Biomass burning and oceanic primary production estimates in the Sulu Sea area over the last 380 kyr and the East Asian monsoon dynamics

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Abstract

Coccolithophorid assemblages and micro-charcoal content were analysed in giant piston core MD97-2141 recovered in the Sulu Sea (Philippines). These proxies help to reconstruct respectively the dynamics of the oceanic primary production (PP) and biomass burning in that area. PP in the Sulu Sea intensifies during the East Asian winter monsoon (EAWM) and therefore PP constitutes a proxy for EAWM dynamics. Most of the precipitation in the Sulu Sea region occurs during the East Asian summer monsoon (EASM). Because the intensity of biomass burning is related to dryness of the surrounding area, the sedimentary micro-charcoal content can be used as an inverse proxy for EASM intensity. Our results show that the EAWM intensifies during glacial times in agreement with previous studies. Precessional forcing appears to act directly on EAWN because of the early response of PP in that frequency band. The micro-charcoal record exhibits complex dynamics, which we attribute to the competing influence of the long-term El Niño Southern Oscillation (ENSO)-like forcing and the glacial/interglacial cycle on EASM. These influences create an unusual frequency spectrum with power around 30 kyr and 19 kyr attributed to the non-linear response to the 100-kyr cycle (glacial) and the 23-kyr (ENSO) cycle. A factor of two increase in the amplitude of the micro-charcoal variability between 51 and 10 ka BP could correspond to \textit{Homo sapiens} biomass burning in the style of the fire-stick farming of the Australian Aborigines. We also find, on precession cycles, an opposite phase between EASM and EAWM records and an advance of $\delta^{18}O$ and $\delta^{18}O$ respectively by 2000 yr.

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1. Introduction

The monsoon climatic system is characterised by a seasonal reversal of the wind direction due to differential seasonal heating of the continents and oceans. In Asia there are two major monsoon systems: the Indian monsoon and the East Asian monsoon. Monsoon winds come from the north during boreal winter and from the south during boreal summer. The relationship existing between...
long-term (orbital-scale) variability of these monsoon systems and the seasons has been difficult to determine. The long-term dynamics of the Indian summer monsoon (ISM) is well established from sediment tracers in the Arabian Sea (e.g. Clemens and Prell, 1990), where it produces upwelling that strongly imprints the quality of the sediments (Anderson and Prell, 1993). It appears that Arabian southern ISM decreases during glacial periods and is significantly influenced by the direct insolation forcing of precession (Prell and Kutzbach, 1992). The dynamics of the Indian winter monsoon are more difficult to resolve using oceanic sediments because in most places the southern monsoon shadows the winter monsoon imprints. However, in the Bay of Bengal, Duplessy (1982) and Sarkar et al. (1990) have shown that the winter monsoon was stronger during the last glacial period. Also Reichart et al., 1998, using the abundance pattern of two deep-dwelling planktonic foraminifera, have shown that the winter monsoon in the Arabian Sea was in almost phase opposition with the summer monsoon.

Does the East Asian monsoon system show the same long-term dynamics? The areas that are at the origin of the monsoons are different: the Tibetan Plateau and Indian Ocean in the case of the Indian monsoon and Siberia and the Pacific Ocean for the East Asian monsoon. Most of what we know from the past dynamics of the East Asian monsoon come from China lands and seas. Chinese loess–palaeosol sequences provide good records of both summer and winter East Asian monsoon dynamics (An et al., 1990; Liu, 1985). They indicate that the intensity of the two monsoon seasons has varied inversely (Xio et al., 1999), similar to the Indian monsoon. However, one can argue that this strict phase opposition results from the interpretation of the palaeo-loess sequences as a ratio between winter and
summer monsoon (see for example Porter and An, 1995). The relation between summer and winter monsoon is not yet known at lower tropical latitudes in East Asia, an area where El Niño Southern Oscillation (ENSO) may complicate the high-latitude dynamics shown in the loess sequence (Porter and An, 1995). In order to study the dynamics of both phases of the monsoon it is necessary to develop proxy records of the two monsoon seasons from the same core. The Sulu Sea is well suited for such a study because of its unique geography (enclosed basin away from major oceanic currents) and because both phases of the monsoon have distinct meteorological effects in the basin.

