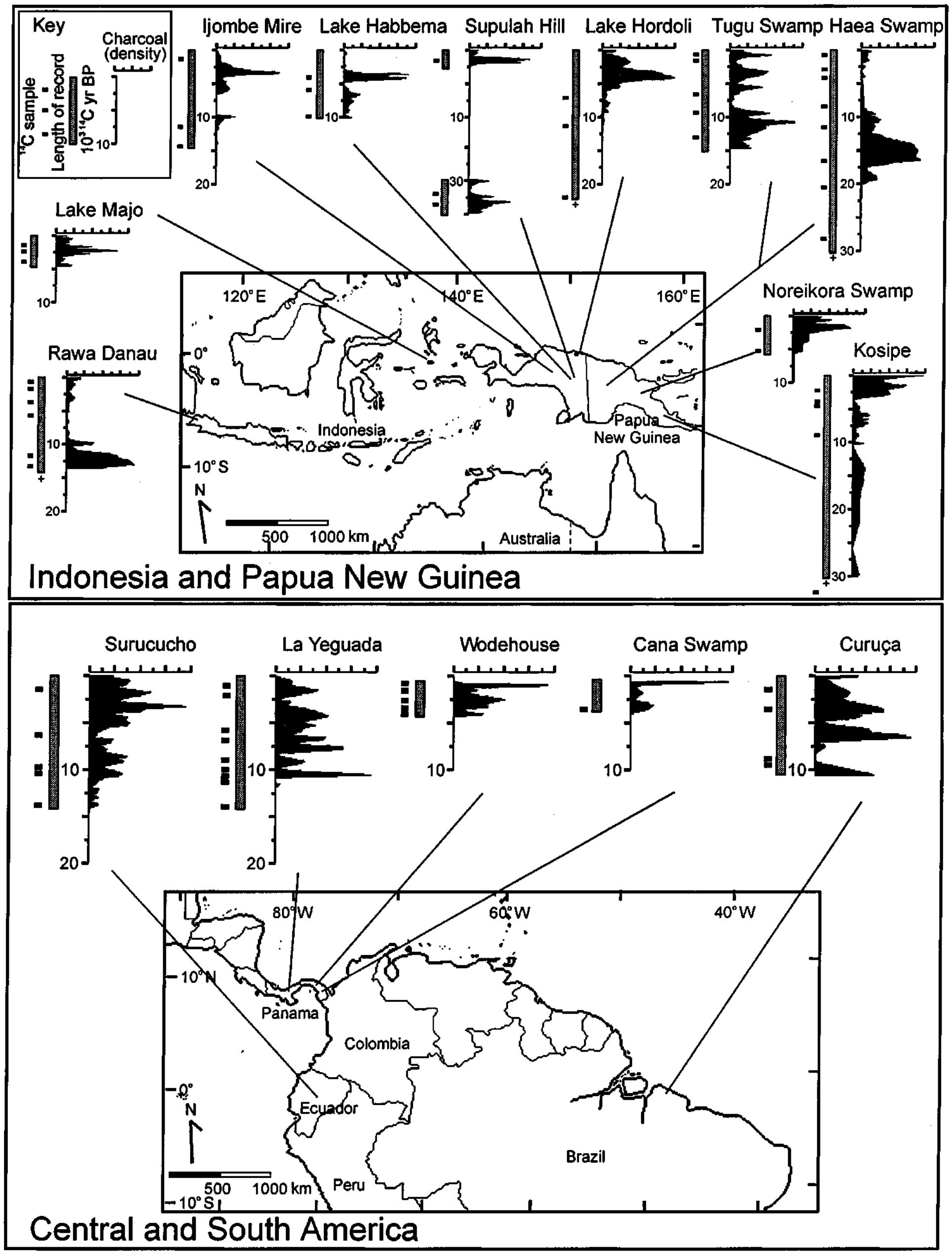

FIG. 1. Charcoal records from sites in Indonesia and Papua New Guinea (adapted from Haberle *et al.* (in press)) and in Central and South America. Standardized charcoal curves (dimensionless) are plotted against ^{14}C age with sample location shown in black squares (see Table 1 for details).

latitudinal equator, it is changes in the position and strength of the meteorological equator on century to millennial time scales that remains poorly understood.

