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# A biomass burning record from the West Equatorial Pacific over the last 360 ky: methodological, climatic and anthropic implications

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## Abstract

To reconstruct past fire activity in the West Pacific–East Asian region, we investigate a 360-ky pelagic sedimentary record from the Caroline Basin in the Western Pacific Warm Pool (WPWP) (core MD97-2140, 2547 m. water depth). In the lack of standard protocol and terminology for the determination of combustion-derived products in sedimentary archives, three proxies were compared with other paleoceanographic proxies: Oxidation Resistant Elemental Carbon (OREC), Black Carbon (BC), and microcharcoal. The mineral ( $\text{CaCO}_3$ ) and organic carbon (Corg) records of core MD97-2140 generally covary with the oxygen isotope record, suggesting that glacial/interglacial changes in ice-volume and sea-level control the preservation of  $\text{CaCO}_3$  and terrigenous inputs of Corg in the deep-sea sediments of the Caroline Basin. The changes in OREC are primarily connected to the changes in Corg record. In contrast, the BC and charcoal records, which better reflect the input of biomass burning products, are tightly connected with changes in the precession band (23 ky), likely through low-latitude atmospheric circulation patterns. A peculiar control by the competing influence of the long-term El Niño–Southern Oscillation (ENSO)-like forcing and the glacial/interglacial cycle on the East Asian summer monsoon (30-, 19-, 11- and 6-ky periods) is suggested. In addition, large increases in BC and charcoal are observed between ca. 53–43 and 12–10 ky BP. These events strikingly correspond to the main Late Pleistocene and Early Holocene periods of human colonization in the region, likely suggesting an additional anthropic impact on the fire activity.

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## 1. Introduction

Biomass burning in the tropics releases today about one third of annual anthropogenic  $\text{CO}_2$  emissions and large amounts of aerosol particles, making an impor-

tant contribution to atmospheric chemistry and climate (Seiler and Crutzen, 1980; Crutzen and Andreae, 1990; Hao et al., 1990). Despite such major environmental effects, the dynamics of tropical fires, however, are still poorly documented and understood for the past, especially at the scale of global, glacial–interglacial climate cycles (Verardo and Ruddiman, 1996; Bird and Cali, 1998). The aim of this paper is to provide a Late

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Pleistocene multiproxy record of biomass burning from the West Pacific–East Asian region, a fire-sensitive region where present-day devastating forest fires are triggered and sustained by the combination of landuse practices, and regional changes in length and intensity of the dry season, in association with the El Niño–Southern Oscillation (ENSO) (Goldamer and Seibert, 1990; Goldamer, 1997).

The plant-derived material partially altered and blackened by fire is generally referred as charred by microscopic or chemical characteristics (Jones et al., 1997): (i) Small charcoal particles are formed in fuel by low-temperature pyrolysis processes, and carried out with smoke emissions. (ii) Soot particles (sub-micrometer or up to less than 1- $\mu\text{m}$  size) are emitted with smoke and form via gas-phase processes, and gas-to-particle conversion. (iii) During a fire at high temperature and intense flaming combustion, more carbon is reduced nearly its elemental state and referred as Black Carbon (BC) (Cofer et al., 1997). This fire-originating aromatic to graphitic carbon particulate fraction (0.01–1- $\mu\text{m}$  size range) is primarily produced in fire-altered material or smoke particulate (Penner et al., 1992; Jacobson, 2001).

To date, most marine sedimentary records of past biomass burning have been reconstructed by using physical analysis (microcharcoal counting) and/or chemical analysis (analysis of Oxidation Resistant Elemental Carbon or OREC, and Black Carbon or BC) of carbonaceous products (Bird and Cali, 1998; Gustafsson and Gschwend, 1998; Moss and Kershaw, 2000). Given the complex sources and behavior of carbonaceous material in sedimentary environments, and the lack of intercalibration between the existing methods (Schmidt and Noack, 2000), we combined OREC analysis, BC analysis, and charcoal counting on the same cored sequence (MD97-2140; 360 ky) to improve the reliability of biomass burning reconstruction.

## 2. General setting

### 2.1. Present-day oceanography and climate in the West Pacific warm pool area

The WPWP occupies the East Indian–West Pacific areas and is centered on the Equator, northeast of

Papua New Guinea, close to the MD97-2140 coring site (Fig. 1). The WPWP region contains the warmest surface waters of the global ocean. At annual scale, the precipitation pattern over the region is tied to the seasonal shift of the Intertropical Convergence Zone (ITCZ) and the trans-equatorial cold surge of the northern winter monsoon (An, 2000). The resulting Australian summer monsoon and summer rainfall affect an area which includes Indonesia, Papua New Guinea, and northern Australia. In the central highlands of Papua New Guinea, a double rainfall maximum occurs around March and September/October of each year (Haberle et al., 1998), approximately in phase with the two insolation maxima at the Equator (spring and autumn equinoxes). Previous regional studies have shown that biomass burning in the region is closely controlled by the seasonal distribution of rains, and by its orbital (precession) forcing (Haberle and Ledru, 2001).

At interannual scale, zonal changes of the WPWP associated with ENSO-like oscillations cause dramatic changes in duration and regional extent of the dry season: at the onset of “warm” ENSO events (El Niño), the WPWP surface waters are driven eastward by the near-equatorial westerly winds associated to the East Asian winter monsoon. Consequently, atmospheric convection and rainfall dramatically weaken above the tropical rainforests of the East Indian–West Pacific area. The recent massive fires that occurred in Indonesia were caused by the association between such dry, ENSO-driven conditions, and human forest clearing and occupation. It has thus been suggested that similar dramatic events possibly occurred during the Late Pleistocene (Goldamer, 1993, 1999; Haberle and Ledru, 2001; Siebert et al., 2001).

### 2.2. Depositional setting of core MD97-2140

The 37.4-m-long core MD97-2140 (02°04' N, 141°76' E; 2547-m water depth) was collected on the southwestern slope of the NNE–SSW trending Eauripik ridge, during the IPHIS-IMAGE III cruise of the R/V Marion Dufresne, in 1997 (Fig. 1). The coring site is located ca. 400 km North off Papua New Guinea in the Caroline Basin. Because it is separated from the continent by the New Guinea trench, the coring site is relatively isolated from the

