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Aspects of atmospheric circulation: the Late Pleistocene (0–950,000 yr) record of eolian deposition in the Pacific Ocean

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ABSTRACT

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The eolian component of pelagic sediments provides a proxy record of atmospheric circulation intensity and of dust transport. Examination of those records indicates that the atmosphere responds to orbital forcing but that during the past 875,000 years the variability in atmospheric circulation is at much shorter periods than the 100,000-year variation in ice volume. The mid-Brunhes climatic event is well characterized in the eolian grain-size records and in records of sea-surface phenomena; it is not seen in proxy indicators of either deep-water or ice-volume variability. The increase in the amplitude of paleoclimatic variability that denotes the early/late Pleistocene transition occurs suddenly about 875,000 years ago in the eolian grain-size record, which is about 20,000 to 25,000 years earlier than the transition in the deep sea $CaCO_3$-dissolution record or in the $\delta^{18}O$ proxy record of ice volume.

Introduction

To understand changes in the Earth's environment it is useful to examine the several climatic subsystems separately because each has its own history of variability. Climatic subsystems are those naturally occurring systems that are capable of storing, transporting, and releasing both heat and moisture. These systems, the atmosphere, the surface ocean, the deep ocean, the cryosphere, the land surface and the biosphere, all have different response times to climatic stimuli and all interact either directly or indirectly. These interactions seem to occur at three distinct time scales that range from the rather short *oceanic* timescale of several hundred to a few thousand years, to the *orbital* timescale of tens of thousands to several hundred thousand years, to the *tectonic* timescale of one-half million or more years.

An important goal of the study of paleoclimatology is to understand the variability of each climatic subsystem for all timescales and to determine how the various subsystems relate to each other. Deep-sea sediments are the single best recorder of this range of climatic variability; the challenge is to sort out all the different paleoclimatic signals. The relative abundance of planktic and benthic organisms provides the history of those ecosystems, and fluxes of biogenic materials provide some record of past productivity. Isotopic and geochemical data provide information about ice volume, ocean temperature, sea-floor hydrothermal activity, and ocean circulation patterns. The terrigenous mineral component of ocean sediments contains a record of the nature of the continental source area and the pertinent transport process.

Far from shore, the terrigenous component of deep-sea sediments consists of eolian dust. This component of pelagic sediments is the best proxy indicator of the nature and variability of atmospheric circulation, and is a useful proxy of continental climate (Rea et al., 1985). This paper examines the record of variability of eolian transport and deposition processes as documented

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on orbital time scales over the past 950,000 years and, when possible, relates that atmospheric record to that of other climatic subsystems. The longer Cenozoic eolian-dust record of atmospheric processes on tectonic timescales is summarized by Rea (1989). Rea and Leinen (1988) have described the eolian-dust record of the prevailing westerlies during the transition from the last glacial to the Holocene from a series of cores from the Northwest Pacific.

Eolian transport and deposition processes

Most of the dust incorporated into pelagic sediments originates as soil in an arid or semiarid climate. That soil material is entrained and elevated by the strong winds of large spring storms and then is transported long distances in the middle and upper troposphere by the zonal wind systems (Prospero, 1981a, 1981b). Most of the total flux of dust to the ocean is accomplished by a few storms each year. The mineral component of pelagic sediments, away from the influence of hemipelagic, turbidite, and ice-rafting depositional processes, is considered to be eolian because (1) dust mineralogy matches that of terrigenous materials in the surface sediments (Ferguson et al., 1970; Blank et al., 1985) and (2) quartz, an unequivocally eolian mineral when found in the middle of the ocean, shows distribution patterns in surface sediments that correspond to the zonal wind patterns (Leinen and Heath, 1981; Leinen et al., 1986).

The longest temporal record of modern dust transport is of material that crossed the North Atlantic Ocean from North Africa, compiled by Prospero and his colleagues. These data relate the dust flux to climatic conditions in the Saharan–Sahelian source region. Monthly values show increased dust transport during the late spring and early summer and indicate an order of magnitude variation in any specific year (Prospero, 1981a; Prospero et al., 1981). The annual transport cycles correlate with the climatology of arid North Africa, but not with the general intensity of the Northern Hemisphere tradewinds that are stronger during the winter months. On a longer term basis, the annual flux of dust across the North Atlantic had a threefold to fivefold increase during the height of the Sahelian droughts in 1973–1974 and 1983–1984, known to be times of significantly decreased rainfall (Middleton, 1985; Nicholson, 1985), in comparison to the more normal years (Prospero and Nees, 1977, 1986). This data set and a similar data set which was collected in the North Pacific (Uematsu et al., 1983; Parrington et al., 1983) provide evidence for the working assumption that on orbital and longer time scales the flux of dust to the ocean depends not on wind strength but primarily on the availability of material which, in turn, depends on the degree of aridity of the eolian source region (Middleton, 1985).

