A 350,000 year summer-monsoon multi-proxy stack from the Owen Ridge, Northern Arabian Sea

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Abstract

Five summer-monsoon proxies from the Northern Arabian Sea are combined using stacking and principal components analysis (PCA) to create two very similar multi-proxy records of summer-monsoon variability. The five individual proxies all respond to monsoon variability but are largely independent in terms of the processes that complicate their interpretation as summer-monsoon indicators (e.g. preservation, dissolution, diagenesis, sediment reworking). As such, stacking and PCA average out non-monsoon variance, yielding a more pure monsoon signal. These stacked and PCA records (hereafter summer-monsoon stack and summer-monsoon factor) allow evaluation of relative monsoon strength through time as well as the relative concentration of variance within orbital bands; these two parameters are less reliable when estimated from individual proxy records. In fact, the summer-monsoon factor (SMF) accounts for only 33% of the total variance in the five records, suggesting that relative amplitude variations in each individual proxy time series are influenced by non-monsoon processes. The summer-monsoon stack (SMS) and SMF are spectrally very similar, dominated by variance in the 41-k.y. (obliquity) and 23-k.y. (precession) bands; there is very little variance at the 100-k.y. (eccentricity) band associated with large-scale changes in global ice-volume. Indeed, equally strong monsoons occur in both glacial and interglacial intervals. Within the 23-k.y. precession cycle, monsoon maxima fall at −121° relative to precession minima (June 21 perihelion, maximum Northern Hemisphere (NH) summer insolation). This phase falls midway between δ18O minima (−78°) and December 21 perihelion (−180°) indicating that two mechanisms exert equal influence in determining the timing of strong summer monsoons within the precession band: (1) sensible heating of the Asian Plateau which is maximized at times of ice-volume minima (−78°), and (2) latent heat export from the southern subtropical Indian Ocean which is maximized at times of December 21 perihelion (−180°). The seasonal cycle at December 21 perihelion is characterized by warm Southern Hemisphere (SH) summers followed by cold SH winters, a combination that preconditions the ocean to export latent heat during the boreal summer-monsoon season. Summer-monsoon winds transport this latent heat into Asia where it is released during precipitation, enhancing the Asian monsoon low. Within the 41-k.y. obliquity cycle, monsoon maxima are in phase with obliquity maxima. This indicates that two mechanisms, quite similar to those in the precession band, influence the timing of strong summer-monsoons in the obliquity band: (1) sensible heating of the Asian Plateau but with no ice-volume delay, and (2) latent heat export from the southern subtropical Indian Ocean which is maximized at times of obliquity maxima. Again, the seasonal cycle at obliquity maximum is characterized by warm SH summers followed by cold SH winters, ideal for maximizing latent heat export during the boreal summer monsoon.

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1. Introduction

Evaluating past paleoceanographic and climate change through the analysis of deep ocean sediments necessarily involves the use of proxy indicators. Proxy indicators, by virtue of their physical, biological, and/or chemical origins, are indirectly related to paleoceanographic and climatic variables of interest (e.g., sea surface temperature, wind speed, nutrient content, precipitation, global ice-volume). No proxy currently in use can claim a unique, direct association with a single climate variable; all have the potential to be influenced by processes other than changes in the climatic or oceanographic variable of direct interest. Some of the more common processes that complicate the interpretation of proxy variables include regeneration of biological material in the intermediate and deep waters, post-depositional diagenesis, erosion, and redeposition. In addition, some proxies are influenced by a number of climatic and/or oceanographic variables making it difficult to isolate the effect of any single variable. A good example is the marine δ18O record, which is influenced by temperature, salinity, and the isotopic composition of the ocean waters. Such complications have led to the use of a multi-proxy approach, which relies on evaluation of a number of independent proxies for a given variable. Each is physically, chemically, biologically, or isotopically linked to the variable of interest but impacted differently by unrelated processes such as diagenesis, dissolution, etc. Only the variance held in common among the proxies is attributed the variable of interest. This approach is most useful when the proxies are of sufficiently different origin (chemical, physical, biological, isotopic) and from an array of sites such that the variance not held in common is largely independent. In this work, five 350-k.y. records that fit these optimum criteria are combined using two different approaches, simple stacking and principal components analysis (PCA). The results are very similar. Relative to the five individual records, both provide much improved estimates of changes in monsoon strength over time and estimates of the relative concentrations of variance within the orbital periods.

2. Monsoon circulation and proxy response

Summer-monsoon circulation is characterized by atmospheric heating and the development of low pressure over Asia relative to higher pressure over the southern subtropical Indian Ocean (SSIO; Fig. 1). Two heat sources drive the Asian summer-monsoon low (Webster, 1987, 1994): sensible heating of the Asian landmass and condensational (latent) heating within the troposphere over Asia. The latent heat source derives from surface evaporation over the SSIO; latent heat is transported across the equator in low-level monsoon flow and released during precipitation as moisture-laden winds ascend the slopes of the Himalayas, thus enhancing the strength of the Asian low (Cadet and Reverdin, 1981; Cadet and Diehl, 1984; Cadet and Greco, 1987; Hastenrath and Lamb, 1979a,b, 1980; Hastenrath and Greischar, 1993; Liu et al., 1994; Webster, 1987, 1994). We make the case that this latent heat source is a critical factor in driving paleomonsoon variability.