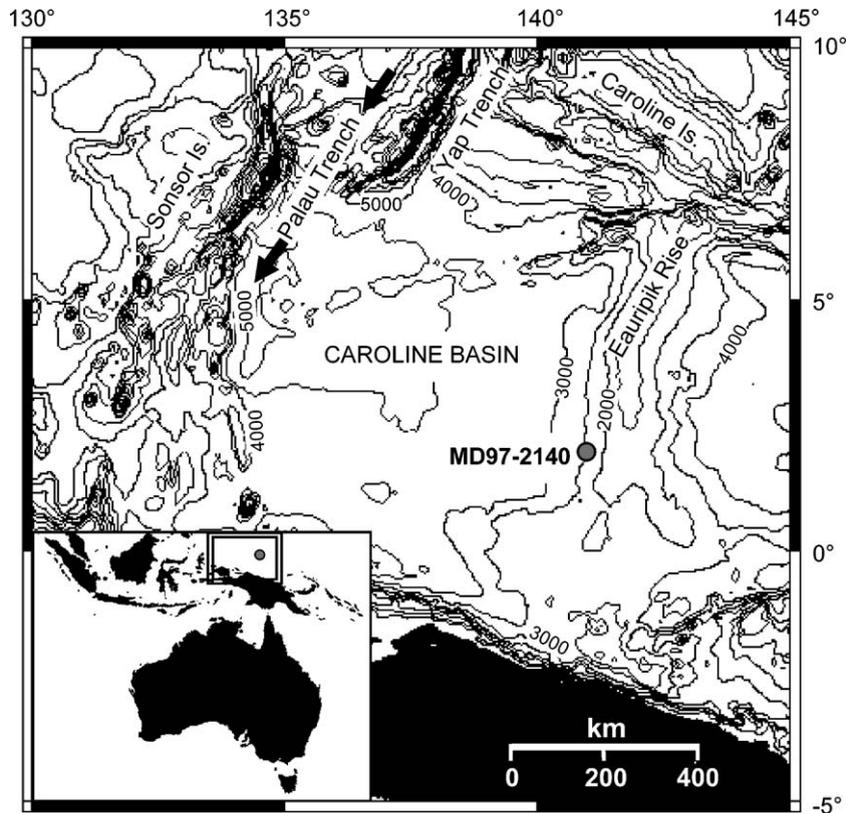


Fig. 1. Regional map of the Papua New Guinea and Northern–Australian region, and bathymetric map of the West Caroline Basin showing the location of the core MD97-2140 at 2°04' N, 141°76' E, 2547-m depth. The dark arrow represents the predicted primary flow path of the Pacific bottom and deep waters into the West Caroline Basin (modified from Kawahata et al., 1997).

influence of terrigenous inputs from Papua New Guinea, which deposit in the deep hemipelagic region through a very narrow shelf (Woolfe and Lecombe, 1998).

The present lysocline is at ca. 3400-m water depth in the western equatorial Pacific (Groetsch et al., 1991). The Caroline Basin (maximum depth: ca. 5000 m) is isolated from the western North Pacific by a rise including the Sonsorol and Caroline Islands, and the Eauripik Rise (Fig. 1). This basin is located at the western end of the flow path of bottom waters from the equatorial Pacific. Below ca. 3000-m water depth, poor preservation of carbonates in the western Caroline Basin has been documented and may be caused by a strengthened alkalinity and carbon dioxide enrichment of bottom waters, in response to the isolation of the Caroline basin (Kawahata et al., 1997).

### 3. Lithology, stratigraphy and sampling

The core MD97-2140 sediment consists of bioturbated, olive-gray nannofossil ooze. Like most pelagic carbonate sequences from the tropics, low amplitude changes in volume magnetic susceptibility  $\chi$  (shipboard measurements) indicate a cyclic dilution of the carbonates by dominantly paramagnetic clays (Yamazaki and Loka, 1997).

A well-constrained oxygen isotope chronostratigraphy of core MD97-2140 has been established from the  $\delta^{18}\text{O}$  record of *Globigerinoides ruber* (planktonic foraminifera) by de Garidel-Thoron et al. (manuscript in preparation). For the upper part of the core (i.e. above the isotope stage 3–2 transition), the depth–age model was obtained by correlation with the radiocarbon-dated *G. ruber* isotopic record of the neighboring core MD97-2138. Correlation

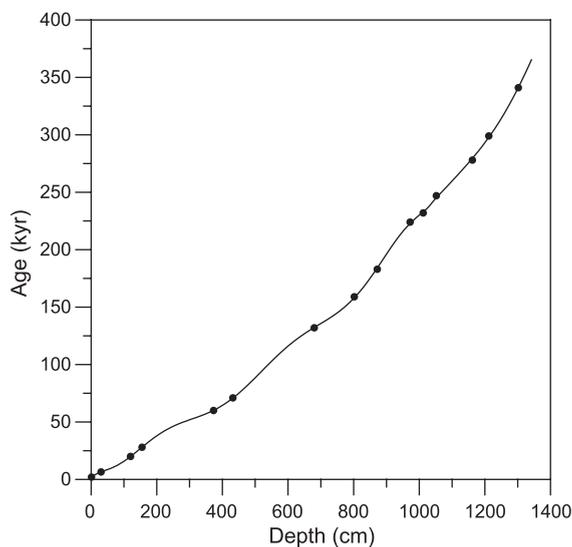


Fig. 2. Age–depth relationship in core MD97-2140.

with the SPECMAP stack (Imbrie et al., 1984) was used for the bottom part of the record. The resulting depth–age curve (Fig. 2) locates the upper 14 m of the core in the last 360 ky, from Marine Isotope Stage (MIS) 10.

For the purpose of this study, the upper 13.5 m of the core were sub-sampled at 5-cm depth-interval and stored at  $-20\text{ }^{\circ}\text{C}$  until analysis in CEREGE laboratories. According to the variable apparent deposition rate, the time sampling interval of our record lies between  $\sim 0.8$  and  $-3$  ky.

## 4. Methods

### 4.1. Microscopic charcoal counting

The method used for the automated counting of microcharcoal particles as isolated by the image analysis of microscopic slides is described by Thevenon et al. (2003a). The treatment was performed on 50–60 mg dry sediment. The results were expressed in charcoal area per gram of sediment and converted in Charcoal Accumulation Rate (CHAR):

$$\text{CHAR} = \text{charcoal area} \times \text{sedimentation rate} \times \text{density}$$

CHAR is expressed in  $\text{mm}^2 \text{cm}^{-2} \text{ky}^{-1}$ , charcoal area in  $\text{mm}^2 \text{g}^{-1}$ , sedimentation rate in  $\text{cm ky}^{-1}$ , density in  $\text{g cm}^{-3}$ .

In the absence of dry density measurements, we used the wet bulk density (WBD) in the calculation of accumulation rates.

### 4.2. Organic carbon (Corg) and calcium carbonate ( $\text{CaCO}_3$ ) analyses

The analysis of organic and inorganic carbon (calcium carbonate) was also performed to reconstruct the general variability of deposition at the coring site, and to ensure that the determination of the combustion-derived products was not biased by these carbonaceous components.