Dust raised from the land surface is a complex mixture of sizes and shapes. However, as injection height and distance from source increases, the grain size of the eolian load becomes smaller and the size distribution more normal (Nickling, 1983; Hobbs et al., 1985). Air and land-based sampling (Gillette et al., 1974; Johnson, 1976; Glaccum and Prospero, 1980) and theoretical calculations (Windom, 1969; Jaenicke, 1979; Schutz et al., 1981) indicate that beyond 1000–2000 km from the source area the size distribution of the grains changes minimally. Along a 3600 km transect that extends west to east in the North Pacific, a transect that lies about 4000–8000 km downwind from the source area in China, the median grain size of the dust in surface sediments changes from 8.5φ (2.8 μm) to 8.7φ (2.4 μm) (Janecek, 1985). The second working hypothesis, then, is that these small grains, which are thousands of kilometers from the source area, are essentially in equilibrium with the transporting zonal winds; and, therefore, their size is a measure of the intensity of those winds. Parkin and his colleagues (Parkin and Shackleton, 1973; Parkin, 1974; Parkin and Padgham, 1975) first used the size of wind-blown grains to make estimates of past wind intensity. Janecek and Rea (1985) have made preliminary attempts to quantify the grain size to wind intensity relation based on the work of Gillette (Gillette et al., 1974; Gillette, 1981), but such quantification is not yet well resolved [see also Tsoar and Pye (1987)]. Demonstration that the Milankovitch periodicities are preserved in the grain-size record, both in the westerlies (Janecek and Rea, 1985) and in the tradewinds (Pisias and Rea, 1988), and that the grain-size record of wind intensity can be linked quantitatively to the radiolarian record of equatorial divergence (Pisias and Rea, 1988), shows that this size parameter of the eolian signal does preserve significant paleoclimatic information.

In summary, almost a decade of studying the eolian component of pelagic sediments has shown that dust flux to the sea floor records the supply function and that supply is related to the climatology of the source region. Dust grain size is determined by the transport process and records the intensity and the variability of that process. The flux and size values vary independently (Janecek and Rea, 1985; Chuey et al., 1987), and can be represented as the atmospheric analogy of the capacity and competence of rivers. Useful reviews of eolian processes have been done by Windom (1975), Prospero (1981b), Rea et al. (1985), and Pye (1987).

Methodology and sedimentology

The cores selected for study are obtained from locations that are remote from continents to ensure that the mineral component, usually a minor percentage of the bulk sediment, is not affected by river-borne hemipelagic material. The sediments, usually calcareous oozes from the equatorial Pacific cores and siliceous oozes or pelagic clays from the North Pacific cores, are subjected to a sequential extraction process that isolates the mineral component. These extractions remove calcium carbonate, oxides, hydroxides and zeolites, and opaline silica; details of the procedure are described in Rea and Janecek (1981). Volcanic ash survives this extraction procedure. Therefore in places like the Northwest Pacific that are directly downwind from island arcs disseminated ash may occur in the sediment and accordingly smear slides of the extracted material are examined for volcanic glass.

The weight-percent of mineral component that results from the extraction process is only an intermediate value. Percentages of minor sedimentary components fluctuate in response to and antithetically from the predominant component, usually carbonate, thus obscuring any paleoenvironmental signal. To discover the actual input values, it is necessary to calculate the true mass flux of the eolian material to the sediment, a value termed the mass accumulation rate or MAR. The MAR (which is measured in $g(cm^2 \times 10^3\ yr)^{-1}$) is the product of the linear sedimentation rate, (LSR, measured in $cm(10^3\ yr)^{-1}$), and the bulk density of the dry sediment (DBD, measured in $g(cm^3)^{-1}$). The DBD is determined by freeze-drying fresh sediments and determining the porosity, then calculated as DBD = (1-porosity) × grain density. Working with dried sediments is more difficult. Various approaches are used such as determining initial porosity from the quantity of dried sea salt present or determining the porosity to lithology variation in a nearby fresh core and applying that relation to the dried sediment (see Chuey et al., 1987, for details). The ability to determine porosity values limits the accuracy of resulting information when working with dried cores. When working with fresh cores, porosity values are much better and the determination of the LSR is the limiting factor for the final accuracy of the sediment-flux information.