The summer monsoon is marked by strong southwesterly winds over the Arabian Sea while cyclonic circulation about the Asian low creates northwesterly winds over the Arabian Peninsula. This circulation pattern typically lasts from mid-June through mid-September. The southwest flow over the northwest Arabian Sea drives strong coastal and open-ocean upwelling systems with consequent responses in a large array of oceanographic variables including sea surface temperature (SST), mixed layer depth (MLD), nutrient content, and productivity (Dickey et al., 1998;
Honjo et al., 1999; Smith, 1998, 1999, 2000; Weller et al., 1998). Cyclonic flow about the Asian monsoon low results in large amounts of dust entrainment and transport to the Arabian Sea, adding a lithogenic component to the biogenic material produced as a result of wind-induced upwelling (Clemens, 1998; Sirocko and Sarnthein, 1989; Tindale and Pease, 1998). Northeast winds develop over the continent and the Arabian Sea during winter as the continent becomes cool and high pressure develops over Asia. These north-easterlies are significantly weaker with relatively little dust transport and no large-scale upwelling response in the northwest Arabian Sea.

2.1. Seasonal sediment trap response

The seasonal response of several climate proxies during a single annual cycle is shown in Fig. 2. Detailed discussions of these data and their relationships to winter- and summer-monsoon dynamics have been previously published (Clemens, 1998; Honjo et al., 1999; Smith, 1998, 1999, 2000). Of relevance to this work is that all proxies have a strong seasonal response to summer-monsoon winds underscoring the dominance of summer-monsoon input to the underlying sediment record. Also of particular relevance is that the organic carbon and lithogenic flux, while highly correlated with summer-monsoon strength, cannot be interpreted as indicators of summer-monsoon variability over geological time scales. These two variables are not included in the summer-monsoon stack (SMS) or the summer-monsoon factor (SMF) for reasons that will be addressed in Section 4.

3. Stacking and principal component analysis

The five summer-monsoon proxies included in the SMS and SMF (Fig. 3) include physical (lithogenic grain size), chemical (barium mass accumulation rate; MAR), isotopic ($\delta^{15}N$), and both the carbonate and silica end members of biological production (abundance of *Globigerina bulloides* and opal MAR). Each proxy is linked to summer-monsoon variability through independent origins within the oceanographic and atmospheric systems. Each record has been previously pub-
lished with discussions of the extent to which it may be influenced by processes that serve to complicate its direct interpretation in terms of summer-monsoon variability.

Briefly, the nitrogen isotopic record (Altabet et al., 1995, 1999) is linked to summer-monsoon strength via export production and the intensity of denitrification within the oxygen minimum zone (OMZ). High export production associated with wind-driven upwelling strengthens the OMZ, thereby enhancing denitrification and increasing $\delta^{15}$N. While diagenesis is not a factor at this site, the $\delta^{15}$N record may be influenced by changing input of Persian Gulf and the Red Sea waters which have sill depths of 60 m and 137 m, respectively. Hence, changes in sea level associated with global ice-volume are a potential source of $\delta^{15}$N variability in the Arabian Sea. *Globigerina bulloides* is a planktonic foraminifer typically found in subpolar waters but is also abundant in tropical upwelling regions characterized by cold, nutrient-rich waters. The *G. bulloides* record is thus linked to monsoon strength through surface productivity driven by wind-induced upwelling (Anderson and Prell, 1993; Clemens et al., 1991; Curry et al., 1992; Naidu and Malmgren, 1996; Peterson et al., 1991; Prell, 1984a,b; Prell and Kutzbach, 1992). The biogenic opal record is similarly linked to monsoon strength through upwelling-induced production of diatoms and radiolaria (Haake et al., 1993; Honjo et al., 1999; Murray and Prell, 1992; Nair et al., 1989; Prell et al., 1992). Processes other than those associated directly with monsoon variability impact both biogenic records. The *G. bulloides* record can be influenced by carbonate dissolution. In addition, it has a significant linear increase toward the mod-

Fig. 2. Sediment trap and meteorological data (normalized) from the JGOFS Arabian Sea project. Data are the average of mid-depth traps (~2000 m) at mooring locations 2, 3, and 4. Moorings 2 and 3 are inshore from cores ODP 722 and RC27-61 while mooring 4 is offshore. The meteorological data are from a buoy deployed near mooring 4. The dashed line in the grain size plot indicates a gap in data collection due to trap turn around. The three vertical dashed lines indicate the onset, maximum, and end of the summer-monsoon season as defined by the meteorological variables.
ern which is not present in the other records and thus has been removed prior to stacking and PCA. The opal record contains distinct intervals of low variability indicative of silica dissolution (e.g. 125–175 ka). The lithogenic grain size record is linked to the transport capacity of the summer northwest winds associated with cyclonic flow about the Asian low (Clemens and Prell, 1990; Clemens, 1998; Clemens et al., 1991, 1996). While the grain size record does not suffer the effects of dissolution, it is impacted by changes in vegetation cover in the source areas. For example, the

