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Extending the historical record by proxy

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16.1 INTRODUCTION

Instrumental records of meteorological and oceanographic variables critical to understanding monsoon variability extend back in time only a century or so. These historical data allow evaluation of seasonal to interdecadal variability in the Asian monsoon system. Extending the historical record through development of terrestrial and marine-based proxies is a useful means of better understanding monsoon variability across all timescales – seasonal to tectonic. The utility of developing monsoon records from geological archives lies in the fact that monsoon variability during older intervals takes place in the context of boundary conditions very different from the modern system. For example, Earth-orbital variations lead to cyclical changes in the amount of radiation received at a given latitude over time (Figure 16.1(a)). In addition to climate change driven by these orbital radiation cycles, the Earth has undergone large, and currently unexplained, changes in the extent of its high-latitude ice sheets; changes of such magnitude that global sea level ranged from +6 m to −120 m relative to the present day within the last few hundred thousand years (Figure 16.1(b)). On the same timescales occurred large changes in the concentration of atmospheric greenhouse gasses such as carbon dioxide and methane as well as atmospheric dust content (Petit *et al.*, 1999; Rea, 1994) indicating a very close coupling between the atmosphere, hydrosphere, cryosphere, and biosphere. Superimposed on the long-term orbital and glacial–interglacial climate cycles are abrupt climate events which occur on centennial to millennial timescales (Figure 16.2(a)). Evaluating the role of the monsoon in the context of climate change on these differing timescales provides important information on how the ocean, atmosphere, and terrestrial components of the Earth's climate system interact as a function of changing boundary conditions.

Monsoon circulation is defined on the basis of seasonal changes in regional patterns of wind and precipitation. These primary meteorological variables are

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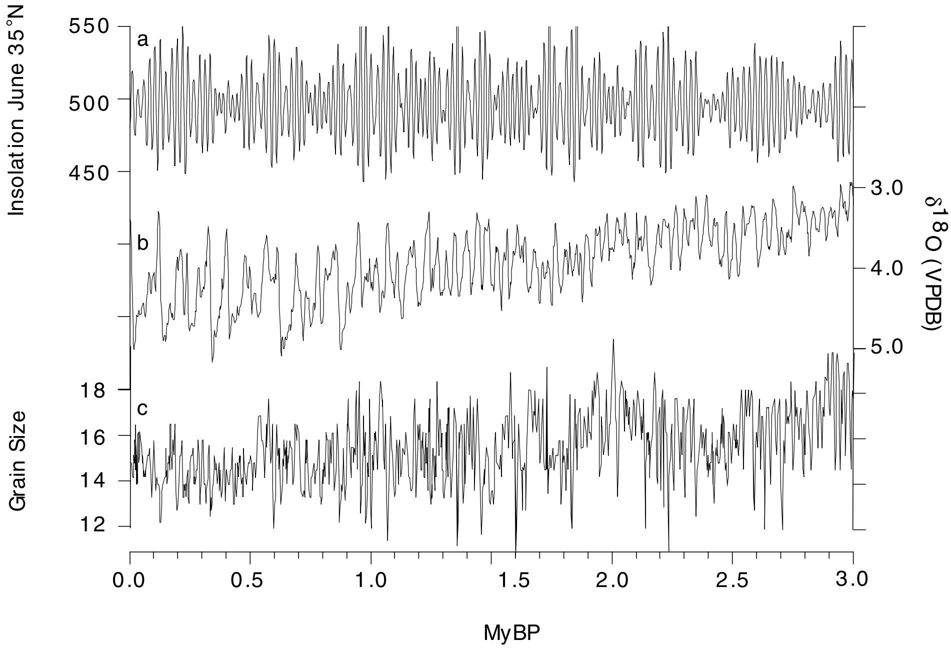

Figure 16.1. Climate change at orbital timescales; external forcing, internal boundary conditions, and monsoon climate change over the past three million years. (a) Time series of incoming solar radiation at 35°N in June (Laskar *et al.*, 1993); (b) the marine oxygen isotopic record of benthic foraminifera, a proxy for changes in global ice volume and sea level; and (c) lithogenic grain size, a proxy for changes in Indian summer monsoon wind strength (Clemens, 1998; Clemens *et al.*, 1996). Radiation variability driven by changes in the precession and obliquity of the Earth's orbit about the Sun is the primary external forcing for climate change at the 10^4 to 10^5-year timescale. Large-scale change in the amount of water stored as high-latitude ice (the Pleistocene ice ages) is a fundamental internal boundary condition associated with climate change on this timescale. Within the time frame of the past 3 million years, the oxygen isotopic ($\delta^{18}O$) composition of benthic foraminifera are an excellent proxy for changing terrestrial ice volume; the maximum amplitude in the younger part of the record corresponds to $\sim$126 m of sea level change (high values correspond to increased terrestrial ice and low sea level). The grain size of lithogenic material transported to the Arabian Sea is a proxy for changes in the strength of Indian summer monsoon winds; large grains correspond to increased circulation. The $\delta^{18}O$ record is a composite of three records from the equatorial Pacific Ocean, Vema 19–30 (Shackleton and Pisias, 1985), Ocean Drilling Program (ODP) Site 677 (Shackleton *et al.*, 1990), and ODP 846 (Shackleton *et al.*, 1995b). Isotope data available from *http://delphi.esc.cam.ac.uk/*.

not directly preserved in the geological record. However, they greatly influence the physics, chemistry, and biology of the associated ocean and land surface processes, aspects of which are readily preserved in geological archives such as deep-sea sediments, terrestrial sediments, corals, lakes, bogs, cave deposits, tree rings, and glaciers. As such, these geological archives preserve indirect, or proxy, measures of