The Sulu Sea (Fig. 1) is located between the Asian continent and the ‘Western Pacific Warm Pool’ (WPWP) where annual sea surface temperature (SST) is above 29°C (Yan et al., 1992). The climate of the Sulu Sea is strongly influenced by the monsoons. The East Asian monsoon results from the difference in potential heating between the WPWP and the Asian continent. During the boreal winter, the main heating source is located in the ocean. The latent heat release associated with intense convective precipitation fuels meridional circulation. Tropical convection in the western equatorial Pacific is connected to the descending branch over the Siberian region, forming a strong local Hadley cell in the East Asian region (Zhang et al., 1997). The East Asian winter monsoon (EAWM) winds in the Sulu Sea result from the merging of the northerly East Asian monsoon with the Pacific trade winds over the South China Sea (McGregor and Nieuwolt, 1998).

EAWM bursts during January to March (Fig. 2) can induce blooms of coccolithophorids in the region (Wiesner et al., 1996). Primary production (PP) rises correlatively with wind stress strengthening, because of the stronger mixing of the upper ocean (Fig. 2) (Nair et al., 1989). Thus coccolith assemblages in the Sulu Sea record information on both palaeoproductivity changes and EAWM variations (de Garidel-Thoron et al., 2001). The winter monsoon corresponds to the dry season in the Sulu Sea whereas the summer monsoon brings

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Fig. 2. Monthly average of precipitation (solid line) and wind speed (dotted line) at the core location taken from the COADS database.
a lot of moisture to the Eastern Philippine Seas (Fig. 2). Originating from the subtropical high region, the air incorporates moisture as it moves west over the WPWP and provokes rain when it arrives in the region of the Philippines (McGregor and Nieuwolt, 1998). Today, the summer monsoon humidity dynamics are strongly modulated by ENSO. For example, during ENSO events such as 1997/1998, the strength of the summer monsoon is reduced and rains are extremely weak. During these events, the surrounding islands of the Sulu Sea experience drought and intense fires. Small charcoal particles from these fires (microscopic charcoal) are transported by wind and eventually fall into the sea and may be incorporated in the sediment. Downcore records of microscopic charcoal may serve as a proxy for dryness and therefore for summer monsoon intensity.

2. Material

A 36 m giant piston-core (IMAGES MD97-2141) (08°47'N, 121°17'E, 3633 m depth) was retrieved during the IPHIS–IMAGES III cruise of the R/V Marion Dufresne in May 1997 (Fig. 1). The core is located in the vicinity of ODP Site 769 (Linsley, 1996). The core is located on the Cagayan ridge, which protects the site from downslope processes and is above the present lysolcine depth (about 3800 m), allowing for good preservation of carbonates (Linsley et al., 1985; Miao et al., 1994). The sediments are composed predominantly of well-preserved nannofossil–foraminifera oozes and do not contain significant amounts of detrital mineral.

3. Methods

3.1. Microscopic charcoal

The preparation techniques for charcoal were adapted from the work of Winkler (1985) and classical palynology methods. About 0.07 g of bulk sediment was sampled every 10 cm down the core and precisely weighted. Carbonate was removed by adding 50 ml of 10 N HCl. The residue was rinsed with de-ionised water and 50 ml of nitric acid was added. After rinsing, the remaining sediment was transferred to a beaker containing 100 ml of hydrogen peroxide and soaked for 24 h. This procedure dissolves the most common opaque minerals, such as pyrite, oxides (e.g. magnetite) and hydroxides (e.g. hematite), that can be found in the sediment and leaves charcoal, micro-charcoal and soot, which are the only dark particles left in the remaining sediment. The sediment was rinsed, diluted in 500 ml water and 250 ml of this suspension was filtered onto a cellulose membrane of 47 μm nominal porosity and 45 mm in diameter (the effective filtering diameter is 20 mm). A portion of the membrane was mounted onto a slide with Canadian balsam. The slides were scanned with an automated Leica DMRBE microscope and 100 view-fields of 1.7 mm² were grabbed with a standard 756 × 582 pixels digitising camera. In a controlled light adjustment of the microscope, all charcoal appeared darker than a given threshold (grey level below 35 with the light adjustment used in that study). The number of pixels ($P$) having grey level values lower than 35 was counted in the scanned area. Knowing $P$, the surface of a pixel ($S_p = 25.6 \times 10^{-9}$ mm²), the surface of the filter ($S_f = 1250$ mm²), the surface scanned by the microscope ($S_s = 170$ mm²), the weight of sediment ($W$) and the dilution ($D = 0.5$), the results are given as the total surface area (in mm²) occupied by charcoal (MC) per gram of sediment:

$$MC = \frac{(P \cdot S_p \cdot S_f)}{(D \cdot W \cdot S_s)} = 0.00376P/W \text{ mm}^2/\text{g}$$

This automated optical method provides reproducible results (the difference of results between two different areas of the same slide is less than 1%), and the standard error given for the analysis of the 100 view-fields is 5.7 mm²/g on average for a mean value of 41.7 mm²/g.

A limiting feature of this method is that it is not possible to be certain that some non-completely dissolved opaque minerals were not measured. For most of the peaks, we visually checked that they did not correspond to increases of non-dissolved pyrite but to higher concentrations of
charcoal. In another oceanic core this method had been satisfactory compared with two chemical methods for measuring the black carbon content in sediment bases (Thevenon et al., manuscript in preparation).

Also we should mention an interpretative limit in the use of charcoals. Some charcoal could have arrived by river discharge and not reflect eolian input. Because the Western Pacific marginal seas experience strong sea-level variations, it is possible that fluvial detrital material varied considerably and that the record would reflect sedimentation variations rather than climatic dynamics. However, (1) the core has been retrieved from a ridge and therefore the record is protected from fluvial transport, (2) the Sulu Sea has relatively steep margins and did not vary extremely in size during glacial–interglacial cycles, (3) the core consists of an homogeneous hemipelagic mud with no visible downcore variations, and (4) as we will see the micro-charcoal record does not follow a clear glacial–interglacial cycle.

3.2. Coccoliths

3.2.1. Counting

The coccolith data and results have been published recently (de Garidel-Thoron et al., 2001). The core was sampled every 2 cm in the upper 6 m, allowing for a resolution of about 70 yr and every 3–4 cm in the lower 30 m for a resolution of about 200–500 yr. A smear slide was prepared for each sample, and at least 300 coccoliths were counted for each slide (mean 357 coccoliths) on a Zeiss Axioscop at a 1000× resolution. Percentages of *Florisphaera profunda* (Fp) were computed using the following equation:

\[
\%\text{Fp} = 100 \left( \frac{\text{number Fp}}{\text{total coccoliths}} \right)
\]

The 95% confidence interval for %Fp varies between ±2% and ±6% depending on the percentage of Fp (Mosimann, 1965).

3.2.2. Primary production transfer function

Most of the coccolithophorid species live in near-surface waters where light for photosynthesis is abundant. Phytoplankton also requires nutrients, and thus thrives where a shallow thermocline brings subsurface nutrients into the upper euphotic zone. In contrast, the coccolithophore *F. profunda* lives in the deep-photic zone where nutrients are relatively abundant and light is rare (Okada and Honjo, 1973). Where the thermocline is deep, total PP is low, and the dominant coccolith species in fossil assemblages is *F. profunda* (Molinino and McIntyre, 1990). As productivity increases, the relative abundance of *F. profunda* decreases. Thus, estimates of PP are made from counts of the relative abundance of *F. profunda* (%Fp) converted to PP using the following equation: %Fp by PP = 316log(%Fp+3) (Beaufort et al., 1997). This equation derives from the fit of Indian Ocean core-top coccolith counts with modern productivity based on satellite chlorophyll (Antoine et al., 1996). The correlation between the estimated and observed productivity in the calibration data set is \( r = 0.94 \) and the standard deviation of the residuals is \( \pm 26 \) g carbon m\(^{-2}\) yr\(^{-1}\). This transfer function has been shown to be reliable in the equatorial Atlantic (Henriksson, 2000) and Pacific Ocean (Beaufort, unpublished).