Fig. 3. Normalized Arabian Sea summer-monsoon proxies. The top five plots show the Arabian Sea SMS (dashed line) plotted with the five individual proxy records which make up the SMS (solid lines). The bottom plot shows the SMS (dashed line) relative to the Arabian Sea SMF (solid line) derived from principal components analysis. A linear trend was removed from the *Glo-bigerina bulloides* record prior to incorporation in the SMS and SMF. The SMF starts slightly later and ends slightly earlier than the SMS due to the way the two procedures handle differing record lengths. Age models for cores RC27-61 and 722B can be found in Clemens and Prell (1991) and Murray and Prell (1992). Four exceptionally large peaks have been truncated for the purposes of plotting this figure, two in the Ba record (~10 and 123 ka) and two in the Opal record (9 and 255 ka). The data for the SMS and SMF are archived at [http://www.ngdc.noaa.gov/paleo/paleo.html](http://www.ngdc.noaa.gov/paleo/paleo.html).
four other records indicate strong monsoons during the early Holocene whereas the grain size record shows a distinct low. This is attributed to increased regional vegetation cover in the dust source areas during the early Holocene pluvial. Sedimentary barium is mechanistically linked to biogenic particles in the surface waters and to export productivity (Francois et al., 1995). Thus, the excess Ba accumulation record is tied to monsoon strength through productivity associated with wind-induced upwelling (Clemens et al., 1991; Shimmield et al., 1990). Though these chemical and biogenic indicators all are associated with productivity, the dissolution processes associated with each (carbonate, silica, and barium) are largely independent.

Visual comparison of the five records shows that the timing of events is similar, indicating a common link to monsoon strength (Fig. 3). Hence the individual proxies yield similar phase responses at the 23- and 41-k.y. periods (Altabet et al., 1995, 1999; Clemens et al., 1991). However, the relative amplitude of monsoon events is quite different. For example, the largest events are indicated at ∼50- and 255-ka in the Globigerina bulloides record, at ∼214- and 319-ka in the grain size record, and at ∼124-ka in the Ba record. These amplitude differences, in part, reflect the differing impacts of dissolution, preservation, and source area changes on these chemical, biological and physical monsoon indicators. These amplitude differences lead to differences in the spectra of the individual proxies (Altabet et al., 1995, 1999; Clemens et al., 1991). Thus, the phase response of the monsoon system can be determined from individual proxies, but the relative strength of monsoon events through time and the relative concentration of variance in the orbital bands cannot. To evaluate these aspects of the paleomonsoon one must assess the variance in common among all the five proxies. This can be approached by simple stacking of the records or by PCA.

Stacking was accomplished by interpolating each record to a common 2-k.y. sample interval (similar to the resolution of the individual records), normalizing each to unit variance, and averaging. The result was cross checked by PCA of the five summer-monsoon records using CABFAC (Klovan and Imbrie, 1971) (Fig. 3). Similar approaches have been used to evaluate the variance in common among multiple carbonate dissolution proxies (Peterson and Prell, 1985; Thunell, 1976). Factor 1 (unrotated) of the summer-monsoon proxies (the SMF) indicates that only 33% of the total variance in the five records is

<table>
<thead>
<tr>
<th>Proxy record</th>
<th>δ¹⁸O-41 coh</th>
<th>δ¹⁸O-41 phase</th>
<th>Tilt phase</th>
<th>δ¹⁸O-23 coh</th>
<th>δ¹⁸O-23 phase</th>
<th>P phase</th>
</tr>
</thead>
<tbody>
<tr>
<td>δ¹⁸O 722</td>
<td>0.95</td>
<td>7 ± 16°</td>
<td>0.98</td>
<td>−11 ± 10°</td>
<td>−11 ± 10°</td>
<td>0°</td>
</tr>
<tr>
<td>δ¹⁸O RC27-61</td>
<td>0.98</td>
<td>−3 ± 10°</td>
<td>0.99</td>
<td>7 ± 6°</td>
<td>7 ± 6°</td>
<td>0°</td>
</tr>
<tr>
<td>δ¹⁸O SPECMAP</td>
<td>1.00</td>
<td>0 ± 0°</td>
<td>−69°‡</td>
<td>1.00</td>
<td>1.00</td>
<td>0°</td>
</tr>
<tr>
<td>Summer Monsoon Stack</td>
<td>0.84</td>
<td>57 ± 28°</td>
<td>−12°</td>
<td>0.96</td>
<td>0.96</td>
<td>−12°</td>
</tr>
<tr>
<td>Summer Monsoon Factor</td>
<td>0.86</td>
<td>63 ± 18°</td>
<td>−6°</td>
<td>0.97</td>
<td>0.97</td>
<td>−12°</td>
</tr>
<tr>
<td>RC17-98 SST (cold)</td>
<td>0.71</td>
<td>68 ± 29°</td>
<td>−1°</td>
<td>0.82</td>
<td>0.82</td>
<td>−12°</td>
</tr>
<tr>
<td>NIOP 464 average</td>
<td>0.91</td>
<td>32 ± 73°</td>
<td>−37°</td>
<td>0.92</td>
<td>0.92</td>
<td>−12°</td>
</tr>
<tr>
<td>RC27-61 Corg MAR</td>
<td>0.92</td>
<td>138 ± 21°</td>
<td>69°</td>
<td>0.81</td>
<td>0.81</td>
<td>170°‡</td>
</tr>
<tr>
<td>RC2761 Lith. MAR</td>
<td>0.92</td>
<td>−168 ± 20°</td>
<td>123°</td>
<td>0.91</td>
<td>0.91</td>
<td>172 ± 2°</td>
</tr>
</tbody>
</table>