Terrestrial paleoenvironmental data has increased considerably for the tropics within the last decade, and apart from the observation of a synchronous expansion of pollen taxa representing cool and moist vegetation between Africa and South America (Servant *et al.*, 1993), comparisons have been focused on interhemispheric transects (PEP I, Markgraf, 2000; PEP II, Rokosh *et al.*, 1997). In this paper we use a zonal intertropical comparison of long-term cumulative charcoal records from the Indonesian and Papua New Guinea region and Central and South American region (Fig. 1) to evaluate the role of climate and other factors, such as human activity, in fire occurrence through time.

METHODS

Microscopic charcoal preserved in soils and lake and swamp sediments has been used, in association with fossil pollen records, to reconstruct the occurrence of "regional fire periods" in the landscape. A relatively small number of the total paleoecological sites available from the tropics have detailed information on changes in the abundance of microscopic charcoal preserved in the sediment. A total of 10 sites in the Indonesian and Papua New Guinea region (mostly south of the equator) and 5 sites in the Central and northern South American region (mostly north of the equator) have been selected for this analysis on the basis of sampling resolution and chronological control (Table 1). Criteria for site selection include: (1) small basin area (<1 km^2), (2) sampling resolution of <250 yr/sample, (3) good chronological control, and (4) consistent particulate charcoal measurements. At each site detailed counts of microscopic particulate charcoal were expressed either as concentrations using the point counting method (Clark, 1982) or as a ratio of particles to pollen sum. Charcoal particle density values were standardized (unity) for each site. Using a linear age model and interpolation, a cumulative charcoal record was constructed by summing the 200-yr values for each site, and a corrected charcoal curve was calculated (cumulative charcoal per site) (Anderson and Smith, 1997; Haberle *et al.*, in press). Within each region, high corrected charcoal values are assumed to be related to periods of synchronous regional burning.

RESULTS AND DISCUSSION

The individual charcoal records from each region (Fig. 1) have been derived from a wide geographical spread covering different vegetation types and variable influence from human activity. Vegetation types vary from alpine grasslands to montane and lowland forests, all with marked differences in biomass and susceptibility to fire, producing a variable pattern of fire frequency across the landscape. A comparison between past charcoal records from regions with different human histories provides one of the most reliable means available of decoupling climate influences from human influences in the fossil record.

In the Indonesian and Papua New Guinea region the earliest sustained rise in charcoal values occurred at different times across the island of New Guinea (Supulah Hill and Haea Swamp, Fig. 1) and is considered to reflect the strong impact of initial human occupation in the region about 32,000 ^{14}C yr B.P. (Haberle, 1998; Hope, 1998), supported by archeological evidence for initial colonization (Swadling and Hope, 1992). Corresponding pollen records show that burning appears to precede the expansion of grassland at the expense of forest cover, suggesting that persistent burning may have forced forest retreat. Human occupation of the Americas occurred much later, possibly between 13,000 and 11,000 cal yr B.P. (Cooke, 1998), although in this case corresponding pollen records show that burning occurred at, or slightly after, the expansion of grasslands (Surucucho and La Yeguada, Fig. 1), suggesting that increased burning may have been a response to grassland expansion rather than a cause of this vegetation change. That subsequent increases in burning occurred in both regions at times of increasing agricultural activity highlights the difficulty in separating human from natural causes of biomass burning, even under a regional comparative approach. A more parsimonious interpretation is that human activity is strongly influenced by climate change and should not be treated as mutually exclusive.