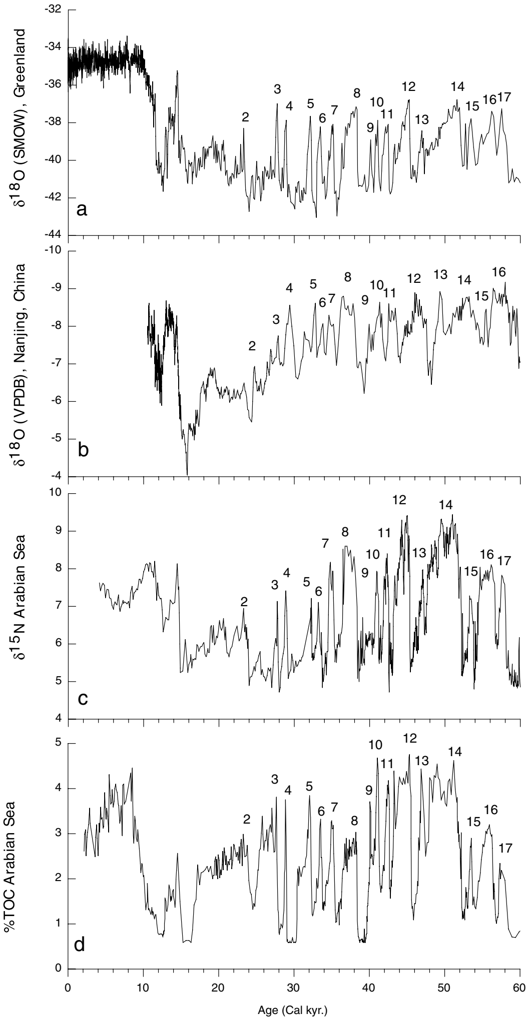

Figure 16.2. Climate change at millennial timescales. Superimposed on longer timescale climate change (Figure 16.1) are abrupt changes that occur at the centennial to millennial timescale. At the millennial timescale climate variability is thought to be largely internal to the climate system; there is no known source of solar variability at the millennial timescale. (a) The Greenland Ice Sheet Project 2 (GISP2) δ^{18}O record, a proxy for temperature change over Greenland (Dansgaard *et al.*, 1993). (b) Hulu Cave δ^{18}O, a proxy for Asian summer monsoon strength as it impacts the oxygen isotopic composition of stalagmite calcite near Nanjing, China (Wang *et al.*, 2001b). (c) Total organic carbon, a proxy for the strength of the Indian summer monsoon as it impacts the strength of the oxygen minimum zone on the Pakistan Margin (Schulz *et al.*, 1998). (d) δ^{18}O^{15}N, a proxy for the strength of the Indian summer monsoon as it impacts the productivity driven denitrification on the Oman Margin (Altabet *et al.*, 2002). Within the error of estimated ages, the numbered interstadial (warm) events are thought to be coincident, indicating strengthened Asian summer monsoons in association with abrupt warming events in the northern Atlantic region.

past monsoon variability. Depending on the proxy, changes in monsoon climate can be resolved on timescales ranging from seasonal (e.g., laminated sediment sections, corals, snow pits) to tectonic (e.g., marine and terrestrial sediment sections). Here we discuss paleomonsoon history at centennial (10^2 years) to tectonic (10^6 years) timescales. In terms of paleoclimate signals, monsoon circulation is exceptionally strong, often leaving a clear and dominant signal in the geological record. However, additional processes, sometimes unrelated to monsoon variability, may also influence the chemical, physical, and biological composition of these geological archives. For example, diagenesis, a post-depositional chemical and/or physical alteration of the archive, may take place under certain conditions, masking the monsoon signal. Such alterations are not always readily evident. For these reasons, paleoclimatologists often employ a multiproxy approach to evaluating climate variability preserved in the geological record (e.g., Clemens and Prell, 2003). The strategy is to develop a number of different proxies; each is physically, chemically, or biologically linked to changes in monsoon circulation but impacted differently by non-monsoon processes. Statistical techniques are then employed to extract the variance common to all the proxies, suppressing the variance unrelated to monsoon circulation. The common variance is taken as the best estimate of monsoon variability.

Another aspect of primary importance in evaluating paleomonsoon variability from geological archives is the development of accurate age models (chronostratigraphies) such that the time history of climate change recorded at different geographic locations can be compared with confidence. This is critical for evaluating temporal leads and lags in the system documenting how the monsoon initiates or responds to changes in other parts of the climate system. Sufficiently accurate chronostratigraphies are generally available for studies at the tectonic and orbital-scale. Sufficiently accurate chronostratigraphies are generally lacking for studies of abrupt climate change on millennial and finer scales although significant progress is being made in this area and a great deal of useful information is derived from the geological record in spite of current chronostratigraphic limitations (Sarnthein *et al.*, 2002).

The rapid increase in global-scale, instrument-based meteorological and oceanographic data has fueled tremendous advances in the understanding of modern monsoon variability and interactions among the various monsoon subsystems. However, the spatial coverage of paleomonsoon records is limited (Voelker, 2002). Thus, from a paleomonsoon perspective, the subsystems within the Asian monsoon have largely been studied independently both in terms of geographic variability and timescales of variability. While the number of high-quality paleomonsoon records is increasing, the coverage remains insufficient to fully assess regional differences in paleomonsoon circulation.

16.2 INTERNAL AND EXTERNAL FORCING

Potential mechanisms driving changes in the ancient monsoon can be broadly categorized as 'internal' or 'external'. External forcing is solar in origin, including insolation cycles associated with eccentricity (100,000-year cycle), obliquity (41,000-year cycle), and precession (23,000-year cycle) of the Earth's orbit about the Sun (Hays *et al*., 1976; Imbrie *et al*., 1993; Imbrie *et al*., 1992; Imbrie *et al*., 1984; Laskar *et al*., 1993). External forcing at the suborbital scale is thought to be associated with sunspot, solar magnetic polarity, and cosmic ray activity (Beer *et al*., 2000; Bjorck *et al*., 2001; Bond *et al*., 2001; Solanki, 2002; Soon *et al*., 2000; Svensmark, 1998). In terms of radiation amplitude, the 23,000 and 41,000-year Earth-orbital cycles are very large, altering the incoming radiation at a given latitude by the order of 10% relative to present-day levels. Irradiance variability associated directly with the orbital eccentricity cycle and suborbital solar phenomenon are very much smaller, of the order of 0.1% of the present day (Beer *et al*., 2000; Imbrie *et al*., 1993; Solanki, 2002; Soon *et al*., 2000). Paleoclimate records indicate that the ancient monsoon does exhibit variability at timescales where the external radiation forcing is very small and, as well, exhibits orbital-scale variability significantly different in amplitude and phase from that of the primary radiation forcing at the 23,000 and 41,000-year cycles. This indicates that the ancient monsoon, like the modern, is strongly influenced by climatic feedback mechanisms internal to the climate system.