In all cases the phase of δ¹⁸O has been multiplied by −1 such that large values correspond to ice minima.

b Test statistics for non-zero coherence at the 80% and 95% confidence intervals are 0.79 and 0.91, respectively.

c Phase set by hypothesis (Imbrie et al., 1984).

d Phase calculated by application of the SPECMAP-defined phase lags of −69° and −78° for tilt and precession, respectively (Imbrie et al., 1984). Phase error is the same as in column to the left.

e Average composed of δ¹⁵N, Ba/Al, and Globigerina bulloides relative to δ¹⁸O from the same core (Reichart et al., 1998).
shared in common (i.e. related to summer-monsoon strength). Viewed from another perspective, this indicates that 67% of the total variance is linked to non-monsoon processes (e.g. dissolution, preservation, sediment reworking), highlighting the utility of the multi-proxy approach.

The plot of factor 1 loadings (the SMF) is very similar to the SMS in terms of the timing and relative amplitude of events. The only exceptions are increased amplitude in the SMS at 9- and 255-ka which are the result of spikes in opal accumulation at these times; simple averaging emphasizes them whereas factor analysis does not. In this respect, the SMF plot is a better representation of relative amplitude through time. Visually, orbital-scale variability is concentrated in the higher-frequency precession and obliquity bands, with little variance at the 100 k.y. glacial-interglacial scale. In fact, equally strong monsoon events occur within glacial stages (MIS 6 and 8) and interglacial stages (Fig. 4).

3.1. Spectral results

Cross-spectral analysis allows the quantification

![Fig. 5. Cross-spectral relationships between the SMS and (a) insolation at 30° north for the month of June, when insolation is greatest in the region of the Asian monsoon low, (b) ETP which combines the normalized eccentricity, obliquity (tilt), and precession parameters from Laskar et al. (1993) into a single record, convenient for time series analysis, (c) marine δ^18O using the SPECMAP stack of Imbrie et al. (1984) which was used to create the age models for the two cores in the SMS, and (d) SMF. Periods (1/frequency) of the main spectral peaks (66, 41, 23, and 16 k.y.) are labeled in (d). The 41-k.y. peak in the SMS (SMF) contains 26% (27%) of the total variance. Similarly, the 23-k.y. peak contains 18% (17%) of the total variance, the 66-k.y. peak contains 11% (13%) of the total variance and the 16-k.y. peak contains 8% (12%) of the total variance in the SMS (SMF) records. Spectral densities are normalized and plotted on log scales. The coherence spectra are plotted on a hyperbolic arctangent scale. The horizontal solid lines indicate confidence at the 80% and 95% levels.](image-url)
of coherence and phase relationships among variables at specific periods, in this case, Earth-orbital periods. Coherence defines the extent to which two variables are linearly related while the phase indicates temporal leads and lags critical to understanding forcing/response relationships. Coherence and phase relationships between the SMS and Northern Hemisphere (NH) summer insolation, ETP, $^{18}$O, and the SMF are presented in Fig. 5 and Table 1. The cross-spectral relationships discussed below are based on the SMS. The results would be no different were they based on the SMF record; the two are coherent with one another well above the 95% confidence limit (Fig. 5d) and in phase.

### 3.1.1. Variance

The spectrum of the SMS indicates that the amount of variance in the obliquity band (26%) is slightly greater than that in the precession band ($\sim$18%). The radiation spectrum, however, is clearly dominated by variance in the precession band (Fig. 5a). Thus, more monsoon variance resides in the obliquity band than can be accounted for by direct insolation forcing. This is an important observation in that most discussion of summer monsoon variability in the scientific literature has been focused on the precession response. An additional 22% of the total variance resides in peaks at the 16- and 66-k.y. periods (Fig. 5d), possibly representing heterodynes asso-
associated with beating of obliquity and precession. These non-primary peaks are better defined in the SMF spectrum compared to the SMS spectrum.