The size of the eolian grains within the range of 1–16 μm (10–6φ) was determined using a Coulter Counter particle size analyzer. Data are reported as median grain size (φ_{50} of Folk, 1974) and have a precision of $\pm 0.03\varphi$. Phi units are a logarithmic measure of grain size such that $\varphi = -\log_2 D_{mm}$, where D_{mm} is the grain diameter in millimeters, thus $10\varphi = 1$ μm, $9\varphi = 2$ μm, $8\varphi = 4$ μm, $7\varphi = 8$ μm etc.

Much of the information discussed below is from piston core RC11-210 which was raised from 4420 m depth in the equatorial Pacific at 1°49′N, 140°03′W (Fig.1). That core provides a 946,000-year long record of a complete suite of paleoclimatic variables: $\delta^{18}O$, $\delta^{13}C$, the MAR of $CaCO_3$, opal, organic carbon, and dust, the changing assemblages of radiolarians, and the grain size of the eolian dust. Analysis of 175 samples from core RC11-210 was done and resulted in an average temporal sample spacing of about 5000 years. Sedimentation rates were determined by comparing the oxygen-isotope record of RC11-210 to the SPECMAP timescale of Imbrie et al. (1984) by using the inverse-mapping technique developed by

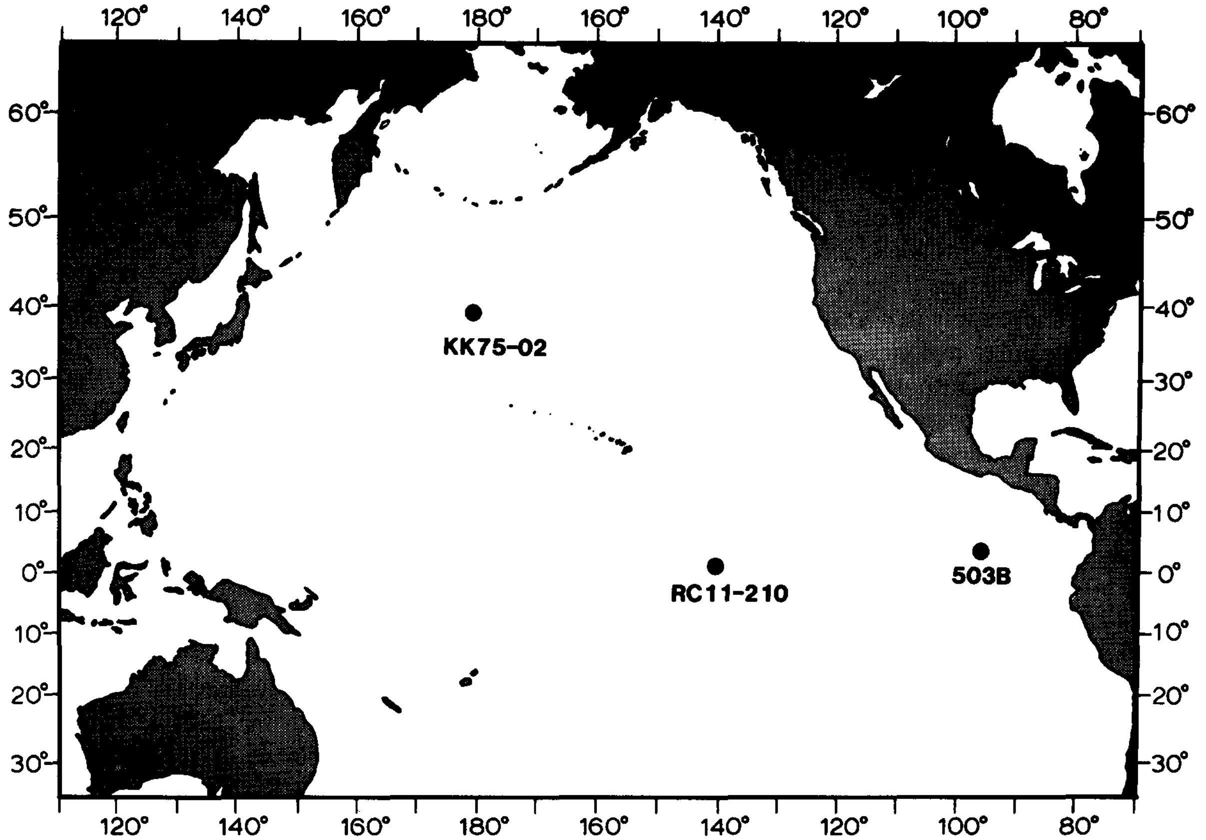

Fig.1. Index map of the Pacific Ocean showing locations of the cores discussed in the text.