Potential internal forcing mechanisms include any interactions among the atmosphere, ocean, lithosphere, cryosphere, and biosphere which alter the seasonal pressure gradients between land and the ocean, thus altering monsoon circulation. Broadly defined, these internal mechanisms, or boundary conditions, range from the El Niño/Southern Oscillation (ENSO) system to plate tectonics.

16.3 TECTONIC VARIABILITY

The collision of the Indian continent with Asia in the early to middle Eocene, approximately 50 Myr (million years) ago, set the stage for what we experience as modern Asian monsoon circulation. Prior to collision, fossil evidence from the sedimentary basins of eastern China indicates that inland basins were characterized by arid environments whereas nearshore basins were under the influence of arid/humid climate cycles, most likely driven by eustatic changes in sea level. After collision and the initiation of plateau uplift in the late Eocene ($\sim$40 Ma), the inland basins came under the influence of humid or alternating arid/humid climates indicating the initiation of summer monsoon circulation penetrating the Asian interior (Chenggao and Renaut, 1994). While these data provide information regarding the initiation of monsoon circulation in terms of timing, they don't provide information regarding changes in circulation strength. Geological data indicating changes in monsoon strength between the late Eocene ($\sim$40 Ma) and late Miocene ($\sim$10 Ma) are currently lacking. However, general circulation model (GCM) experiments indicate that retreat of the Paratethys seaway and increased elevation and lateral extent of the Tibetan Plateau play important roles in strengthening monsoon circulation (Fluteau *et al*., 1999; Kutzbach *et al*., 1993; Prell and Kutzbach, 1992). Translating these model results into time histories of monsoon circulation awaits reliable estimates for the timing the Paratethys retreat and plateau uplift (Fort, 1996). The long-term history of these events are recorded in land sections as well as the deep-sea fans off India and south-east Asia. These fans, including the Indus Fan and the Bengal Fan, are the largest on Earth, having accumulated and stored the erosional products of the Himalaya and the Tibetan Plateau for the past 50 Myr (Clift *et al.*, 2002; Curray *et al.*, 2003).

Reasonably well-dated geological records related to monsoon evolution have been generated for the last 11 Myr. These records, coupled with an array of evolving GCM experiments have begun to unravel the evolution of the monsoon as it relates to tectonism since the Miocene. A number of increasingly sophisticated GCM experiments examining the impact of plateau uplift on monsoon circulation have been generated since the late 1980s (deMenocal and Rind, 1993; Kutzbach *et al.*, 1989; Kutzbach *et al.*, 1993; Prell and Kutzbach, 1992; Prell and Kutzbach, 1997; Ruddiman and Kutzbach, 1989). These experiments are designed to understand the sign and magnitude of regional changes in surface variables (temperature, precipitation, precipitation minus evaporation, and runoff) as they relate to the elevation and lateral extent of the Himalayan–Tibetan Plateau complex and to changes in orbital forcing. Broadly, uplift enhances both the Asian summer and winter monsoon systems with regions south and east of the plateau becoming more humid and regions north and west of the plateau becoming more arid (Kutzbach *et al.*, 1993).

For south-east Asia in particular, the effects of solar radiation on the hydrological cycle are greatly amplified by increased plateau elevation and lateral extent (Prell and Kutzbach, 1997). Model results indicate that runoff in south-east Asia, a parameter which is readily recorded in marine deposits, is highly sensitive to uplift. For a plateau approximately 1/4 its modern elevation/extent, summer monsoon runoff in south-east Asia increases by 42% between intervals of radiation minima and maxima over orbital timescales. For a plateau of modern elevation/extent runoff increases by 118% between intervals of radiation minima and maxima. This represents a 180% increase in range, attributed to uplift from 1/4 to full modern plateau elevation/extent.

The impact of uplift on the wind regime in the north-west Arabian Sea is also important in terms of monitoring monsoon variability using the geological archive. The exceptionally strong south-west winds off Oman support open ocean and divergent upwelling which, in turn, drives high regional productivity. This upwelling-induced productivity signal is preserved in underlying sediments, serving as a proxy for the strength of summer monsoon circulation. GCM simulations by Prell and Kutzbach (1992) found that the south-west summer monsoon wind regime was not firmly established until plateau elevations reached $\sim$1/2 that of the modern plateau.

The effects of plateau uplift differ for central Asia and the east Asian Loess Plateau. The importance of these regions in defining the history of monsoon circulation derive from the geological record of eolian (wind-derived) deposition which created the Loess Plateau. These sediment sections are composed of an eolian red clay deposit overlain by alternating loess–soil horizons creating a combined sequence of the order of 300 m thick. The loess is transported to the plateau from the north by winter monsoon winds. The interbedded soils are formed in situ by pedogenesis during periods of enhanced summer monsoon circulation (An, 2000). These processes have resulted in an extraordinary and well-dated archive of monsoon circulation. GCM simulations indicate that uplift results in drying within central Asia, the eolian source region, and increased precipitation within the east Asian region of eolian deposition (see An *et al.*, 2001, figure 3). These dynamics resulted in the development of the Loess Plateau with an areal extent of 500,000 km^2.