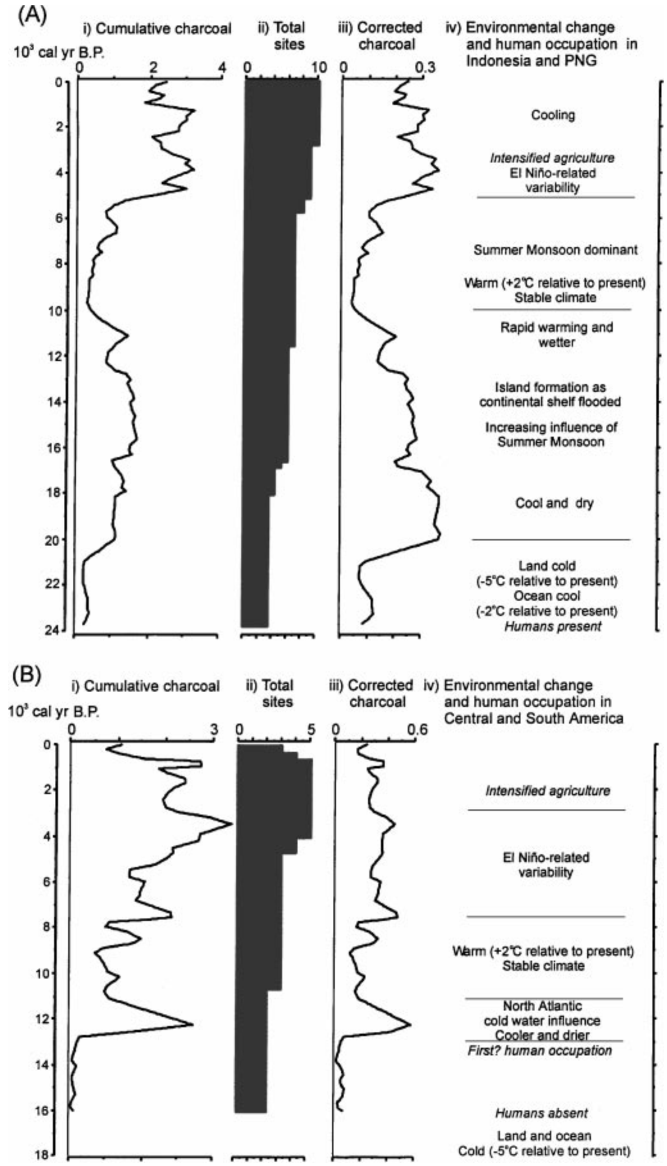

FIG. 2. (A) Cumulative charcoal record for the Indonesia and Papua New Guinea region adapted from Haberle *et al.* (in press). (i) Cumulative charcoal record constructed by summing the 200-yr values for each site. (ii) Number of cores with data for each 200-yr period. (iii) Value in (i) divided by the corresponding value in (ii). (iv) Summary of climate change and human occupation in the Indonesia and Papua New Guinea region (after Huang *et al.*, 1997; Flenley, 1998; Haberle, 1998). (B) (i) Cumulative charcoal record constructed by summing the 200-yr values for each site in the Central and South American region. (ii) Number of cores with data for each 200-yr period. (iii) Value in (i) divided by the corresponding value in (ii). (iv) Summary of climate change and human occupation in the Central and South American region (after Markgraf (2000)).

In general, periods of frequent regional burning in each region can be shown to correlate broadly with periods of rapid climate change or high climate variability, or prehistoric human activity (Figs. 2A and 2B). In the Indonesian and Papua New Guinea region, reduced biomass burning during the last glacial maximum (24,000–20,000 cal yr B.P.), at a time when people

TABLE 1
Location, Description, Radiocarbon Dates, and Calibrated Dates (Stuiver and Reimer, 1993) for Sites Used in This Analysis