3.1.2. Coherence

Cross-spectra indicate that the summer monsoon is strongly coherent with NH summer radiation and ETP at the 41-k.y. and 23-k.y. periods (Fig. 5a,b). However, the summer monsoon is much more coherent with $\delta^{18}$O at the 23-k.y. period than at the 41-k.y. period (Fig. 5c). Coherence between the summer monsoon and $\delta^{18}$O at the precession and obliquity periods may seem odd given that the summer monsoon (SMS and SMF) has no variance peak at the 100-k.y. period, where the majority of ice-volume variance resides. Phase relationships, discussed below, provide some insight into this issue.

3.1.3. Phase

Phase results are displayed using phase wheels (Table 1; Fig. 6). Note that the phases reported in Table 1 are calculated relative to $\delta^{18}$O, not ETP. Direct comparison of phase results measured relative to ETP from different cores is valid only if the isotopic age models are comparable. While most isotopic age models in the 0–350-k.y. interval use the SPECMAP age model of Imbrie et al. (1984), some investigators use slightly modified ages for some isotopic events. Of relevance to this work, Reichart et al. (1998) use an isotopic age model which yields phase lags (relative to ETP) that are decreased by 2 k.y. at the precession band and 3 k.y. at the obliquity band compared to those of the SMS, which is based on the SPECMAP age model. By calculating phases relative to $\delta^{18}$O from the same core, the phases of proxies from different cores can be directly compared. Age models for the data presented here were developed by $\delta^{18}$O correlation to the SPECMAP stack so it is used in the cross-spectra with the SMS.

At the obliquity band, strong monsoons are in

Fig. 7. Time series and cross-spectra for organic carbon (C$_{org}$) mass accumulation rate (MAR) and lithogenic MAR. Unlike the Arabian Sea SMS and SMF, these two variables show very strong 100-k.y. cyclicity which is coherent with $\delta^{18}$O over all orbital bands. See caption to Fig. 5 for details of the cross-spectra.
phase with NH summer insolation maxima but lead ice minima by 6.5 k.y. (57°/360°×41 k.y.; Table 1). At the precession band, strong monsoons lag NH summer insolation maxima (June 21 perihelion) by 8 k.y. (−121°/360°×23 k.y.) and ice minima by 3 k.y. (−43°/360°×23 k.y.).

4. Discussion

4.1. Lithogenic and organic carbon accumulation

Before discussing the variability of the summer monsoon, the exclusion of the organic carbon and lithogenic accumulation records from the SMS and SMF needs to be addressed. Both the lithogenic and organic carbon records are highly correlated with summer monsoon winds on the seasonal time scale as was previously discussed (Fig. 2). However, unlike the summer-monsoon proxies, lithogenic and organic carbon accumulation are dominated by 100-k.y. cyclicity suggesting that changes in climate associated with large-scale glacial-interglacial variability strongly influence these proxies (Fig. 7). Phase relationships further differentiate them from the summer-monsoon proxies, indicating that they are not useful indicators of paleomonsoon strength, at least not at the Owen Ridge sites. All five of the individual summer-monsoon proxies have near-zero phase relationships with one another at both the 41- and 23-k.y. periods (Altabet et al., 1995, 1999; Clemens et al., 1991). The only exception is the δ15N record at the 41-k.y. period, which lags maximum obliquity by 48 ± 18°. The carbon and lithogenic flux records are distinctly out of phase with phase with the summer monsoon and with one another indicating that they are not responding directly to summer-monsoon forcing (Table 1; Fig. 6). As discussed by Murray and Prell (1992), the phase of the organic carbon signal resides halfway between strong monsoons (high biogenic productivity) and high sedimentation rates associated with lithogenic flux (enhanced preservation). Thus, organic carbon is a slave to two masters and a direct indicator of neither. High lithogenic accumulation is in phase with maximum ice-volume (Clemens and Prell, 1990) as is found in many globally distributed dust records (Rea, 1994). This record is driven by global changes in aridity associated

### Table 2

<table>
<thead>
<tr>
<th>Orbital configuration</th>
<th>Month</th>
<th>Ocean/atmosphere response</th>
<th>Impact on monsoon</th>
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<tbody>
<tr>
<td><strong>Precession maxima</strong></td>
<td></td>
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<td></td>
</tr>
<tr>
<td>Warmest SH summer</td>
<td>June</td>
<td>Strengthened Asian low</td>
<td>Stronger monsoon</td>
</tr>
<tr>
<td>Warmest SH winter</td>
<td>June</td>
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<td>Late onset, shorter season</td>
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with glacial–interglacial cyclicity, not by changes in regional monsoon strength. These examples serve to highlight the utility of the multi-proxy approach. Interpretation of the lithogenic and organic carbon records as monsoon proxies would yield vastly different and incorrect interpretation of mechanisms driving monsoon variability at orbital time scales.