The samples were lyophilized and ground into a fine powder with an agate mortar. Carbon measurements were performed by using the automatic Na-1500 Elemental Analyzer. An overall reaction scheme is presented in Fig. 3. Total carbon (Ctotal) was measured on ca. 10 mg of bulk sediment within Sn caps ( $3 \times 5$  mm). Organic carbon (Corg) was determined on ca. 15 mg samples, after removing carbonate by acidification (2 N HCl) in Ag caps

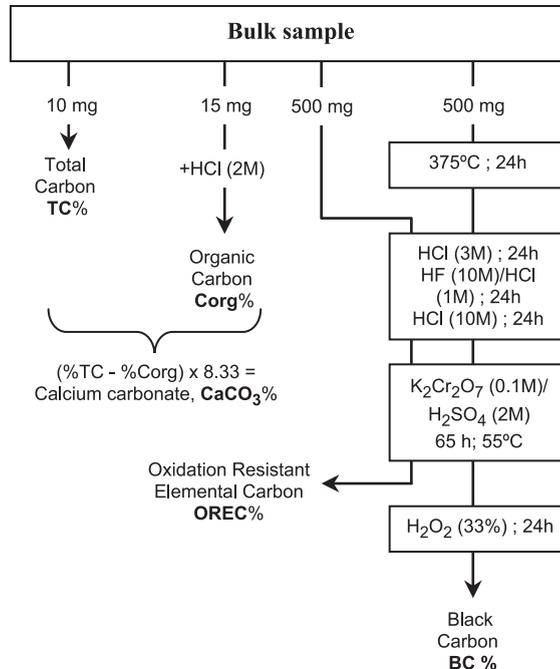


Fig. 3. The sequence reaction scheme of the various elemental-carbon analyses: Total Carbon (TC), Organic Carbon (Corg), Calcium Carbonate ( $\text{CaCO}_3$ ), Oxidation Resistant Elemental Carbon (OREC), and Black Carbon (BC).

(3×5 mm). Ctotal and Corg experimental blank values did not differ significantly, yielding limit of quantification ( $L_Q$ ) of ca. 1.6 and 4.3  $\mu\text{g}$  carbon (10 times the standard deviation of the mean blank value; Currie, 1968), and relative  $L_Q$  of 0.016% and 0.029%, respectively. The calcium carbonate ( $\text{CaCO}_3$ ) concentration was then calculated as the difference between Ctotal and Corg contents (Verardo et al., 1990).

#### 4.3. Oxidation Resistant Elemental Carbon (OREC) and Black Carbon (BC) analyses

Two methods (OREC and BC extractions) were tested to remove mineral and organic carbon compounds, and to concentrate the elemental refractory carbon components (Fig. 3). The OREC method has been previously applied to extract the elemental carbon (soot, charcoal) in sample of the Cretaceous–Tertiary boundary clay (Wolbach and Anders, 1989). However, Bird and Gröcke (1997) and Bird et al. (1999) suggested that an additional peroxide or thermal oxidation step can improve the removal of any remaining refractory organic carbon.

Oxidation Resistant Elemental Carbon (OREC) extraction and measurement was performed on ca. 500 mg bulk sediment samples by using the method of Wolbach and Anders (1989). The samples were placed into centrifugation tubes, and first treated by three treatments with 3 M HCl and 10 M HF/1 M HCl, and 10 M HCl, respectively. The following step consisted of oxidation during 65 h into a sulfo-dichromate solution maintained at 55 °C (0.1 M  $\text{K}_2\text{Cr}_2\text{O}_7$  in a 2 M solution of  $\text{H}_2\text{SO}_4$ ), which removed organic compounds. Finally, the remaining insoluble residue was filtered with a vacuum pump on a pre-cleaned silica filter (heated at 550 °C for 10 h) and the resulting OREC was measured by elemental analysis within big Sn caps (10×10 mm).

The BC extraction was also performed on ca. 500 mg bulk sediment samples. The samples were first treated by thermal oxidation (375 °C, 24 h) before chemical oxidation in sulfo-dichromate (OREC extraction). Finally, a peroxide treatment (33%  $\text{H}_2\text{O}_2$ , 24 h) removed any remaining refractory carbon (Bird and Gröcke, 1997). Although the  $L_Q$  of OREC or BC measurements (ca. 7.8  $\mu\text{g}$  carbon) was slightly higher than for Ctotal and Corg measurements, the treatment of ca. 500 mg samples allowed the applicability of the

method to marine sediments (relative  $L_Q$  of ca. 0.001%). To further compare with CHAR, the BC concentration was converted to Mass Accumulation Rate (MAR) by using the following equation:

$$\text{MAR BC} = 1000 \times \text{concentration} \\ \times \text{sedimentation rate} \times \text{density}$$

MAR is expressed in  $\text{mg cm}^{-2} \text{ky}^{-1}$ , concentration in wt.%, sedimentation rate in  $\text{cm ky}^{-1}$ , density in  $\text{g cm}^{-3}$ .

The BC method has been tested on experimental matrices of known composition. In order to reduce BC loss from these synthetic samples (during the rinsing as performed between each steps of the procedure), the dissolution of silica with HF was not conducted, and the carbonate fraction was dissolved just before the filtering phase, with 3 M HCl. The results are presented in detail in Table 1. We used silica and  $\text{CaCO}_3$  powders,  $\pm$  acetanilide (as a reference material of organic matter), oceanic (core MD97-2140) and lacustrine (core MM8) sediments, in which we add some graphite (synthetic or lithogenic). The correlation between the graphite content (wt.%) and the BC measured (%BC) is remarkable ( $r^2=0.99$ ,  $n=26$ ) (Fig. 4). The linear relation ( $Y=0.83X-0.008$ ) indicates that inorganic and organic carbon were completely removed by the procedure. The mean recovery ( $100 \times \text{BC measured} / \text{BC theoretical}$ ) with this method accounts for ca. 83% (S.D.=9;  $n=26$ ). Independent tests have shown that the experimental loss of BC mostly results from the oxidation by dichromate (ca. 2%) and the filtration technique (ca. 6%). However, the applicability of the BC extraction to clay-poor sediments should be undertaken with caution, since clays most likely retain BC during rinsing and filtration steps.

## 5. Results

### 5.1. Organic, mineral, and oxidation resistant carbon records

The  $\text{CaCO}_3$  and Corg records (Fig. 5) of core MD97-2140 show opposing trends over the whole record ( $r=-0.74$  for the entire data set), with  $\text{CaCO}_3$  values ranging between 40% and 75%, and Corg values ranging between 0.15% and 0.7%, respectively. Such a

Table 1

The results of the thermo/chemical oxidative treatment (BC extraction) on synthetic matrices and sediment of known composition (wt.%): Graphite synthetic (Gte stc, CAS No. 7782-42-5; TC=99.46%) or lithogenic (Gte lito, Alibert Mine, Siberia; TC=99.25%) are introduced in experimental matrices composed of calcium carbonate powder (CaCO<sub>3</sub>; CAS No. 471-34-1), silica powder (SiO<sub>2</sub>; CAS No. 7699-41-4),  $\pm$ acetanilide (Acet, SRM No. 141c, TC=71.21%)

Gte stc 1.269%, CaCO <sub>3</sub> 79.65%, SiO <sub>2</sub> 19.08%			
Mass (mg)			%BC
Sample	BC <sub>th</sub>	BC <sub>m</sub>	
205.61	2.615	2.154	<b>1.048</b>
220.27	2.802	2.438	<b>1.107</b>
216.62	2.755	2.431	<b>1.122</b>
<i>n</i> =3; mean (S.D.)			<b>1.092</b> (0.039)