Given model results indicating strong hydrological sensitivity to tectonic uplift, one can use the well-dated geological record to piece together the monsoon response to uplift since the Miocene. A diverse multiproxy set of geological records indicate that monsoon circulation as we know it today, with strong summer and winter monsoons, initiated in a near threshold-like manner sometime between 9 and 7 Myr ago. The planktonic foraminifer *Globigerina bulloides* as well as the radiolaria *Actinomma spp* and *Collosphaera sp* are diagnostic of cold, nutrient-rich waters not found in the tropics except within upwelling regimes. In the Arabian Sea, off the coast of Oman, these upwelling indicators increased significantly, from near zero values, between 9 and 7 Myr ago indicating the initiation of strong south-west monsoon winds over the Arabian Sea (Kroon *et al.*, 1991; Nigrini and Caulet, 1991; Prell *et al.*, 1992). The record of lithogenic material transported from the Ganges/Brahmaputra River systems to the Bay of Bengal reflect erosion and runoff from the Himalayan–Tibetan orogen. Accounting for the effects of changing sea level on sediment transport (a signal unrelated to monsoon circulation), these records indicate a strong increase in precipitation, erosion, and runoff $\sim$9 Myr ago (Prell and Kutzbach, 1997). Changes in the carbon isotopic composition of soil carbonates indicate paleoecological changes in local plant cover while changes in the oxygen isotopic composition of soil carbonates reflect the isotopic composition of local rainfall. The combined shifts in the oxygen and carbon isotopic composition of soil carbonates in Pakistan at $\sim$7.4 Myr indicate a dramatic ecological shift marking the strengthening of the Asian monsoon system (Quade *et al.*, 1995; Quade *et al.*, 1989). In eastern Asia, the onset of red clay deposition at the base of the Loess Plateau ranges from 8.3 to 7.6 Myr ago depending on location, reflecting the onset of aridity in central Asia and monsoon circulation in east Asia (An *et al.*, 2001; An, 2000; Ding *et al.*, 1998b; Qiang *et al.*, 2001; Sun *et al.*, 1998). Simultaneously, eolian dust deposition peaked downwind of Asia in the North Pacific (Rea *et al.*, 1998).

Taken together, this diverse and independently dated set of proxy climate records indicate increasing seasonality across southern and eastern Asia, as would be predicted by the onset of strong summer and winter monsoon regimes in response to tectonic uplift and extension of the Tibetan Plateau. Evidence for the initiation of normal faulting in Tibet, suggesting that the plateau increased significantly in elevation at approximately 8 ± 3 Myr ago (Molnar *et al.*, 1993), is consistent with the inferred link between the proxy monsoon records and uplift based on GCM results.

16.4 EARTH-ORBITAL VARIABILITY

Orbital-scale paleoclimate studies enjoy a distinct advantage relative to tectonic and suborbital scale studies in that the external (solar) forcing function is accurately known (Laskar, 1999; Laskar *et al.*, 1993). Knowledge of the solar forcing function and the climate response allows one to evaluate the physics of the system using time series analytical approaches. In this regard, there are two important prerequisites. The time series records must have fine enough sample resolution to resolve orbital-scale cycles and chronostratigraphies sufficient to (1) compare variability in cores distributed across broad geographic regions and (2) compare the climate response to variability in external solar forcing. A number of continuous monsoon proxies meeting these requirements are available with lengths spanning the last few hundred thousand to a few million years (Wang *et al.*, 2005).

16.4.1 The Indian summer monsoon

Within the late Pleistocene interval, the two dominant boundary conditions influencing global climate change are variations in solar radiation and global ice volume (Figure 16.1). Figure 16.3 (color section) illustrates the cross-spectra of the Indian summer monsoon relative to insolation and global ice volume over the past 350,000 years (350 kyr). Earth-orbital periods associated with precession of the equinoxes (23-kyr period) and obliquity of the Earth's axis (41-kyr period) dominate the insolation and monsoon spectra whereas the ice volume spectrum is dominated by 100-kyr variance associated with the great ice ages. To a large extent, these 100-kyr ice age cycles are pervasive in paleoclimate records illustrating the global climatic impact of northern hemisphere ice volume. In this regard, the Indian summer monsoon is somewhat of an exception, being dominated largely by variability in the 41 and 23-kyr spectral bands. Much of the orbital-scale monsoon work has focused on understanding monsoon variability within the context of 41 and 23-kyr insolation cycles and changes in global ice volume, the two most obvious external and internal boundary conditions.

Since the mid-1980s a large variety of physical, chemical, isotopic, and biological proxies have been developed to evaluate the variability of the Indian summer monsoon using cores from the Arabian Sea (Almogi-Labin *et al.*, 2000; Altabet *et al.*, 1999; Altabet *et al.*, 1995; Anderson, 1991; Anderson *et al.*, 1992; Anderson and Prell, 1991; Anderson and Prell, 1993; Beaufort, 1996; Budziak *et al.*, 2000; Clemens *et al.*, 1996; Clemens and Prell, 1990; Clemens and Prell, 1991a; Clemens and Prell, 1991b; Clemens and Prell, 2003; Clemens *et al.*, 1991; Murray and Prell, 1991; Murray and Prell, 1992; Overpeck *et al.*, 1996; Prell, 1984a; Prell, 1984b; Prell and Kutzbach, 1987; Prell *et al.*, 1990; Prell and Van Campo, 1986; Reichart *et al.*, 1997; Reichart *et al.*, 1998; Rostek *et al.*, 1997; Shimmield and Mowbray, 1991; Shimmield *et al.*, 1990; Sirocko *et al.*, 1993; Street-Parrott and Harrison, 1984; van Campo *et al.*, 1982; Weedon and Shimmield, 1991). A subset of this proxy array have been shown to possess similar variance at orbital periods, indicating that the summer monsoon signal is not masked by unrelated physical, chemical, or biological processes (Budziak *et al.*, 2000; Clemens and Prell, 2003; Clemens *et al.*, 1991; Reichart *et al.*, 1998). This internally consistent subset of proxies are from cores spanning the past 250 to 400 kyr and is used to assess processed driving monsoon variability at the orbital timescale.