3.2.3. Chronostratigraphy

The age model of the core includes 28 AMS \(^{14}\)C ages (converted to calendar ages) on *Globigerinoides ruber* and *G. sacculifer*. Older ages were selected by correlation of the planktonic foraminifera \(^{81}\)O curve (on *G. ruber* tests) to the SPECMAP stack (Imbrie et al., 1984). The age model is presented in further detail elsewhere (de Garidel-Thoron et al., 2001; Oppo et al., 2003). Two ninth-order polynomial regressions were used from the core top to 400 cm and from 440 to 920 cm on the \(^{14}\)C ages and SPECMAP tie-points to smooth the sedimentation rate for the last 60 kyr. This smoothing is indispensable for the spectral analyses to avoid spurious peaks linked with sedimentation rate changes.

The average sedimentation rate is about 10.5 cm/kyr, with a maximum during glacial stage 2 of 34 cm/kyr. For example, the sedimentation rate during stage 3 is about 30 cm/kyr, which allows a 70 yr resolution (2 cm sampling). Lower sedimentation rate (4.4 cm/kyr) occurs in a short interval of 40 cm between 28 and 19 ka BP. This could be indicative of the presence of a hiatus in
that interval. We check that the use or the withdrawal of that interval has no influence on the results of time series analysis.

3.3. Time series analysis

To extract the significant periodicities contained in the PP signal, we performed spectral analysis using Blackman–Tukey and Cross–Blackman–Tukey methods provided in the package Analyseries (Paillard et al., 1996).

4. Results

The present annual PP in the Sulu Sea is 148 g carbon m$^{-2}$ yr$^{-1}$ (Antoine et al., 1996), which is close to the average reconstructed PP over the last 380 kyr of 149 g carbon m$^{-2}$ yr$^{-1}$. The PP$^1$ oscillates during the last 380 kyr between 80 and 330 g carbon m$^{-2}$ yr$^{-1}$ (Fig. 3). On glacial–interglacial time scales, PP increases during glacial periods and decreases during interglacials (Fig. 3). PP is moderately correlated ($r^2 = 0.49$) with the ice-volume curve (SPECMAP stack of Imbrie et al., 1984). This is shown by spectral analyses of the palaeoproductivity record, which contains peaks corresponding to the main orbital periods of Milankovitch theory (i.e. 1/100 kyr, 1/41 kyr and $\sim 1/20$ kyr) and confirmed by cross-spectral analyses between $\delta^{18}$O and PP (Fig. 4). Cross-spectral analyses indicate that PP leads $\delta^{18}$O by 25° in the precession frequency band and lags by 39° in the obliquity frequency band.

The absolute abundance of microscopic charcoal varies between 1 and 185 mm$^2$/g. The flux of charcoal was also calculated. Flux and absolute abundance are strongly related. We discuss only absolute abundance, because flux data may be biased by discrete chronology and possible core disturbances that occurred during the piston-corring process (Thouveny et al., 2000). The main frequencies are a strong 29-kyr$^{-1}$ and a 19-kyr$^{-1}$ one, the latter related to precession (note the small 23-kyr$^{-1}$ peak) (Fig. 5). Micro-charcoal abundance increases during glacial periods, but does not correlate with $\delta^{18}$O or PP. Only the long-term trend (see dotted line in Fig. 3) shows recurrent increases during glacial times. There is an important increase starting at about 60 ka BP and marked by the highest abundance of charcoal at 51 ka BP. After this event the charcoal abundance remains high in comparison with older sediment. Before 60 ka the average charcoal abundance is 31 mm$^2$/g and after 60 ka the average is 71 mm$^2$/g, indicating a doubling of the charcoal. Charcoal abundance decreased significantly during the last deglaciation, paralleling the $\delta^{18}$O record. Between 380 and 300 ka the charcoal abundance is relatively high with an average of 42 mm$^2$/g. That long trend, described above (dotted line in Fig. 3), partly represents the glacial–interglacial cycle, partly a non-cyclic trend.