Gte stc 1.250%, CaCO<sub>3</sub> 78.97%, SiO<sub>2</sub> 14.84%, Acet 4.93%

Mass (mg)				%BC
Sample	BC <sub>th</sub>	BC <sub>m</sub>		
199.84	2.486	2.022	<b>1.012</b>	
174.18	2.167	1.854	<b>1.064</b>	
161.03	2.003	1.595	<b>0.991</b>	
193.68	2.409	1.961	<b>1.013</b>	
<i>n</i> =4; mean (S.D.)			<b>1.020</b> (0.031)	

Gte stc 0.423%, CaCO<sub>3</sub> 89.50%, SiO<sub>2</sub> 10.08%

Mass (mg)				%BC
Sample	BC <sub>th</sub>	BC <sub>m</sub>		
229.78	0.974	0.705	<b>0.307</b>	
267.76	1.135	0.823	<b>0.307</b>	
<i>n</i> =2; mean			<b>0.307</b>	

Gte stc 0.466%, CaCO<sub>3</sub> 87.05%, SiO<sub>2</sub> 9.87%, Acet 2.61%

Mass (mg)				%BC
Sample	BC <sub>th</sub>	BC <sub>m</sub>		
235.99	1.102	0.889	<b>0.377</b>	
246.21	1.150	0.886	<b>0.360</b>	
243.31	1.136	0.893	<b>0.367</b>	
<i>n</i> =3; mean (S.D.)			<b>0.368</b> (0.008)	

Gte lito 0.735%, CaCO<sub>3</sub> 88.95%, SiO<sub>2</sub> 10.3%

Mass (mg)				%BC
Sample	BC <sub>th</sub>	BC <sub>m</sub>		
321.49	2.369	1.964	<b>0.611</b>	
325.42	2.398	1.906	<b>0.586</b>	
<i>n</i> =2; mean (S.D.)			<b>0.598</b>	

Table 1 (continued)

Gte lito 0.112%, CaCO <sub>3</sub> 94.72%, SiO <sub>2</sub> 5.16%				
Mass (mg)			%BC	
Sample	BC <sub>th</sub>	BC <sub>m</sub>		
514.27	0.591	0.442	<b>0.086</b>	
429.59	0.494	0.316	<b>0.074</b>	
485.87	0.559	0.474	<b>0.097</b>	
510.77	0.587	0.451	<b>0.088</b>	
<i>n</i> =4; mean (S.D.)			<b>0.086</b> (0.010)	

Gte lito 0.177%, SiO<sub>2</sub> 4.97%,  
MM8 sediment 94.85%, BC<sub>i</sub>=0.023%

Mass (mg)			%BC	BC–BC <sub>i</sub> %
Sample	BC <sub>lito</sub>	BC <sub>m</sub>		
445.89	0.789	0.682	0.153	0.130
484.32	0.857	0.768	0.159	0.136
545.58	0.965	0.873	0.160	0.137
475.96	0.842	0.907	0.191	0.168
<i>n</i> =4; mean (S.D.)			<b>0.166</b> (0.017)	0.143 (0.017)

Gte lito 0.113%, SiO<sub>2</sub> 4.79%,  
MD97-2140 sediment 95.09%, BC<sub>i</sub>=0.016%

Mass (mg)			%BC	BC–BC <sub>i</sub> %
Sample	BC <sub>lito</sub>	BC <sub>m</sub>		
516.44	0.580	0.694	0.134	0.118
451.82	0.508	0.586	0.130	0.114
538.43	0.605	0.534	0.099	0.083
437.64	0.492	0.556	0.127	0.111
<i>n</i> =4; mean (S.D.)			<b>0.123</b> (0.016)	0.107 (0.016)

Mass (in mg) for samples, BC theoretical (BC<sub>th</sub>), and BC measured (BC<sub>m</sub>). The corresponding BC content (%BC=100×BC<sub>m</sub>/Sample mass) was calculated. Number of replicate analyses (*n*), mean carbon content (in bold characters), and standard deviation (S.D.). To calculate the recovery within the experiments on sediments (MM8 and MD97-2140), the intrinsic BC concentration (BC<sub>i</sub>%) was subtracted from the measured BC concentration (%BC).

pattern strongly contrasts with neighboring, shallower sediments of the Eauripik ridge where high and constant carbonate concentrations (ca. 80%) exclude significant effects of carbonate dissolution (Kawahata and Eguchi, 1996; Kawahata, 1999). As observed at deeper water depth (Kawahata et al., 1997), the negative correlation between Corg and CaCO<sub>3</sub> is consistent with the combined effects of (i) dilution of organic matter by carbonate under variable productivity conditions, and (ii) dissolution of carbonate in oxygen-deficient conditions. Comparisons of Mass Accumulation Rate (MAR) (Thevenon, 2003b) suggest that changes in carbonate preservation (West Caroline

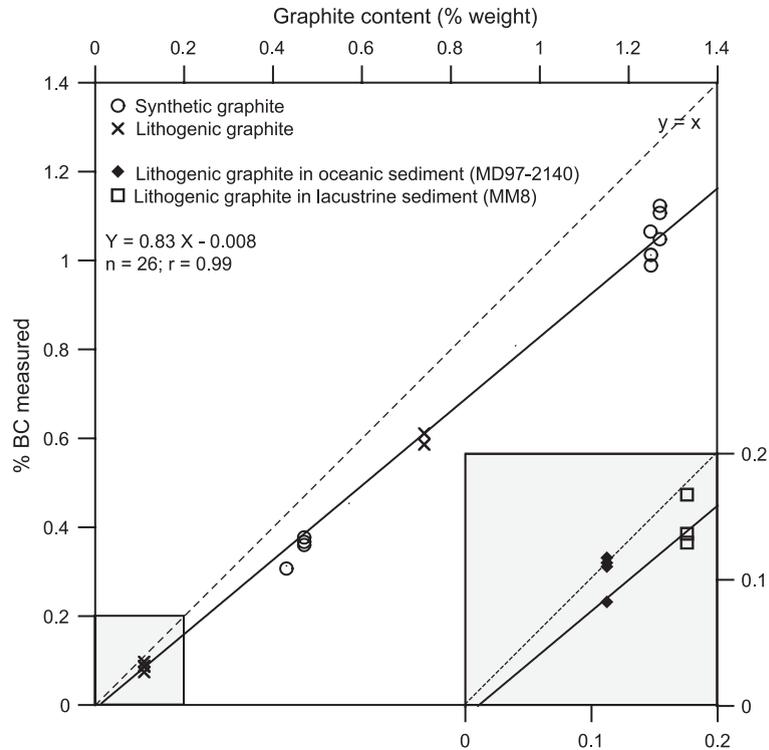


Fig. 4. %BC measured as a function of graphite content (wt.%) in synthetic matrices: Synthetic graphite (circles), lithogenic graphite (crosses), lithogenic graphite in marine sediment (MD97-2140, lozenges) and in lacustrine sediment (MM8, squares).

Basin) and/or deposition rate (Eauripik ridge) were roughly synchronous, increasing dramatically during periods of low and/or rapidly changing sea-level.