Site	Site type, elevation, location	Depth (cm)	^{14}C yr B.P.	Calib. dates 1σ range (cal yr B.P.)	Reference
		Indonesian and Papua New Guinea region			
Kosipe	sedge swamp,	71–82	2070 ± 70	2120–1950	Hope (1980)
	1960 m,	110–125	3530 ± 80	3900–3700	
	8°28′S, 147°12′E, PNG	185–198	4190 ± 80	4840–4780	
		360[a]	9330 ± 250	11,060–10,200	
		540[a]	35,900 ± 850	—	
Noreikora Swamp	grass/sedge swamp,	340–350	1580 ± 160	1700–1300	Haberle (1996)
	1750 m,	500–510	4470 ± 120	5300–4870	
	6°20′S, 145°50′E, PNG				
Haeapugua	grass/sedge swamp,	70–78	890 ± 80	920–710	Haberle (1998)
	1650 m,	185–190	2860 ± 100	3160–2850	
	5°50′S, 142°47′E, PNG	290–300	4380 ± 80	5050–4850	
		344–350	8340 ± 150	9520–9090	
		420–430	11,270 ± 100	13,400–13,150	
		480–490	16,640 ± 250	20,250–19,410	
		495–505	16,990 ± 180	20,600–19,850	
		550–560	20,670 ± 150	—	
		642–650	27,760 ± 390	—	
Tugupugua	grass/sedge swamp,	60–70	880 ± 70	910–710	Haberle (1998)
	2300 m,	195–200	1700 ± 210	1870–1350	
	5°40′S, 142°35′E, PNG	400–415	5750 ± 80	6660–6410	
		600–615	9290 ± 190	10,690–10,240	
		800–815	12,860 ± 250	16,200–14,250	
Lake Hordorli	sedge swamp,	90–98	7140 ± 120	8100–7800	Hope (1996)
	780 m,	190–200	10,750 ± 120	12,950–12,650	
	2°32′S, 140°33′E, Indonesia	390–400	22,500 ± 240	—	
Supulah Hill	shallow lake,	150–170	1770 ± 70	1820–1570	Hope (1998)
	1580 m,	230–250	32,240 ± 880	—	
	4°7′S, 138°58′E, Indonesia	330–350	33,200 ± 1070	—	
Lake Habbema	lake,	215–225	3050 ± 80	3360–3080	Haberle *et al.* (in press)
	3120 m,	382–392	5610 ± 80	6470–6300	
	4°7′S, 138°42′E, Indonesia	635–645	9880 ± 130	11,550–11,200	
Ijombe	sedge pen,	310	6450 ± 100	7430–7270	Hope and Peterson (1976)
	3630 m,	480	10,750 ± 260	12,480–12,370	
	4°2′S, 137°13′E, Indonesia	1000	13,850 ± 260	17,000–16,250	
Lake Majo	lake margin,	220–240	1300 ± 90	1290–1150	Haberle *et al.* (in press)
	140 m,	414–432	220 ± 190	2350–1950	
	1°28′S, 127°29′E, Indonesia	634–648	4270 ± 90	4870–4660	
Rawa Danau	shallow lake,	270–280	350 ± 80	510–310	Van der Kaars *et al.* (in press)
	100 m,	380–390	1810 ± 60	1820–1630	
	6°11′S, 105°58′E, Indonesia	490–500	3890 ± 160	4450–4090	
		860–870	5450 ± 150	6400–6000	
		1400–1410	11,460 ± 160	13,790–13,170	
		1640–1650	13,200 ± 240	16,220–15,500	
		Central and South American region			
Cana Swamp	grass swamp,	925–940	3680 ± 110	4150–3850	Bush and Colinvaux (1994)
	≈500 m,				
	7°45′N, 77°35′W, Panama				
Wodehouse Swamp	grass swamp,	162–175	370 ± 50	500–320	Bush and Colinvaux (1994)
	≈500 m,	435–445	1920 ± 140	2000–1700	
	7°45′N, 77°35′W, Panama	685–700	2600 ± 100	2800–2500	
		830–840	3390 ± 100	3800–3500	
		1030–1040	3980 ± 80	4550–4300	