4.2. Indian summer monsoon

Insolation and global ice-volume are the two most obvious mechanisms driving late Pleistocene climate change at orbital time scales. A-priori, one expects stronger summer monsoons to be associated with the strongest NH summer insolation and with minimum ice-volume. These expectations are supported by climate model sensitivity experiments (deMenocal and Rind, 1993; Kutzbach et al., 1996; Kutzbach, 1981, 1983; Kutzbach and Otto-Bliesner, 1982; Prell and Kutzbach, 1992). However, two important lines of evidence from the SMS and SMF indicate that the summer monsoon does not respond directly to either of these factors. The greater amount of variance in the obliquity band compared to the precession band as well as the $3^{121\text{‡}}$ lag relative to June 21 perihelion are inconsistent with a direct response to NH summer insolation forcing. The absence of a 100-k.y. variance peak in the summer-monsoon spectrum is inconsistent with a direct response to large-scale changes in ice-vol-
ume. Taken together, these lines of evidence imply that the summer monsoon is responding to some additional factor(s). Phase analysis points to a Southern Hemisphere (SH) latent heat source as a primary factor in driving monsoon variability.

4.2.1. Precession band forcing

For precession, the strongest summer-monsoons occur 8 k.y. (~121°) after maximum NH summer insolation and 3 k.y. after ice-volume minima (Table 1; Fig. 6). A simple linear forcing and response system cannot accommodate phase lags over 90°. Thus, the timing of the strongest monsoons is not a direct response to NH summer insolation forcing associated with June 21 perihelion. This large precession-based phase lag is well-established in the literature, not only from Owen Ridge records included in the SMS and SMF (Altabet et al., 1999; Altabet et al., 1995; Clemens and Prell, 1990; Clemens et al., 1991; Shimmield et al., 1990), but also from work on the Murray Ridge, northern Arabian Sea (Reichart et al., 1998) and off Japan in the northwest Pacific (Morley and Heusser, 1997). Using four monsoon proxies established from the 225-k.y. record of core NIOP 464, Reichart et al. (1998) calculate a lag of 6 k.y. relative to NH summer insolation maxima (8 k.y. when converted to the SPECMAP time scale; Table 1). Morley and Heusser (1997) report an 8-k.y. lag based on a 350-k.y. pollen record in core RC14-99 off Japan. Both results are consistent with our estimate of 8 k.y., regardless of the age model employed. While the existence of the lag is firmly established, its interpretation is debated.

Clemens et al. (1991, 1996) argue that the timing of strengthened monsoons is partly governed by the timing of latent heat export from the SSIO. They observed that cold SST in the SSIO (~13° south) is in phase with monsoon maxima at the precession band (Table 1; Fig. 6), a relationship consistent with modern dynamics relating latent heat export, SST, and monsoon strength. Reichart et al. (1998) argue against the latent heat hypothesis using evidence from well-dated Holocene sections in India and Tibet which indicate a precipitation maximum from 10 to 6 ka. They argue that the latent heat hypothesis implies that a summer-monsoon maximum should exist at 3.5 ka (~8 k.y. after the last precession minimum at 11.5 ka), considerably later than the observed maximum in India and Tibet. This reasoning would be valid if all the variance in the monsoon were associated with precession alone. However, the large amount of variance in the monsoon band must also be taken into account. The last obliquity maximum (also latent heat maximum as discussed in detail near the end of this section) was at 9.5 ka. Combined, the latent heat export associated with precession and obliquity would predict strengthened summer monsoons between 9.5 and 3.5 ka, in agreement with the terrestrial data from India and Tibet. The relatively low-resolution isotopic age model of the SMS and SMF indicates a peak from 8 to 6 ka, consistent with the latent heat hypothesis and with the timing of the Holocene rainfall maximum in India and Tibet. High resolution, AMS-dated Globigerina bulloides records from the Northern Arabian Sea indicate a broad monsoon maximum between 11 and 5 calendar ka (Naidu and Malmgren, 1996; Overpeck et al., 1996), in agreement with the terrestrial dates and with the timing of strong Holocene monsoons predicted by the latent heat hypothesis.

Reichart et al. (1998) further suggest that the large precession-based lag (~121°) can be reduced by calling on August or September insolation forcing instead of the maximum NH summer insolation (June 21 perihelion). This is based on the observation that the maximum in seasonal biological productivity measured in sediment traps occurs in August and September, seven to eight weeks after the onset of the summer monsoon. Consequently, they suggest that intervals of increased late-summer insolation result in longer summer-monsoon seasons and hence increased productivity. This lagged biological response is illustrated in Fig. 2 where all biological indicators show a consistent 7–8-week lag relative to summer monsoon onset and increased lithogenic grain size (Clemens, 1998). This seasonal biogenic lag is a function of the response time of the ocean in terms of surface circulation and thermal inertia as well as the response times of biological communities to nutrient supply, mixed layer depth,
reproduction rates, and predation rates (Bartolacci and Luther, 1999; Dickey et al., 1998; McCreary et al., 2001; Prahl et al., 2000; Roman et al., 2000; Weller et al., 1998). Thus, the late timing of the maximum biological flux within the seasonal cycle is not a synchronous response to late summer insolation forcing and should not be extrapolated to processes operating on orbital time scales.