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5. Discussion

5.1. Winter monsoon dynamics

Spectral analysis of the PP records shows all Milankovitch frequencies. The dominant frequencies are 100-kyr and 41-kyr cycles. Cross-spectral analysis also indicates highly (> 99%) significant coherence with \(^{18} \text{O} \) at these frequencies. PP lags \( \delta^{18} \text{O} \) by 40° (5 kyr) in the 41-kyr band and is in phase with \( \delta^{18} \text{O} \) in the 100-kyr band. This implies that PP is responding strongly to global variations expressed by the \( \delta^{18} \text{O} \) record. The PP in the Sulu Sea reacts to variations in winter monsoon dynamics (de Garidel-Thoron et al., 2001) (Fig. 2). The dynamics of the winter monsoon for the last 380 ka are clearly linked with the glacial–interglacial cycles and therefore imply some type of teleconnection between low and high latitudes. The strengthening and weakening of the Siberian high-level pressure cell is often suggested to explain such a relationship between the North Atlantic and EAWM, a relationship that has also been found in Chinese loess records (Porter, 2001; Porter and An, 1995) and in marine records from the South China Sea (Wang et al., 1999a,b). A strengthening of the Siberian high in the cold northern continent during glacial times contrasted with the relatively warm tropical Indo-Pacific Ocean, reinforcing the winter monsoon (Wang et al., 1999a).

In the precession frequency band the phase between \( \delta^{18} \text{O} \) and PP is different from what is observed at other Milankovitch frequencies. In the precession band, changes in PP lead the \( \delta^{18} \text{O} \) changes by 2000 yr. This lead of PP is a common feature of the equatorial Indo-Pacific record (Beaufort et al., 1997, 1999, 2001). Precession appears to induce a direct PP response. Therefore the 20-kyr signal present in the Sulu Sea PP records is not only the response of the 20-kyr signal recorded in the ice-volume but also results from a 20-kyr variation of a low-latitude variability of...
the structure of the thermocline. These fluctuations follow an ENSO-like mechanism equivalent to the present 2–6 yr ENSO cycle but acting on a much longer time scale.

The Sulu Sea being protected by the Philippines from the basinwide variability of the thermocline, one could wonder about the reasons why a similar feature is also observed in that area. We suggest the existence of a link between ENSO-like variability and the East Asian monsoon variability, equivalent to the link observed between modern ENSO and EAWM. For example, (Wang et al., 2000) established a clear teleconnection between warm (cold) events in the eastern Pacific and the weak (strong) EAWM. Strong winter monsoons are the response to colder than normal temperatures in the Siberian highs and/or warmer than normal tropical SST. By analogy with the modern climate, long-term variations of the WPWP SST related to the ENSO-like cycles may have had a strong influence on the East Asian monsoon.

The PP response to two forcings acting at the same frequency induce a non-linear response which is complex and expressed in the frequency spectrum by the shift of frequency from 23- to 18-kyr, 20-kyr, and 28-kyr cycles. This type of complex response is more prominent in the micro-charcoal record and is explained in greater detail in the next section.

A half-precession signal (10-kyr) is significantly revealed in the record. It may correspond to the fact that the Inter-Tropical Convergence Zone (ITCZ) passes over the Sulu Sea twice a year, and therefore there are two important periods of insolation maxima in a precession cycle that can influence the ITCZ and ENSO. The theoretical impact of a twice-yearly passage of the sun across the equator sites, creating peaks at 10 and 12 kyr, is discussed in Short et al. (1991).

Finally, a 5.8-kyr period appears in the record. It corresponds to millennial PP variations that are discussed in detail by de Garidel-Thoron et al. (2001), who use the full resolution of the record (100 years) instead of the 1000-yr resolution we have here. The 5.8-kyr period is interpreted as corresponding to the rapid response of winter monsoon to the northern ice-sheet dynamics. Indeed some degree of resemblance exists between GRIP and PP records for the Dansgaard/Oeschger cycles (de Garidel-Thoron et al., 2001).