The records of Corg and OREC present relatively higher values in carbonate-poor sediment, glacial stages 8, 6, 4 and 2 (Fig. 5). A positive correlation ( $r=0.85$ ) between OREC and Corg values suggests either that “soot” carbon was not uniquely isolated during the OREC extraction procedure, or a common behavior of OREC and Corg in the pelagic depositional environment. The latter case should not be excluded, since OREC (and, to a lesser extent, micro-charcoal and BC) and Corg belong to the fine organic particulate material, which is partly stored in the Dissolved Organic Carbon (DOC) reservoir (Masiello and Druffel, 1998). Because the deposition of DOC is strongly constrained by absorption on fine, clayey material (Premuzic et al., 1982; Ogawa et al., 2001), the Corg and OREC correlation possibly results from increased deposition and preservation of DOC in rapidly deposited carbonate-poor environments.

## 5.2. Carbonaceous biomass burning records

The concentration changes in charcoal and Black Carbon (BC) are shown in Fig. 6. Charcoal values range from 0.1 to 5 mm<sup>2</sup> g<sup>-1</sup> (0.6 to 49.4 mm<sup>2</sup> cm<sup>-2</sup> ky<sup>-1</sup>), and show several high amplitude and short duration increases during the last 360 ky, especially at 54–47 ky BP and after 12 ky BP (>29 mm<sup>2</sup> cm<sup>-2</sup> ky<sup>-1</sup>). BC values range between ca. 0.003% to 0.042% (0.08 to 5.04 mg cm<sup>-2</sup> ky<sup>-1</sup>), with abrupt BC spikes at ca. 53–43 and 12–10 ky BP, i.e. at the same time interval as major charcoal “events”. In contrast with the OREC and charcoal records, however, the BC record shows little change (range 0.003–0.015%) before the critical age of 52 ky BP. Moreover, for the whole 330–220-ky BP time-interval, the BC values (ca. 0.005%) lie near the relative  $L_Q$  (0.001%). Our data suggest that input in BC was not significant (<0.4 mg cm<sup>-2</sup> ky<sup>-1</sup>) until ca. 220 ky in the region, and increased dramatically at 52 and at 12 ky (>2 mg cm<sup>-2</sup> ky<sup>-1</sup>). Similar results are documented in the Sulu Sea (core MD97-2141; Beaufort et al., 2003), in

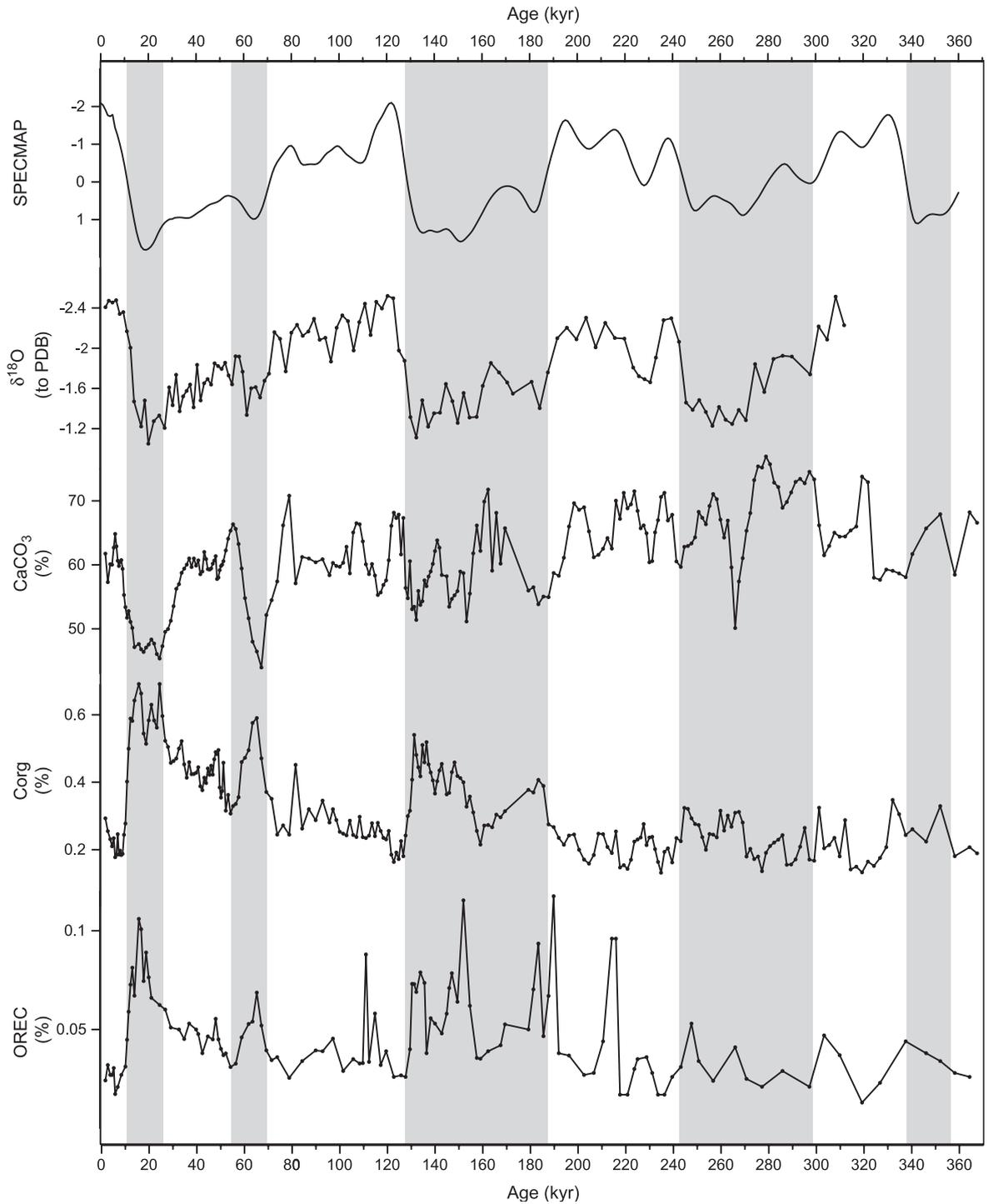


Fig. 5.  $\delta^{18}\text{O}$  (‰ to the PDB standard),  $\text{CaCO}_3$ , Corg, and OREC contents versus age (ky) for core MD97-2140, compared to SPECMAP stack  $\delta^{18}\text{O}$  record. Shaded areas indicate the glacial periods, unshaded areas indicate interglacial periods.