If summer monsoon variability were controlled only by changes in solar forcing over the Tibetan Plateau (sensible heating), monsoon maxima would be in phase with maximum northern hemisphere summer radiation (maximum obliquity and minimum precession; June 21 perihelion) (Figure 16.4). This would be consistent with atmospheric GCM results indicating that the summer monsoon is strongly influenced by changes in sensible heating at the latitude of the Tibetan Plateau (An *et al.*, 2001; Prell and Kutzbach, 1987; Prell and Kutzbach, 1992). Similarly, if summer monsoon variability were controlled only by high-latitude changes in global ice volume, then monsoon maxima should be in phase with minimum ice volume as suggested by atmospheric GCM results indicating weakened monsoons at times of increased global glaciation (Prell and Kutzbach, 1992; Prell and Kutzbach, 1997). Assuming equal sensitivity to both, the phase of the summer monsoon response would fall between northern hemisphere summer insolation maxima and global ice volume minima at both the obliquity and precession periods. The measured monsoon phase response does not entirely fit this simple conceptual model (Figure 16.4). Within each 41-kyr obliquity cycle, the Indian summer monsoon is strongest at the same time northern hemisphere summer insolation is strongest, ~8 kyr before global ice volume reaches a minimum (Clemens and Prell, 1990; Clemens and Prell, 2003; Clemens *et al.*, 1991). Within each 23-kyr precession cycle, the Indian summer monsoon is strongest ~8 kyr after northern hemisphere summer insolation is strongest, ~3 kyr after global ice volume reaches a minimum (Clemens and Prell, 1990; Clemens and Prell, 2003; Clemens *et al.*, 1991; Reichart *et al.*, 1998).

Two explanations have been put forth for these phase relationships. Reichart *et al.* (1998) suggest that the Indian summer monsoon is sensitive to late summer insolation forcing (August or September perihelion), resulting in longer although not necessarily stronger summer monsoons. Clemens *et al.* (2003, 1991), suggest that the observed phase relationships can be explained by the combined influence of sensible heating over the Tibetan Plateau (modulated by changes in ice volume) and the timing of latent heat export from the southern subtropical Indian Ocean. They hypothesize that orbital configurations characterized by warm southern hemisphere summers followed by cold winters (i.e., increased seasonality) precondition the ocean to release large amounts of latent heat at the onset of the Indian summer monsoon. This latent heat is transported to Asia in the low-level, cross-equatorial flow and released over Asia during precipitation, strengthening the monsoon. Only recently have fully coupled ocean–atmosphere GCM's become available that are capable of testing these hypotheses (Liu *et al.*, 2004; Liu *et al.*, 2003). Such experiments are in the planning phase.

16.4.2 The east Asian summer and winter monsoons

Proxy records of both summer and winter monsoon variability have been developed from the loess–paleosol sections of the east Asian Loess Plateau. Grain size and flux

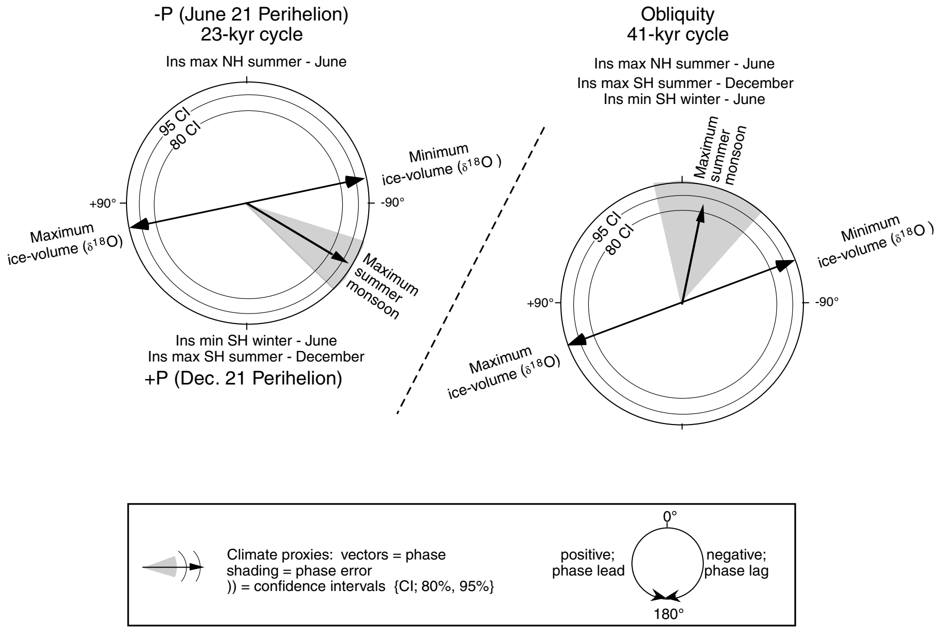

Figure 16.4. Indian summer monsoon phase wheels for precession and obliquity. Phase wheels display coherence and phase relationships among climate proxies and internal/external boundary conditions. The zero phase is set at June 21 perihelion for precession and at maximum obliquity; both correspond to maximum northern hemisphere summer insolation. The phase difference between an orbital component and a climate proxy is plotted as a vector whose angle is measured clockwise (positive, lag) or counterclockwise (negative, lead) from the zero point. The error associated with the phase is illustrated by the shaded area centered on the vector. The length of the vector indicates the degree of coherence between the climate proxy and the orbital parameter (below the 80% CI, between the 80% and 95% CI, or >95% CI). Text at 0° and 180° indicate seasonal insolation relationships associated with each orbital configuration for the southern hemisphere and northern hemisphere. 180° on the precession phase wheel and 0° on the obliquity phase wheel are characterized by warm southern hemisphere summers (December) followed by cold southern hemisphere winters (June), hypothetically ideal configurations for exporting latent heat from the southern Indian Ocean to Asia during the northern hemisphere summer monsoon. At both the precession and obliquity bands, summer monsoon maxima occur between the theoretical times of maximum latent heat export and global ice volume minima indicating that the monsoon is more sensitive to these internal boundary conditions than to direct sensible heating of the Asian Plateau.