Martinson et al. (1982). These records and the various methods used in their construction are described by Chuey et al. (1987), and Pisias and Rea (1988).

We also have examined the eolian record, although at a coarser time interval, contained in the uppermost 5.6 m of siliceous nannofossil ooze that was recovered from hydraulic-piston cores taken at Deep Sea Drilling Project (DSDP) Hole 503B located in the eastern equatorial Pacific at 4°03'N, 95°38'W (Fig.1), in 3672 m of water. This record spans about 420,000 years; sediment ages were determined by visual comparison of the measured $\delta^{18}O$ values with the SPECMAP standard of Imbrie et al. (1984) (Rea, 1982; Rea et al., 1986). The samples from 503B are spaced an average of 10,000 years apart.

Piston core KK75-02, from the central North Pacific, provides information about the last 750,000 years of eolian deposition. That core was raised from 5475 m of water east of Hess Rise at 38°37'N, 179°20'E (Fig.1), and consists of siliceous clay and clayey siliceous ooze. Age control for this core consists of four radiolarian extinction levels and the location of the Brunhes/Matuyama reversal boundary (Janecek and Rea, 1985). The 94 samples from this core have an average temporal spacing of about 8000 years.

RC11-210 is from south of the Intertropical Convergence Zone (ITCZ) in the regime of the Southern Hemisphere tradewinds and Hole 503B is near the present southern margin of the ITCZ. The convergence zone is the site where the atmospheric circulation systems of the hemispheres meet and it is displaced north of the geographic equator by about 5° of latitude. This displacement occurs because the Southern Hemisphere wind systems are more vigorous, basically because the pole-to-equator temperature gradient in the Southern Hemisphere is steeper than that in the Northern

Hemisphere. The location of the ITCZ denotes the dynamic position of the balance in hemispherical wind energy and shifts north and south with the seasons. The ITCZ, characterized by clouds and enhanced rainfall, is an effective interhemispherical barrier to the eolian transport of dust (Raemdonck et al., 1986; Arimoto et al., 1987).

Core KK75-02 is from beneath the prevailing westerlies and apparently provides a record of that zonal wind system throughout the late Pleistocene.

Late Pleistocene eolian record

Equatorial Pacific

The eolian record from core RC11-210 spans the entire late Pleistocene and continues across the transition at about 850,000 or 900,000 years ago into the uppermost portion of the early Pleistocene (Fig.2). During the past 946,000 years dust has accumulated at rates of 4–90 mg$(cm^2 \times 10^3\ yr)^{-1}$ (large values at the top of the core may be spurious, and the sharp peak at 100,000 years ago is an ash layer, see discussion in Chuey et al., 1987). Flux maxima occur in conjunction with both interglacial (Stages 7, 15, and perhaps 19) and glacial (Stages 10 and 12) times. The change in correspondence between the interglacial maxima at Stage 7 and the glacial-aged maxima at Stages 10 and 12 occurs about 300,000 years ago (Fig.2). Dust fluxes revert to the original pattern of interglacial maxima sometime between 480,000 and 560,000 years ago.

Farther to the east at Hole 503B, the MAR of the eolian component ranges from 23 to 183 mg$(cm^2 \times 10^3\ yr)^{-1}$. Flux values are highest during interglacial Stages 5 and 7 and during glacial Stage 10 (Fig.3).

The eolian grain-size record in core RC11-210 provides a remarkable history of the intensity of the southern hemisphere tradewinds (Fig.2). Size of the eolian grains ranges from 9.30φ (1.58 μm) to 7.18φ (6.90 μm). The outstanding characteristic of this time series is that the size variations of the eolian grains occur at a much higher frequency than does a standardized benthic oxygen isotope record of ice volume changes (Imbrie et al., 1984). A change in the character of the record occurs at