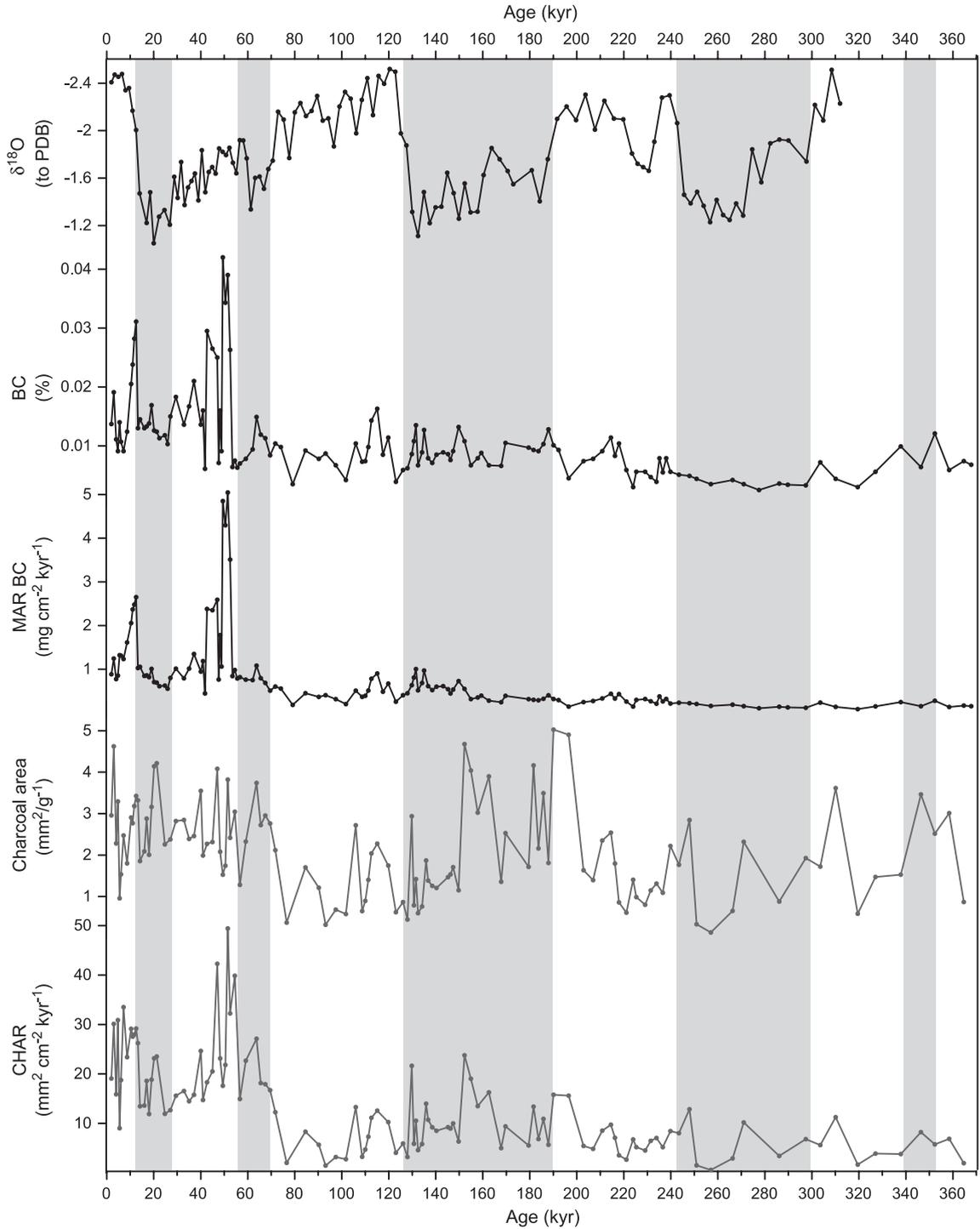


Fig. 6.  $\delta^{18}\text{O}$  (‰ to the PDB standard), and time series records of BC content, BC Mass Accumulation Rate (MAR BC), micro-charcoal area, and Charcoal Accumulation Rate (CHAR), versus age in core MD97-2140. Shaded areas indicate the glacial periods, unshaded areas indicate interglacial periods.

the Banda Sea (core SHI-9014; van der Kaars et al., 2000) and Northeast Australia (Lynch's crater and core ODP 820; Moss and Kershaw, 2000; Turney et al., 2001a), suggesting a regional increase in biomass burning during these periods (Fig. 7).

Although the ca. 220-ky BP increase in BC might reflect a regional trend in carbonaceous transportation processes and/or in fire intensity and frequency, the late 52- and 12-ky increases correspond respectively and remarkably with (i) the earliest major migration of *Homo sapiens* from Asia to Sahul (Australia and Papua New Guinea joined when sea levels were lower, Fig. 7), between 60 and 50 ky BP (Roberts et al., 1994), and (ii) the Early Holocene major colonization of non-coastal areas in Indonesia and Papua New Guinea, associated with a change in regional vegetation (Lake Hordorli in the work of Hope and Tulip, 1994; Haberle, 1998).

The earliest *Homo sapiens* from south-east Asia, which may be old enough to be associated with the first migration of people to Australia, is a skull from Niah Cave in Borneo estimated at around 40 ky BP

(Oakley et al., 1975; Brown, 1997) (Fig. 7). However, the date at which Australian Aborigines entered Australia has now been extended into the Late Pleistocene. This is especially supported by archaeological sites of ca. 40 ky BP in Southwest Australia (Upper Swan; Pearce and Barbetti, 1981), Northwest Australia (Carpenter's Gap; O'Connor, 1995), Papua New Guinea (Huon Peninsula; Groube et al., 1986), and by archaeological sites of ca. 50 ky BP in North Australia (Malakunanja II; Roberts et al., 1990), Southwest Australia (Devil's Lair; Turney et al., 2001b) and Southeast Australia (Lake Mungo; Bowler et al., 2003) (Fig. 7).

### 5.3. Frequency domain comparisons

Because the charcoal and BC records, in contrast to OREC, likely contain a specific biomass burning signal, we performed a cross-spectral analysis between BC and charcoal (Fig. 8). The spectral analysis was computed by using the Analyseries software (Paillard et al., 1996) over the last 220-ky

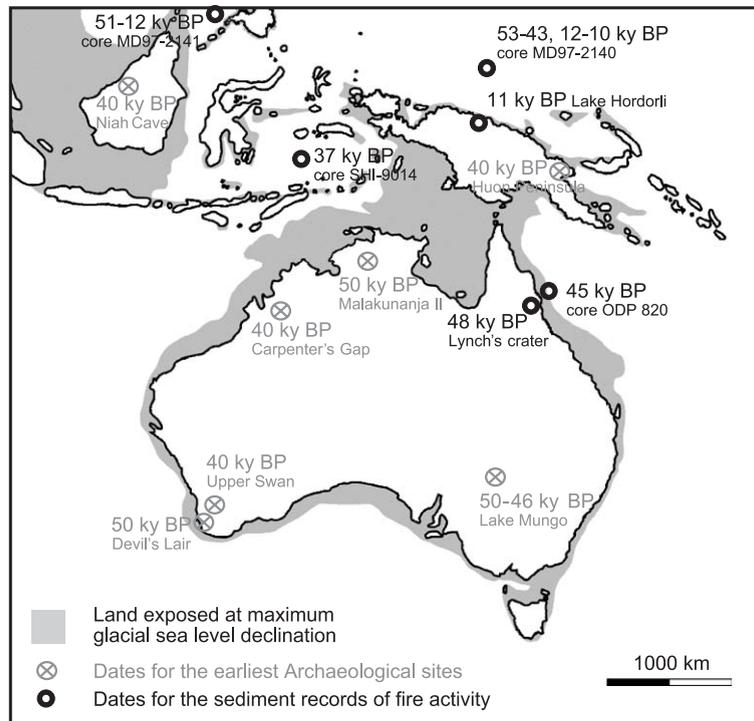


Fig. 7. Regional map of the Papua New Guinea and Australia region with ancient exposed land and shaded (Sahul), indicating the location of archaeological sites (modified from Brown, 1997) and sediment records of biomass burning.