From a strictly NH summer insolation perspective, reducing the $-121^\circ$ lag to zero would require a direct response to October 21 insolation forcing (for precession, the lag is decreased by 30$^\circ$ for each month following June). Using the Reichart and others age model, the $-121^\circ$ lag becomes $-94^\circ$ and would require a direct response to September 21 insolation forcing. Fig. 8 illustrates that, over the last 350 k.y., that the maximum insolation in September is lower than the minimum insolation in June. Thus, the weak late-summer insolation is unlikely to be directly responsible for the large phase lag. Finally, in considering the issue of longer vs. stronger monsoons from the standpoint of NH summer insolation forcing, one should also consider increased insolation in the months prior to June; increased spring insolation. Conceptually this would melt winter snows earlier leading to early sensible heating and early monsoon onset. To evaluate this, one can generate records of the total annual insolation received for all days in which NH insolation exceeds a given threshold. Fig. 9 shows the record of cumulative insolation for all days with radiation greater than 350 watts m$^{-2}$, consistent with the late summer hypothesis of Reichart et al. The frequency, amplitude modulation, and phase are the same as the June 21 insolation record. Similar results are found for any chosen threshold because the insolation curves are arrayed symmetrically about the June 21 insolation maximum. For these reasons June 21 perihelion remains the most appropriate for measuring the monsoon phase response relative to NH summer insolation forcing. However, the large precession phase lag itself indicates that SH dynamics influence the timing of strong monsoons within the precession band. The large phase lag in the precession band and the amplified monsoon response in the obliquity band can both be explained in the context of the latent heat hypothesis.

Table 2 illustrates the impact of various orbital configurations on the seasonal cycle of radiation in the NH and SH. For the precession cycle, the $-121^\circ$ phase falls closer to December 21 perihelion ($-180^\circ$) than to June 21 perihelion ($0^\circ$). The seasonal cycle at December 21 perihelion is characterized by relatively hot SH summers and cold SH winters. The hot SH summer (December–February) results in increased energy storage in the SH ocean during the months prior to the onset of the NH summer monsoon in June. This preconditioning the ocean for increased latent heat export. Decreased insolation during the SH winter (June–August) results in cold SH air temperatures during the boreal monsoon months. Thus, the orbital configuration at December 21 perihelion results in cold, dry air over warm SH ocean waters, maximizing the ocean-to-atmosphere transfer of latent heat available for transport and release over Asia during the NH summer monsoon. December 21 perihelion is also characterized by relatively warm NH winters. This would presumably decrease snowfall accumulation, resulting in an early melting and a longer monsoon season. On the other hand, the relatively cold NH summers associated with December 21 perihelion would serve to weaken the Asian low from a sensible heating standpoint, acting in opposition to the increased latent heat export. Relationships for June 21 perihelion are the opposite, resulting in stronger Asian lows but shorter seasons due to colder than average winters and decreased latent heat export (Table 2). From an insolation standpoint, the $-121^\circ$ phase indicates that the monsoon is more sensitive to the latent heat dynamics associated with December 21 perihelion ($-180^\circ$) than to direct sensible heating of the Asian Plateau associated with June 21 perihelion ($0^\circ$). If no other factors were involved, the equilibrium monsoon phase would presumably be $-180^\circ$. However, as suggested by the high coherence with $\delta^{18}O$, the summer-monsoon is also sensitive to changes in glacial boundary conditions at the precession period. In this case, changes in ice-volume are interpreted as modulating the timing of maximum sensible heating over Asia, delaying it until ice-
minima at $-78^\circ$. Over the course of the precession cycle, maximum monsoon strength ($-121^\circ$) is located between these two primary forcings; maximum sensible heating over Asia ($-78^\circ$ due to the ice-volume delay) and maximum latent heat export from the SSIO ($-180^\circ$).

Within the precession band, the measured phase of cold SST from core RC17-98 (13°S, 65°E) is $-123^\circ$, not $-180^\circ$ as would be consistent with insolation forcing alone. Cold SST is, however, in phase with maximum monsoon strength (Clemens and Prell, 1990; Clemens et al., 1991). The $-123^\circ$ phase of the cold SST likely reflects a positive feedback between monsoon strength and evaporative cooling of the sea surface in the SSIO.

4.2.2. Obliquity band forcing

Within error, monsoon maxima are in phase with obliquity maxima (Table 1; Fig. 6). Obliquity maxima are characterized by warm NH summers leading to stronger Asian lows (Table 2). Thus, monsoon strength in the obliquity band is strongly related to NH summer insolation. However, Fig. 5a indicates a greater concentration of monsoon variance in the obliquity band than would be consistent with direct radiative forcing alone; most radiation variance resides in the precession band. This suggests that the monsoon response at the obliquity band is amplified by feedbacks internal to the climate system. The additional variance can be accounted for by latent heat export. Like December 21 perihelion in the precession band, obliquity maxima are characterized by insolation maxima during SH summer and insolation minima during SH winter; ideal for maximizing latent heat export (ocean to atmosphere) during the NH monsoon season.