TABLE 1—*Continued*

Site	Site type, elevation, location	Depth (cm)	^{14}C yr B.P.	Calib. dates 1σ range (cal yr B.P.)	Reference
		Central and South American region			
La Yeguada	shallow lake,	144–160	600 ± 110	650–500	Bush *et al.* (1992)
	650 m,	335–345	2320 ± 70	2350–2200	
	8°27′N, 80°51′W, Panama	740–760	5700 ± 100	6600–6350	
		867–877	6790 ± 130	7750–7500	
		980–995	8840 ± 130	10,200–9600	
		1030–1050	10210 ± 130	12,300–11,600	
		1140–1150	10530 ± 100	12,800–12,350	
		1200–1210	11250 ± 140	13,400–13,000	
		1310–1320	11610 ± 180	13,800–13,400	
		1630–1640	13670 ± 210	16,750–16,050	
Lake Curuça	shallow lake,	25–30	1520 ± 70	1500–1300	Behling (1996)
	35 m,	42–47	2740 ± 60	2900–2800	
	0°46′S, 47°51′W, Brazil	72–77	9340 ± 60	10,700–10,400	
		90–97	9430 ± 70	10,750–10,550	
Lake Surucucho	lake	88–100	1720 ± 70	1700–1550	Colinvaux *et al.* (1997)
	3180 m,	402–415	6500 ± 80	7450–7350	
	3°3′45′S, ca. 78°W, Ecuador	580–595	10330 ± 140	12,600–11,800	
		642–653	10000 ± 140	11,900–11,200	
		730–745	13070 ± 270	16,100–15,300	

[a] AMS dates.

had already occupied the region, may have been a function of low land temperatures, as evidenced from vegetation histories (Flenley, 1998; Haberle, 1998) and glacial records (Hope and Peterson, 1976). However, there is sufficient discrepancy between estimates of mean annual temperature from sea-surface records and land-based records (Rind and Peteet, 1985) that a causal link between lower temperatures and low biomass burning is uncertain. Between 20,000 and 10,000 cal yr B.P. burning appears to have been exacerbated by an unstable convection regime brought about by major environmental influences: (1) rising sea level and rapidly decreasing continentality, and (2) the increasing strength of the Southern Hemisphere Summer Monsoon (Huang *et al.*, 1997). The early Holocene, between 10,000 and 5000 cal yr B.P., is characterized by very infrequent regional burning at a time when high mean annual temperatures (+2°C relative to present; Nix and Kalma, 1972), combined with reduced summer insolation and increased winter insolation, may have decreased the climate variability. Subsequent increases in charcoal occurred only after 5000 cal yr B.P. and appear to be linked to the interacting forces of intensified El Niño-related climate variability (McGlone *et al.*, 1992) and agricultural intensification (Haberle, 1998).

In the Central and South American region the late-glacial period is characterized by infrequent regional burning that persists until about 12,500 cal yr B.P., when the highest corrected charcoal values are recorded prior to the beginning of the Holocene. The combined impact of rapid climate change during the late-glacial transition and the arrival of people on the landscape may have contributed to increased biomass burning between 12,500 and 11,000 cal yr B.P. (Fig. 2B). The strong correspondence between high charcoal values and the Younger Dryas Stade is suggestive of a regional climate change conducive to increased biomass burning. High-resolution paleoclimate records from pollen (Islebe *et al.*, 1995; Van der Hamman and Hoogheimstra, 1995) and ice cores (Thompson *et al.*, 1995) lend some support to the notion of a rapid cooling, and possibly drying, event during the Younger Dryas Stade in Central America and the northern South American region. However, the close correspondence between human arrival and high climate variability associated with a Younger Dryas Stade influence would likely have led to an exacerbation of biomass burning in this region, so the separation of these two influences becomes problematic.

Corrected charcoal values equivalent to the late-glacial period values occur during the earliest part of the Holocene, between 11,000 and 7800 cal yr B.P. The climate had shifted to warmer and wetter conditions and in conjunction with reduced seasonality (Martin *et al.*, 1997) appears to have been sufficient to reduce biomass burning, despite the presence of people in the landscape. The subsequent increase in regional burning, most notable between 7800 and 3000 cal yr B.P., corresponds to paleoclimate records from pollen cores (Absy *et al.*, 1991), debris-flow deposits (Rodbell *et al.*, 1999), and beach-ridge terraces (Martin *et al.*, 1993) that suggest the climate became drier and more variable. Martin *et al.* (1993) relate this to the onset of El Niño-related variability.