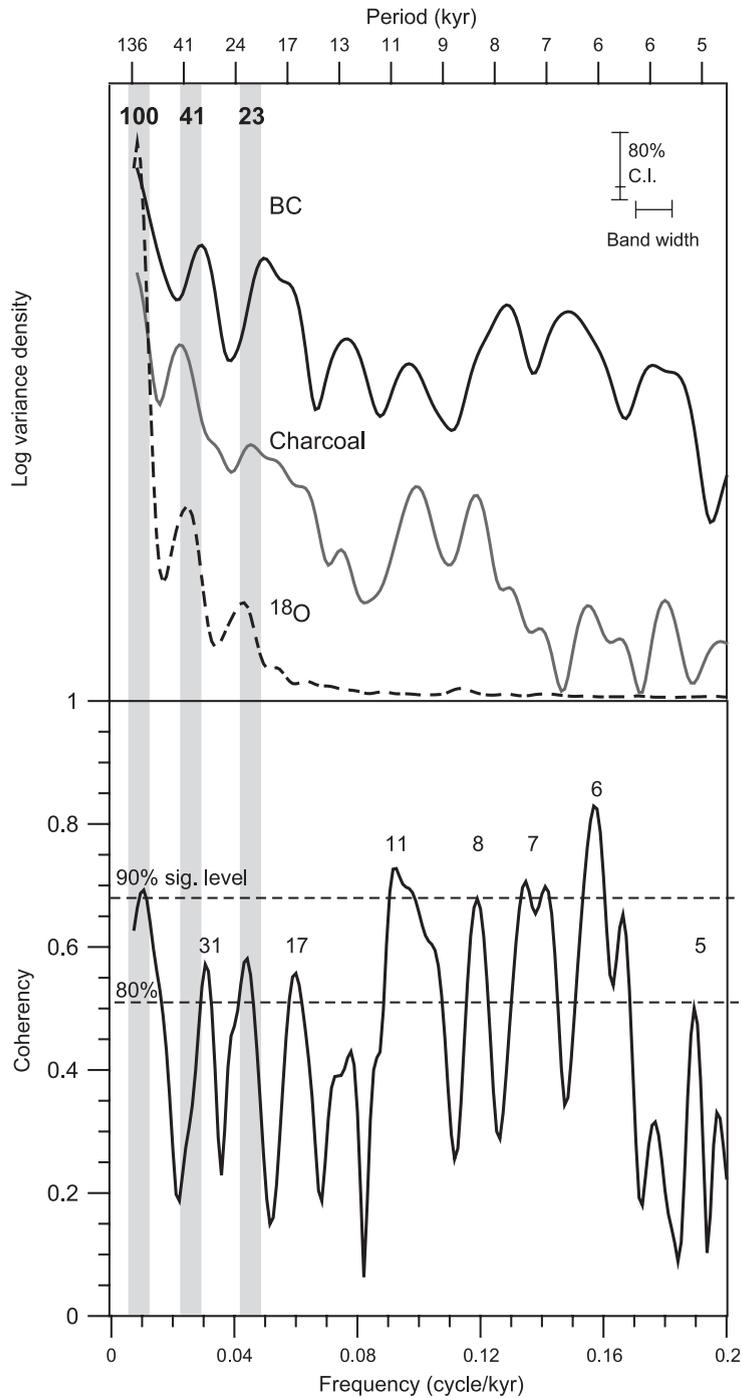


Fig. 8. Cross-spectral (Blackman–Tuckey) analysis between BC and charcoal.

time interval, where the data are significant. In addition to the ca. 23-ky precession band, charcoal and BC are coherent for several periods, especially 31,

17, 11, 7, 6 and 5 ky. These observations suggest that BC and charcoal biomass fire proxies in core MD97-2140 are strongly constrained, by low-latitude, pre-

cession-controlled insolation change, and monsoon dynamics. The ca. 30- and 19-ky cycles have been identified on charcoal records from the Sulu Sea (core MD97-2141; Beaufort et al., 2003) and the Northeast Australia (core ODP 820), but also on the pollen of those taxa susceptible to burning in cores ODP 820 and SHI-9014 (Fig. 7) (Kershaw et al., 2003). It has been attributed to the competing influence of the long-term ENSO-like forcing and the glacial/interglacial cycle on the East Asian summer monsoon (Beaufort et al., 2003). The non-linear response to the 23-ky cycle (ENSO) and the 100-ky cycle (glacial/interglacial) induces a shift of energy from the 23-ky period to periods of 30 and 19 ky ( $1/100 - 1/23 = 1/29.9$  and  $1/100 + 1/23 = 1/18.7$ ).

The 11-, 7- and 5-ky periods might correspond to precessional harmonics ( $p^2=12$  and  $9.5$ ;  $p^3=7.3$ ;  $p^4=5.5$  ky; Pokras and Mix, 1987). The half-precession cycle (11 ky) is attributed to the twice-yearly insolation maximum at the Equator radiation (March/September), which strongly controls convection and rainfall over Papua New Guinea (Short et al., 1991). Studying the past-fluctuations of some East-African lake levels has also proved that half-precession cycles strongly influenced the extension of the ITCZ and summer monsoon dynamics in intertropical Africa (Thevenon et al., 2002; Trauth et al., 2003).

In addition to the precession control, the highest coherency between BC and charcoal change is found at 6.4–6 ky. The ca. 6-ky period has been identified as one of the main frequency domains of submillennial variability of the East Asian winter monsoon in the Sulu Sea (de Garidel-Thoron et al., 2001), and might reflect a connection with the dominant ca. 6 ky pacing of the northern hemisphere climate and iceberg discharges in the North Atlantic (Bond et al., 1993, 2001; Porter and Zhisheng, 1995).

## 6. Discussion

### 6.1. Reliability and significance of carbonaceous biomass burning proxies

Atmospheric transport and fallout of particulate products by fire are strongly affected by atmospheric circulation and precipitation patterns, and the biomass burning relationship to the stratigraphic record can

differ with sedimentary environment (Garstang et al., 1997; Novakov et al., 1997; Stocks and Kauffman, 1997). Indeed, the fire-signal can be biased through secondary remobilization of material remaining on the ground after a fire by wind, and/or by redistribution from terrestrial soils and sediment through erosion and runoff processes (Griffin and Goldberg, 1975; Andreae et al., 1984; Emerson and Hedges, 1988). Smoke particulate emissions mostly consist in BC-like carbonaceous species, with fine particulate carbon in the submicrometer-size range accounting for 50–80% of the total particulate carbon concentration (Cachier et al., 1985). In fact, the gaseous and particulate composition, as well as the grain-size distribution of particulate carbon, strongly depend on the importance of the ratio of flaming to smoldering combustion, and to the type of burned vegetation i.e. plants, wood or fossil fuel (Cachier, 1989; Lobert and Warnatz, 1993).