measurements have been used extensively as winter monsoon proxies (An *et al.*, 1991a; Ding *et al.*, 1994; Liu *et al.*, 1999; Xiao *et al.*, 1995) while magnetic susceptibility, ^{10}Be, and chemical weathering indices have been used as summer monsoon proxies (An *et al.*, 1991b; Shen *et al.*, 1992). The coarse age models for these records is derived from the geomagnetic polarity reversal timescale, which is based on radiometric dating methods. The finely tuned age models are derived by correlation to the marine $\delta^{18}O$ record of global ice volume via astronomical tuning, assuming that loess deposition is in phase with the growth and decay of the northern hemisphere ice sheets (Ding *et al.*, 1995; Ding *et al.*, 2002; Shackleton *et al.*, 1995a). As such, there is no means of assessing changes in the timing (phase) of monsoon dynamics measured from the loess records relative to evolving glacial boundary conditions or insolation forcing; the monsoon response is phase locked to ice volume by virtue of the age model. Ding *et al.* (2002) address this issue, noting that the loess grain size and marine $\delta^{18}O$ record are more similar to one another over the past 1.8 Myr than prior to this time (see also Liu *et al.*, 1999). They suggest that the assumption of a phase lock between loess and $\delta^{18}O$ may not be appropriate prior to 1.8 Myr ago. This issue notwithstanding, the loess record provides an excellent means of assessing paleomonsoon variability. As summarized in An *et al.* (2001), proxies derived from the red clay and loess–soil sequences indicate that both the east Asian summer and winter monsoons strengthened between 3.6 and 2.6 Myr ago. After 2.6 Myr ago, the east Asian summer monsoon began to weaken while the winter monsoon strengthened. Within the interval of the large-amplitude 100-kyr ice age cycles, the summer and winter monsoon proxies are out of phase indicating an alternation between winter and summer monsoon dominance with stronger winter monsoons and weaker summer monsoons during glacial intervals and stronger summer monsoons and weaker winter monsoons during interglacial intervals (An *et al.*, 2001; An, 2000).

A large variety of east Asian monsoon proxies from marine cores indicate similar dynamics, strengthened winter monsoons and weakened summer monsoons during glacial intervals of the late Pleistocene. However, these inferences are largely based on visual inspection of monsoon proxies relative to the marine $\delta^{18}O$ record; very few present phase analyses of summer and winter monsoon variability at specific orbital frequency bands. When viewed in the context of spectral (amplitude and phase) analyses, the relationship between monsoon strength and ice volume is more complicated than that inferred from visual inspection alone. Beaufort *et al.* (2003) analyze a coccolithophore (*Florisphaera profunda*) proxy for winter monsoon productivity in the Sulu Sea. They report that strong winter monsoons are in phase with maximum ice volume at the 100-kyr band and lag maximum ice volume by 40° ($\sim$5 ky) in the obliquity band (Figure 16.5). The in-phase relationship at the 100-kyr period is consistent with the east Asian loess and marine records described previously. However, the significant lag at the 41-kyr obliquity period associated with the obliquity of the Earth's axis suggests that winter monsoon strength at this period responds to both high-latitude ice volume as well as decreased insolation during northern hemisphere winter. The *profunda* record has very little 23-kyr amplitude, the variance in this region of the spectrum is shifted toward the 29-kyr period. This 29-kyr period is interpreted as a heterodyne (beat) between ENSO-induced winter monsoon variance at the eccentricity frequency and insolation-induced variance at the precession frequency ($1/100 \pm 1/23 = 1/19$ and $1/30$; where frequency = 1/period) (Beaufort *et al.*, 2003; Beaufort *et al.*, 2001). Similarly, the charcoal record from the same core, used as a proxy for decreased summer monsoon precipitation (drought-induced fires), indicates 19

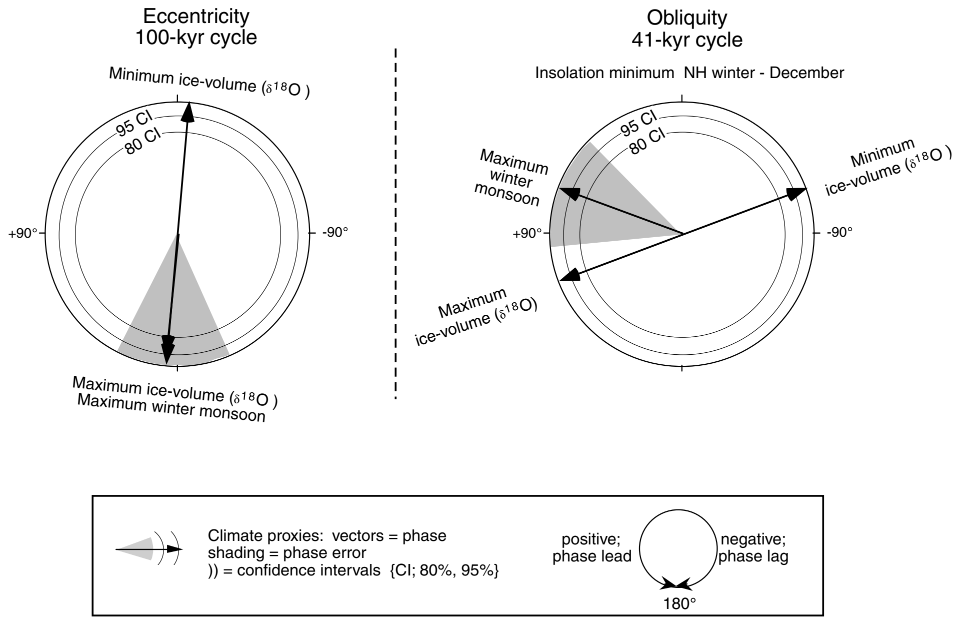

Figure 16.5. Asian winter monsoon phase wheels for eccentricity and obliquity. See Figure 16.4 caption for explanation of phase wheel construct. For the obliquity band, this record from the Sulu Sea indicates that the strongest winter monsoons are driven by a combination of decreased insolation heating over Asia during winter (0° on the obliquity phase wheel) and northern hemisphere ice volume maxima (111° on the obliquity phase wheel). The eccentricity band has no direct insolation forcing although eccentricity of the Earth's orbit modulates the amplitude of precession. At this band, the strongest winter monsoons coincide with times of greatest ice volume.