Comparison of the cumulative charcoal records demonstrate a persistent, though modulating, climate connection between the Indonesia, Papua New Guinea, and the Central and South

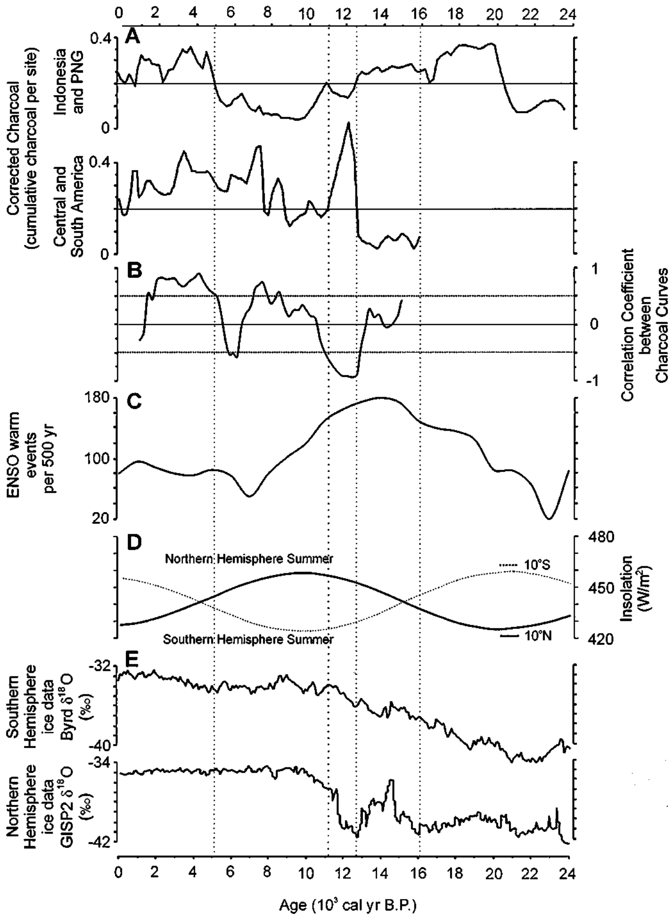

FIG. 3. Comparison among (A) corrected charcoal records (cumulative charcoal per site), (B) correlation between these records (moving 2000-yr interval *r* value), (C) $\delta^{18}O$ ice-core data from the Northern Hemisphere (GISP2; Grootes *et al.*, 1993) and the Southern Hemisphere (Byrd; Broecker, 1998), (D) summer insolation at low latitudes (10°S and 10°N; Berger and Loutre, 1991), and (E) number of El Niño events per 500 yr in an orbitally forced model of the tropical Pacific (Clements *et al.*, 1999).

American tropics (Figs. 3A, 3B). The broad synchrony of changes in corrected charcoal values in each region supports an atmospheric transmission of the climate signal via the dominant large-scale atmospheric circulation systems: (1) Walker Circulation across the tropics that creates zones of convection and subsidence over each region, and (2) Hadley Circulation that redistributes heat from the tropics to the poles. However, fluctuations in the strength of these broad-scale connections, as revealed using a moving 2000-yr interval correlation coefficient calculated between these two records (Fig. 3B), may be explained by altered climate boundary conditions under the influence of changing El Niño-related variability, insolation, sea level, and sea-surface temperature (Figs. 3C–3E).