5.2. Biomass burning

5.2.1. Human impact

The micro-charcoal content is interpreted as related to the aridity of the surrounding terrestrial area of the Sulu Sea, but the long-term variations show a peculiar trend with strong concentration of charcoal in MIS 2 and 3 between 51 and 13 yr BP. These high concentrations are not reached in the other glacial periods (MIS 6, 8 and 10), nor at the onset of the last glacial during MIS 4. Higher values in the last glacial than in the penultimate glacial are difficult to explain. A similar case is found in Australia and has been explained by intense anthropogenic fires (Kershaw, 1986), although other (e.g. direct) evidence has yet to be found (Moss and Kershaw, 2000). The earliest Homo sapiens occupation in the Philippine Archipelago has been estimated at around 50 000 yr BP (Fox, 1970) in the Tabon caves (in Palawan Island located north of the Sulu Sea) that yield artefacts. These artefacts as well as others found in Borneo (Niah in Sarawak) show little sign of any marked stone tool ‘evolution’ until nearly 10 000 yr BP (Harrison, 1976). Early Holocene human skeletons from Malaysia, Tabon and Niah indicate strong morphological affinity with the Australian Aborigines (Matsumura and Zuraina, 1999). By 50 000 yr BP Australian Aborigines reached southwestern Australia (Turney et al., 2001) and they were on the lands around the Sulu Sea until the beginning of the Holocene. These dates correspond surprisingly well with the high charcoal content in core MD97-2141 (51 000–12 000 yr BP). At that time humans did not practise agriculture, but they could have used burning for hunting and clearing the vegetation (fire-stick farming) as the Australian Aborigines did until the end of the 18th century (Nicholson, 1981). The same type of culture may have existed around the Sulu Sea. The large and prominent peak of charcoal at 51 ka could mark the arrival of H. sapiens around the Sulu Sea. Although the charcoal content in MD97-2141 increased by a factor of two after this event, the higher-fre-
quency variability is similar before and after 51 ka. For instance, the relatively low Holocene charcoal content is more diagnostic of wetter conditions during interglacials than of a change of human habits. The ‘natural’ summer monsoon dynamics are therefore the major mechanism responsible for the variability of biomass burning in that area.

5.2.2. Summer monsoon

Because most of the precipitation falls during the summer monsoon (Fig. 2), micro-charcoal concentration is interpreted as an inverse summer monsoon proxy. The record indicates that summer monsoon was significantly reduced during glacial time. Pollen evidence suggests increased aridity in Australia and Indonesia (van der Kaars and Dam, 1995; van der Kaars et al., 2000, 2001). A glacial weakening of the Indian summer monsoon has also been extensively described (Cullen, 1981; Duplessy, 1982; Gasse and Van Campo, 1994; Kutzbach, 1981; Prell and Curry, 1981; Prell and Kutzbach, 1992). Chinese loess also indicates lower summer monsoon intensity in Northeast Asia (Porter, 2001; Xio et al., 1999). It appears that glacial weakening is a common feature of the Indian and Asian branch of the summer monsoon. However, our micro-charcoal data show that the link between summer monsoon intensity and glacial cycles is weak in the Sulu Sea, in comparison, for example, with the winter monsoon, which exhibits a strong relationship with glacial cycles.

ENSO also strongly influences the intensity of the summer monsoon rain in that area: dry weather conditions prevail during El Niño events. On a Milankovitch time scale, model and data indicate that El Niño is more frequent during interglacial times (Andreasen and Ravelo, 1997; Beaufort et al., 2001; Bush and Philander, 1998; Fedorov and Philander, 2000). The strength of summer monsoon related to glacial cycles (Prell and Curry, 1981) and the long-term dynamics of ENSO (Fedorov and Philander, 2000) have an opposite influence on summer rains on a glacial–interglacial time scale. This competing influence provokes a non-linearity between precession (insolation) and eccentricity (glacial/interglacial variations) expressed by the strong 30-kyr cycle revealed by the spectral analyses of the charcoal record.