and 29-kyr periods (Beaufort *et al.*, 2003). These are also taken to represent an ENSO influence on the strength of the summer monsoon. This interpretation of the 19-kyr and 29-kyr periods stems from the work of Federov and Philander (2000) indicating strengthened and/or more frequent El Niño's during interglacial intervals. Monsoon pollen records from the southern hemisphere (north of Australia) also indicate significant amounts of variance at the 30-kyr period suggesting that this signal is pervasive throughout the Indonesian region (Kershaw *et al.*, 2003).

Morely and Heusser (1997) report phase estimates for a pollen-based proxy record of summer monsoon variability off Japan. They report a summer monsoon phase of $-136°$ for the precession band, within error, the same as the Indian summer monsoon estimates from the Arabian Sea.

This sparse number of records analyzed in terms of amplitude and phase indicate that the summer and winter monsoons of the late Pleistocene (~0–400 ka) respond to processes beyond summer season radiation forcing over the Asian continent and glacial–interglacial changes at the 100-kyr cycle. The Indian summer monsoon records indicate that latent heat export from the southern hemisphere Indian Ocean may also be an important feedback strengthening the monsoon low over Asia. The in-phase response with the summer monsoon index off Japan suggests that this feedback may be strong enough to drive the timing of the east Asian summer monsoon as well. These few records also suggest the beginning of a regional pattern in which monsoon records from outside the western Pacific Warm Pool (The Arabian Sea and off Japan) have variance concentrated at the primary (41 and 23-kyr) orbital periods, whereas those from within the Warm Pool exhibit a combination of orbital and heterodyne periods linked to an additional ENSO influence.

16.5 CENTENNIAL AND MILLENNIAL VARIABILITY

Superimposed on the longer term orbital-scale variations are abrupt centennial and millennial-scale events during which climate changes from one state to another within decades or less, remaining so for hundreds to one or two thousand years (Figure 16.2). Records with sufficient temporal resolution to resolve these scales of variability are typically from within the latest Pleistocene interval, usually from within the last 60 kyr.

Centennial-scale variations have been reported in a limited number of highly resolved Indian and east Asian paleomonsoon records (Agnihotri *et al.*, 2002; Buehring, 2001; Neff *et al.*, 2001; Thompson *et al.*, 1997; von Rad *et al.*, 1999; Wang *et al.*, 1999b; Wang *et al.*, 1999c). These records indicate a periodicity in the range of 84–102 years as well as at $\sim$210 years and are interpreted as a monsoon response to the $\sim$88-year solar (Gleissberg) cycle and the $\sim$205-year solar (de Vries or Suess) cycle. The issue is that measured short-term solar irradiance variations are of the order of 0.1% ($1.5\,\mathrm{Wm^{-2}}$), seemingly too small to drive significant climate change. However, there is some suggestion of climate sensitivity to parts of the solar spectrum that have larger amplitude variability. For example, Beer (2000, 2002), and references therein, suggest the possibility that variability in the UV portion of the spectrum may influence climate via changes in ozone production and the radiation balance at high latitudes. Specific mechanisms by which the monsoon might respond to this forcing at centennial (and millennial) scales are unknown. Lack of theory in this regard is not unique to the monsoon system; it is common to the climate field in general. Thus, while paleoclimate data indicate significant variance at the centennial and shorter timescales typically associated with solar variability, no unifying theory exists linking the two (Benestad, 2002).

Millennial-scale variability, sometimes referred to as Dansgaard–Oeschger (DO) variability was first recognized in Greenland ice cores (Dansgaard *et al.*, 1993; Johnsen *et al.*, 1992) but is now known to occur globally (Leuschner and Sirocko, 2000; Voelker, 2002). DO variability is characterized by a 1,470-year cyclicity (Grootes and Stuiver, 1997; Hinnov *et al.*, 2002; Rahmstorf, 2003; Schulz, 2002; Yiou *et al.*, 1997) although it appears sometimes to be manifest in terms of harmonics at periods of ~700 to ~775 years (Sarkar *et al.*, 2000; Wang *et al.*, 1999a) as well as ~3,000 and ~4,500 years (Alley *et al.*, 2001). Based on currently available records, DO variability is best expressed during marine isotope stage 3 (MIS 3) spanning the interval ~30 to 60 kyr before present but has been documented in older intervals as well (to 400 kyr BP) although at considerably lower amplitude (de Garidel-Thoron *et al.*, 2001; Oppo *et al.*, 2003).

The direct (solar) forcing for orbital-scale variability (10^4 years) is well known and the chronology is generally sufficient to evaluate leads and lags among system components as previously described. For DO variability, the forcing is uncertain and the chronology at this point in time is insufficient to assess leads and lags among system components measured at globally distributed sites, information that would be very useful in deriving the underlying physics of millennial-scale climate variability. For example, the Greenland ice core records are very well dated (by layer counting) from the present to 40,000 years ago whereas the marine records rely on less accurate radiocarbon-based chronologies within this interval. This discrepancy in age model reliability precludes direct analysis of phase among records from these archives, information that might implicate high or low-latitude processes as initiating these abrupt climate change events.