In summary, NH summer insolation and SH latent heat export work in opposition to one another at the precession band (out of phase) but with one another at the obliquity band (in phase). This may explain why there is more monsoon variance in the obliquity band than in the precession band and at the same time indicates that the timing of strong summer monsoons is more sensitive to latent heat export than to external insolation forcing.

4.3. East Asian winter and summer monsoon

The Arabian Sea SMS and SMF are useful references for analyzing late Pleistocene records of East Asian monsoon variability. In particular, records from the South China Sea are currently being generated from sediments recovered during Ocean Drilling Program Leg 184 as well as previous coring efforts (Sarnthein and Wang, 1996; Wang et al., 2000). Due to the geographic location of the South China Sea, the ocean and atmospheric responses to monsoon dynamics should yield sediment records of both summer- and winter-monsoon variability (Wang et al., 2000). Conceptually, if the combined effects of insolation and glacial boundary conditions dominate winter monsoon variability, the phase response should lie between minima in NH winter insolation and ice maxima. Using the SPECMAP time scale, these phases would lie between zero and $+111^\circ$ on the obliquity phase wheel of Fig. 6 and between zero and $+102^\circ$ on the precession phase wheel of Fig. 6.

The $-121^\circ$ precession phase of summer-monsoon variables from the Arabian Sea and off Japan suggests that the Indian and East Asian summer monsoons are coupled, presumably by the strength of the Asian low as influenced by latent heating and ice-volume. If so, summer monsoon proxies from the South China Sea should also have a $-121^\circ$ phase at precession band. However, if the East Asian monsoon is more sensitive to direct NH summer insolation and glacial conditions then one would expect the phases to lie between NH summer insolation maxima ($0^\circ$) and ice minima ($-69^\circ$) on the obliquity phase wheel of Fig. 6 and between NH summer insolation maxima ($0^\circ$) and ice minima ($-78^\circ$) on the precession phase wheel of Fig. 6.

5. Summary

(1) Simple stacking and PCA are used to combine five individual summer monsoon proxies into single 350-k.y. time series that integrate the biological, isotopic, geochemical, and physical response of the ocean and atmosphere to summer-
monsoon forcing. Stacking and PCA remove non-monsoon variance (e.g. dissolution, diagenesis, and preservation) associated with each individual proxy resulting in more pure records of summer-monsoon variability. The SMS and SMF records are interchangeable in terms of spectral characteristics (coherence and phase). They are also very similar in time series form although the SMF record is a better estimate of changing amplitude through time.

(2) Approximately 26% of the total variance resides in the obliquity band and 18% in the precession band. An additional 22% of the total variance is distributed among non-primary orbital peaks at 66- and 16-k.y. periods, possibly associated with beating of precession and obliquity. Greater variance in the obliquity band than in the precession band is not consistent with a direct response to external insolation forcing which would be dominated by variance in the precession band. Further, monsoon maxima lag NH summer insolation maxima by 121°, inconsistent with a direct response to NH summer insolation forcing. These results indicate that monsoon strength and timing are highly sensitive to internal forcing mechanisms.

(3) Within the precession band, monsoon maxima lag NH summer insolation maxima (June 21 perihelion) by 121°. This −121° phase falls between δ18O minima (−78°) and December 21 perihelion (−180°). This is interpreted to indicate that the timing of the summer-monsoon is influenced by two mechanisms: (1) sensible heating of the Asian Plateau as modulated by ice-volume such that maximum sensible heating is delayed until ice minimum (−78°), and (2) latent heating within the troposphere over the Asian Plateau which reaches a maximum at December 21 perihelion (−180°), an orbital configuration that maximizes latent heat export from the SSIO and its transport into Asia.

(4) Within the obliquity band, monsoon maxima are in phase with obliquity maxima. Like the precession band, obliquity maxima are characterized by maximum sensible heating of the Asian Plateau (with no ice-volume delay) and by maximum latent export from the SSIO and its transport into Asia.

(5) NH summer insolation and SH latent heat export work in opposition to one another at the precession band (180° out of phase) but with one another at the obliquity band (in phase). This may explain why there is more monsoon variance in the obliquity band than in the precession band. For both the precession and obliquity cycles, monsoon maxima are associated with orbital configurations which maximize latent heat export (ocean to atmosphere) indicating that the summer monsoon is more sensitive to the export and cross-equatorial transport of latent heat than to external insolation forcing over the Asian Plateau.

(6) The summer-monsoon has little variance and low coherence with δ18O at the 100-k.y. eccentricity cycle associated with large-scale changes in global ice-volume. Indeed, equally strong monsoons occur in both glacial (e.g. MIS 6 and 8) as well as interglacial stages.

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