The similar behaviors of Corg and OREC suggest that OREC is associated with organic deposition and organic content. Given the complex origin of pelagic organic matter (Meyers-Schulte and Hedges, 1986), the chemical extraction of soot carbon (OREC) was probably not achieved without including any additional thermal treatment. The resulting bias on OREC data would especially occur if the MD97-2140 organic matter contains significant contributions of (i) oxidized, soil-originating organic matter such as humic acids, or (ii) marine, bacterially oxidized organic matter (Masiello and Druffel, 1998). Further analyses are therefore needed to better understand the similarity between OREC and Corg records.

In contrast, charcoal and BC records show a relatively independent behavior from global  $\delta^{18}\text{O}$  record. This is most strikingly observed with the 54–52- and 12–10-ky large increases, the ENSO-like ca. 30- and 19-ky periods, and half-precession ca. 11-ky period. This suggests that the input of biomass fire products is dominantly constrained by the regional climate. Similar changes in BC and charcoal likely originate from the input of smoke particulate emissions, while different BC and charcoal features likely result from independent processes (e.g., intensity of combustion, soil erosion and runoff). For example, the lack of obliquity (41 ky) signal in the BC record (Fig. 8) suggests that ca. 41 ky changes in charcoal fluxes rather originated from changes in ice-volume and sea-level. Based on such criteria, major aerosol emissions

probably occurred at (i) 54–52 and 12–10 ky BP and (ii) at the precession-originating frequencies which control the rainfall distribution and the resulting probability of fires (Haberle and Ledru, 2001) over Indonesia, Papua New Guinea and Northern Australia.

### 6.2. High latitude forcing on pelagic deposition and atmospheric circulation in the Caroline basin

Although numerous low-latitude pelagic and terrestrial records show a connection with high-latitude, glacial–interglacial climate (Beaufort et al., 2001; Barker and Gasse, 2003), the connection between major, low-latitude atmospheric circulation features (such as monsoon or ENSO) and glacial–interglacial changes still needs better understanding. This is especially the case in the marine realm, where glacial–interglacial changes in global sea-level and bottom circulation strongly constrain the global depositional environment, with major effects of continental shelf erosion, bottom ventilation, and associated carbonate and organic matter preservation processes. The identification, in core MD97-2140, of high-frequency changes in aerosol emissions that present the same timing as high-latitude Dansgaard–Oeschger cycles (ca. 6 ky) may provide a further, independent evidence of a coupling between high and low-latitude atmospheric circulation patterns. As given by the (i) present-day association between biomass fire and ENSO-originating droughts in the study area, and (ii) the past association between monsoon circulation, ENSO-like patterns and BC/charcoal inputs, a potential high latitude forcing on the East-Asian winter monsoon and ENSO-like patterns should not be excluded.

### 6.3. Late Pleistocene–Early Holocene human-induced changes in fire activity

Despite the apparent climatic control over fire activity, human colonization and social change is almost everywhere marked by a new fire regime (Pyne and Goldamer, 1990). Natural fires occur when the fuel is dry, while most human fires are ignited while the fuel is moist: As a result, a shift in fire regime can double or even triple the emissions of non-CO<sub>2</sub> emissions, such as microcharcoal or BC (Saarnak, 2001). This should be considered to explain the two

major increases in charcoal and BC from ca. 53 ky and 12 ky BP, respectively. Migration of *Homo sapiens* from Asia to northern Australia occurred prior or around 50 ky. From that time, higher burning levels are supported by palynological data, assessing the impact of Aboriginal people on the vegetation and landscape (Kershaw, 1986). Archaeological sites and artifacts (Fig. 7) show that Sahul was colonized by people adapted to a coastal way of life, with initial colonizing routes located around the coasts and then up the major river systems. Non-aquatic adaptations, such as desert and montane economies, came relatively late in the archaeological record, from ca. 12–10 ky BP in Papua New Guinea (Bowdler, 1977; Haberle, 1998). Such observations therefore suggest that the two major charcoal and BC peaks of core MD97-2140 at 54–52 and 12–10 ky BP closely correspond to the main Late Pleistocene and Early Holocene periods of human colonization and changes in practices.

Peculiarly, both 54–52- and 12-ky events occur during relatively cold glacial events of the last climatic cycle in the Greenland–North Atlantic region (Dansgaard et al., 1993). The 54–52-ky event would locate between warm Dansgaard–Oeschger (DO) events 16 and 14, while the 12-ky BP event would correspond to the Younger Dryas. Both events clearly belong to the ca. 6-ky periodic component of the DO record. Taken together, such data therefore suggest that strengthened smoke emissions and fire activity, at the time of major human colonization events, did not occur independently of relatively dry, ENSO-like conditions.

## 7. Conclusions

Our study provides a pelagic record of carbonaceous biomass burning proxies in the West Pacific Warm Pool, north of Papua New Guinea. Several conclusions are stressed from this work:

- (1) Except from the BC record, the MD97-2140 record is strongly constrained by depositional changes coupled to global changes in ice-volume. This confirms the tremendous effects of global changes in ice-volume and sea-level on (i) the continental surface of this equatorial region, and the resulting inputs of terrestrial

matter by (shelf) erosion, and (ii) deep-sea chemistry and circulation in relatively closed “glacial” basins such as the Caroline basin.

- (2) The concentration of organic carbon which survives to chemical oxidation (OREC) shows a strong association with the Corg content. The OREC record thus encompasses variable forms and sources of organic carbon and poorly reflects on the input of biomass fire products.
- (3) In contrast with Corg and OREC, the independent and coherent changes in BC and charcoal content consistently reflect the regional emission of smoke carbonaceous particulate material.
- (4) The BC/charcoal record shows a close relationship with precession-originating insolation change. This suggests that biomass burning emissions, associated with drought and ENSO-like conditions over Papua New Guinea and Northern Australia, were strongly constrained by monsoon circulation. Additional half-precession periods in such records further suggest that the insolation in the equatorial band strongly controlled the seasonal variations in precipitation and aridity.
- (5) The high BC/charcoal coherency around ca. 6-ky periods suggests that the northern hemisphere climate dynamics as recorded in Greenland ice-cores are connected with monsoon dynamics and ENSO-like patterns in the Indo-Pacific region.
- (6) The major Late Pleistocene and Early Holocene periods of human colonization in Australia and Papua New Guinea at ca. 54–52 ky and ca. 12–10 ky BP, respectively, resulted in dramatic increases in emission of carbonaceous aerosols in the environment. Although such events still belong to the ca. 6-ky periodic component of biomass burning emission, amplitude of such events likely illustrate the consequence of a shift from natural to human fire regime.

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