Rahmstorf (2003) recently analyzed well-dated Greenland ice core records, finding that the 1,470-year cycle is stable to within a few percent, suggesting that the initial forcing may be solar in origin. In this case, the large-amplitude DO events of MIS3 might be thought of resulting from a stochastic resonant response to low-amplitude solar forcing. The geographic origin of this amplified signal and its mode of global transmission are unknown although two schools of thought exist: (1) that the signal initiates in the high-latitude North Atlantic in response to changes in freshwater influx and is transmitted globally by atmospheric and deep-ocean thermohaline circulation (e.g., Ganopolski and Rahmstorf, 2001); or (2) that the signal initiates in the tropical Pacific ocean–atmosphere system via ENSO-like physics and is transmitted globally via changes in atmospheric circulation (e.g., Cane and Clement, 1999).

In the first case, millennial-scale monsoon variability is likely a downstream response to changes in the North Atlantic transmitted across Eurasia by changes in snow and ice cover (Ganopolski and Rahmstorf, 2001; Overpeck *et al.*, 1996) leading to weakened winter and strengthened summer monsoons during Greenland interstadial (warm) periods and visa versa, strengthened winter and weakened summer monsoons during Greenland stadial (cold) periods. This general phasing, stronger summer monsoons during Greenland interstadials, is supported by the radiometrically-dated Hulu Cave $\delta^{18}O$ record from eastern China (Wang *et al.*, 2001b). In this case, given sufficient chronologies, monsoon proxies would lag North Atlantic proxies by a small amount.

An ENSO-like mechanism initiating in the tropical Pacific may result in monsoon proxies leading North Atlantic proxies at the millennial timescale. In general, the modern summer monsoon weakens in association with El Niño conditions although the strength of this relationship varies on decadal timescales (Clark *et al.*, 2000a; Kumar *et al.*, 1999; Wang *et al.*, 2001a) and the Indian and east Asian subsystems respond somewhat differently to ENSO variability (Wang *et al.*, 2003a; Wang *et al.*, 2001a). Extrapolating the El Niño/weak summer monsoon relationship to paleoclimate timescales, Stott *et al.* (2002) interpret salinity records from the western pacific warm pool as indicating stronger and/or more frequent El Niños (weaker summer monsoons) during Greenland stadials. Again, this overall phasing is consistent with the Hulu Cave record. However, if the initial trigger for millennial-scale climate change resides in the tropical ENSO system, then monsoon proxies should lead North Atlantic proxies.

Current chronologies are not reliable enough to differentiate the small lead and lag relationships associated with these two hypotheses. The best that can be said at this point is that within the limitations of current chronologies, strong summer monsoons are generally coincident with Greenland interstadials as indicated by proxy records from the Arabian Sea (Altabet *et al.*, 2002; Schulz *et al.*, 1998), South China Sea (Wang *et al.*, 1999a), Sulu Sea (Dannenmann *et al.*, 2003), and Philippine Sea (Stott *et al.*, 2002) as well as terrestrial cave and loess profiles from Asia (Ding *et al.*, 1999; Ding *et al.*, 1998a; Wang *et al.*, 2001b) (Figure 16.2). Improved ability to accurately date geological archives is important to unraveling the origin of millennial-scale climate variability, including the role of monsoon circulation. Advances are being made in this area with the goal of eventually placing new and existing paleoclimate records on common age scales sufficient to evaluate phase relationships among globally distributed proxy records. Sarnthein *et al.* (2002) provide a brief review of these issues and advances.

16.6 SUMMARY

Geological archives contain records of paleomonsoon variability with temporal resolution ranging from seasonal to tectonic. These records offer the ability to evaluate monsoon climate change under boundary conditions different from the present and within intervals of time well beyond the reach of the instrumental record.

At the centennial to millennial scale (10^2 to 10^3 years), proxy records of east Asian and Indian monsoon circulation indicate that the summer monsoon strengthens during times of abrupt warming in the North Atlantic region as recorded in the GISP2 ice core. However, age control is not yet sufficient to assess the detailed timing of changes in monsoon circulation relative to high-latitude warming. As such it is not possible to determine if monsoon circulation changes prior to, coincident with, or after changes in the North Atlantic climate. At this point, enhanced understanding of the dynamics involved come from documenting the geographic distribution of millennial-scale variability in terms of the system components involved (atmosphere, cryosphere, biosphere, surface, intermediate, and deep-ocean circulation).

At orbital timescales (10^4 to 10^5 years), the late Pleistocene Indian summer monsoon is dominated by variance in the obliquity (41-kyr) and precession (23-kyr) bands. For both bands, the strongest summer monsoons take place between times of minimum northern hemisphere ice volume and maximum latent heat export from the southern hemisphere Indian Ocean. The late Pleistocene east Asian winter monsoon is dominated by variance in the eccentricity (100-kyr) and obliquity (41-kyr) bands. Within the eccentricity band, the winter monsoon is strongest during times of maximum ice volume. Within the obliquity band, the winter monsoon is strongest between times of ice maxima and insolation minima during northern hemisphere winter. In spite of strong solar forcing at orbital time-scales, proxy records of the east Asian and Indian monsoon systems indicate changes in monsoon circulation are driven at least as much by changing internal boundary conditions as by direct solar forcing. Limited geographic coverage suggests that monsoon circulation in regions outside the Pacific Warm Pool appear to be dominated by variance more closely associated with orbital forcing whereas those from within the Warm Pool appear to be influenced by variance associated with ENOS-like dynamics as well.

At tectonic timescales (10^6 to 10^7 years), monsoon circulation varies as a function of changes in location/extent of seaways and the location/elevation of continental landmasses. Monsoon circulation initiated with the collision of India and Asia approximately 40 Myr ago. Modeling results suggest that strengthened summer season circulation (similar to that of the modern) initiated between 9 and 7 Myr ago, likely in response to the Tibetan Plateau having achieved approximately half its modern elevation and lateral extent. Further strengthening of the winter monsoon and weakening of the summer monsoon took place as glacial conditions increased over the past three million years.