A simple way to explain that non-linearity is to separate the 100-kyr cycle, which we will arbitrarily call the ‘glacial cycle’, from the 23-kyr cycle, called the ‘precession cycle’. In a glacial cycle, during a warm phase the summer monsoon strengthens (Prell and Curry, 1981) and El Niño is stronger (and/or more frequent) (Fedorov and Philander, 2000). In a precession cycle, δ¹⁸O warm phases lead strong monsoon by about 30° (Clemens et al., 1991) and strong El Niño by 150° (Beaufort et al., 1997, 1999, 2001) (which is equivalent to saying that El Niño-like variability leads −δ¹⁸O by 30°). During the cold period of a glacial cycle, weak monsoons limit the amount of rainfall in the Sulu Sea. Monsoon is therefore the ‘limiting factor’. Changes in the summer monsoon induced by local insolation should increase the amount of rain. In consequence, precipitation follows the phase of the monsoon of precession cycles during the cold period of a glacial cycle. During the warm period of a glacial cycle, El Niño is more frequent and stronger and therefore ENSO constitutes the limiting factor for rainfall. During these periods changes in precipitation are following the ENSO phase of precession. Therefore, the phase in the precession band will shift once by 120° during a glacial cycle. This is what is shown in the upper panel of Fig. 6, where the detrended charcoal record appears to vary between being in and out of phase with low-latitude insolation. Therefore, this suggests a phase modulation of the climate response to precession by eccentricity. That induces a shift of energy from the 23-kyr period to periods of 30 and 19 kyr (1/100 − 1/23 = 1/29.9 and 1/100 + 1/23 = 1/18.7). These frequencies are those observed in the charcoal spectra (Fig. 5). To illustrate this hypothesis, in the time domain, we inverted the precession signal of the SPECMAP record for each interglacial period as in figure 6, lower panel, of Beaufort et al. (2001). The relation between the smoothed charcoal record and this new series (SPECMAP-transformed) shows a good phasing between the peaks. The cross-spectral analysis between the charcoal record and the transformed SPECMAP stack in-
dicates significant coherence at periods near 30 and 19 kyr.

Half-precession cycles (12 and 9.5 kyr) are well expressed in the series (Fig. 5), indicating the importance of the migration of ITCZ for the summer monsoon dynamics.

5.3. Relation between winter and summer monsoons

Micro-charcoal and PP show significant coherence (95%) around 21 kyr$^{-1}$ (Fig. 7). The two series vary in phase, meaning that summer monsoon (low micro-charcoal) and winter monsoon (high PP) are in total opposite phase, as in the Arabian Sea (Reichart et al., 1998) and loess sequences (Porter and An, 1995). Therefore it seems that this opposition is a common feature in the different areas. The phase, however, is different from what has been found in the Arabian Sea since the winter and summer monsoons are maximum about $\sim 3$ kyr after ice maxima and ice minima, respectively, whereas in the Sulu Sea the two monsoon records precede by $\sim 2$ kyr the ice volume variations. At longer time scales (obliquity and eccentricity) there are no common dynamics, which is logical since the micro-charcoal does not show any variability in this frequency band.

6. Conclusions

Our results show that summer and winter monsoons in the Sulu Sea vary in opposition on most time scales because they have an inverse response to many climatic parameters:

1. Summer monsoon decreases and winter monsoon increase during glacial time.

2. Present-day climate shows that ENSO events positively influence winter monsoon winds (Wang et al., 2000) whereas they decrease summer precipitation (e.g. 1997 ENSO event). Precession-induced long-term ENSO variability (Clement et al., 1999) explains the opposite pattern of summer and winter monsoon dynamics in precession time scale and also the
complex pattern of variability of the summer monsoon.

3. The relation between the two records is difficult to establish because winter monsoon dynamics is essentially driven by glacial/interglacial dynamics whereas the summer monsoon is more strongly influenced by ENSO cycles. However, on the precession time scale the summer monsoon and the winter monsoon appear to be in complete phase opposition, and in advance to δ18O (or −δ18O) by about 2000 yr.

4. *H. sapiens* might be responsible for the doubling of charcoal abundance after 51,000 yr BP. Local biomass burning increased without changing the natural dynamics of fire frequencies that are under the direct influence of summer monsoon dynamics.

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