There is a general trend of increasing correlation from 16,000 cal yr B.P. to the late Holocene, with periods of inverse correlation occurring during the Younger Dryas Stade and between 7000 and 5500 cal yr B.P. Weakened or low-frequency El Niño-related variability in the Central and South American region during the late-glacial to early Holocene period (Rodbell *et al.*, 1999) is at odds with recent paleoclimatic modeling experiments (Clements *et al.*, 1999) that suggest the amplitude and frequency of El Niño-related climate variability, influenced by solar forcing, was high during this period (Fig. 3C). However, there is a strong correspondence between reconstructed El Niño frequency and biomass burning recorded over the last 24,000 yr in the Indonesian and Papua New Guinea region. That the same cannot be said for biomass burning in the Central and South American region during the late-glacial period implies that climate boundary conditions had altered, possibly under the enhanced influence of relatively cool equatorial Atlantic and eastern Pacific waters (Hostetler and Mix, 1999). Alternatively, the weak connection between the two regions may relate to differences in the geometry of the continental shelf, where rising sea level transformed the exposed shelf in the Indonesian and Papua New Guinea region into a mosaic of islands and small ocean basins, creating a complex and rapidly changing meteorological equator. This may have been a sufficient destabilizing force on local climate to generate a high regional burning regime compared to the Central and South American region, where flooding of the continental shelf did not alter the geometry of the coastline to the same extent as in the Indonesian and Papua New Guinea region.

The most significant period of inverse correlation occurs during the Younger Dryas Stade, when enhanced extratropical North Atlantic Ocean influence on Central and South American climate patterns appears to be reflected in the western Pacific by a slightly reduced biomass burning signal. However, supporting evidence for Younger Dryas Stade cooling in the western Pacific region remains scarce (Peteet, 1995).

Holocene palynological data show that vegetation communities adapted to wet and warm climates were established in both regions (Haberle, 1998; Markgraf, 2000). Insolation data (precessional) suggests that low latitudes during the early Holocene were characterized by reduced seasonality, with relatively warmer Southern Hemisphere winters and cooler Southern Hemispher summers and a much more restricted meteorological equator covering both Northern and Southern hemispheres (Martin *et al.*, 1997). After 7800 cal yr B.P., the cumulative fire signal increases and appears to have become synchronous in the two regions, especially after 5000 cal yr B.P. when the low latitudes came under the influence of intensified El Niño activity (Sandweiss *et al.*, 1996; Clements *et al.*, 1999; Rodbell *et al.*, 1999). This is further supported by pollen records showing increased vegetation distrubance across the Pacific Basin (McGlone *et al.*, 1992) after 5000 cal yr B.P. Whether the short-term shift to inverse correlation between the two regions from 7000 to 5500 cal yr B.P. and after 1500 cal yr B.P. corresponds to periodic enhanced extratropical influences on the tropical climate patterns remains unclear.

Records of past regional fire activity appear to reflect most strongly past global climate change, despite different vegetation communities or human influences on a region. The broad synchrony and abruptness of changes in the two cumulative fire records support a strong atmospheric transmission of the climate signal throughout the tropics. That short-term fluctuations in climate are an important determinant of species distributions has been shown by recent ecological studies demonstrating that population-level dynamics are affected by annual to decadal scale climate change (Phillips and Gentry, 1994). Forest dieback due to drought stress and fire, or the patchy destruction of canopy trees after cyclone events, provide evolutionary possibilities because they create gaps, speeding up regeneration dynamics and leading to a maintenance of high biodiversity. Periods of high climate variability, such as 5100 cal yr B.P. to the present, may have been important driving forces for promoting fire in rain forest environments and maintaining diversity of tropical rain forests.

The number of records of charcoal abundance from the tropics remains very sparse. Additional data no doubt will reveal new trends in temporal and spatial patterns of biomass burning, although the current data present a strong argument that tropical forests have been subject to fluctuating levels of burning through time. These fluctuations in tropical biomass burning are at least partly controlled by orbital forcing (precession), although extra-tropical climate influences and human activity are also important causal elements.

ACKNOWLEDGMENTS

We thank Peter Kershaw, Geoff Hope, Vera Markgraf, and an anonymous referee for valuable comments on earlier drafts of the manuscript. The paper was written while the first author was supported by an Australian Research Council QEII Research Fellowship.

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