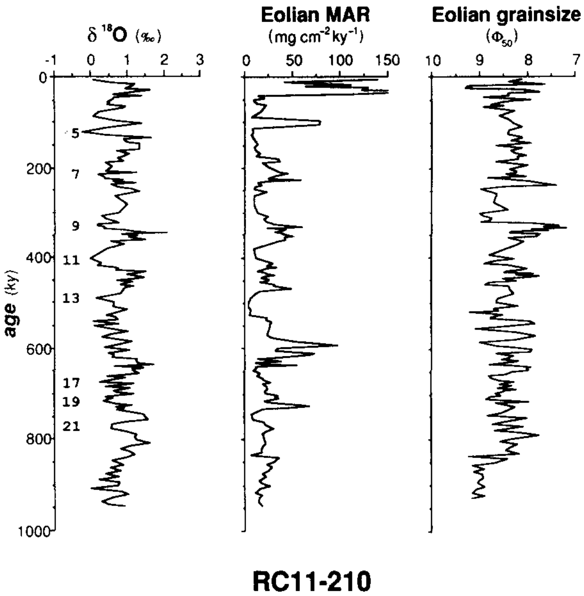

Fig.2. Oxygen isotopic values of planktic foraminifer *Globorotalia tumida* (odd numbers denote interglacial stages), and mass accumulation rate and grain size of eolian dust in core RC11-210.

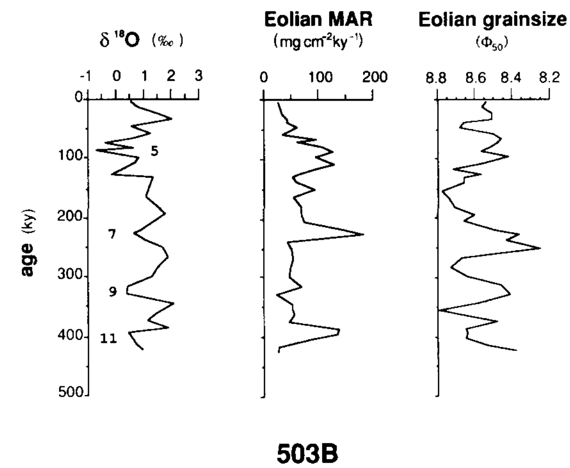

Fig.3. Oxygen isotopic values of planktic foraminifer *Globorotalia tumida* (odd numbers denote interglacial stages), and mass accumulation rate and grain size of eolian dust in DSDP Hole 503B.

about 300,000 years ago and is denoted by a 50,000-year-long period of relatively small grain size with low variability. This interval separates a younger portion of the record characterized by higher frequency, lower amplitude fluctuations from the older portion of the record which was higher amplitude and lower frequency size fluctuations. At about 875,000 years ago the record underwent another pronounced change, such that older dust grains average about 0.5φ smaller and had much less variability than the younger materials.

North Pacific

None of the North Pacific records produced to date are of the temporal resolution of the RC11-210 record. Most cores from that region have been raised from depths greater than the calcium carbonate compensation depth and are without the foraminifers necessary for determination of an oxygen isotope record. The ultimate goal of the North Pacific study is to correlate the detailed eolian-flux data, presumably the distal record of the great Asian loess deposits, and the oxygen isotope record of climate change (Hovan et al., 1988).

A detailed eolian grain-size record (Fig.4) has been produced for core KK75-02 by Janecek

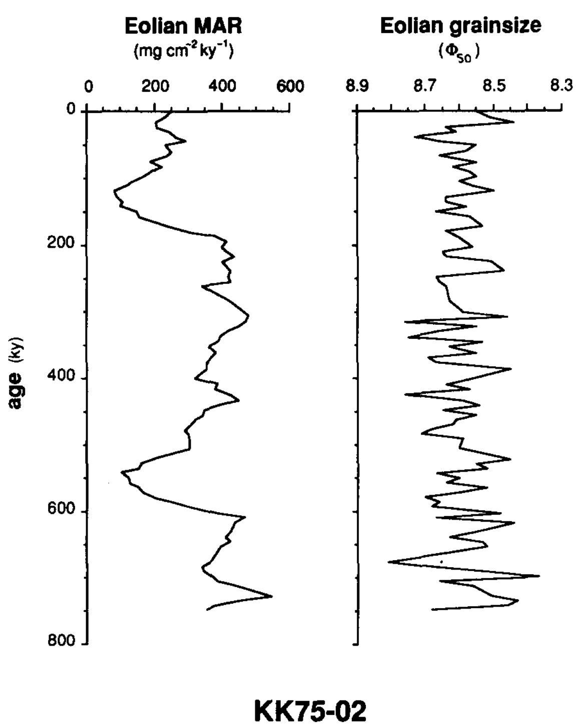

Fig.4. Mass accumulation rate and grain size of eolian dust in core KK75-02.

(1983). Grains vary in size from 8.81φ (2.23 μm) to 8.37φ (3.02 μm), a much smaller range of variability than that in the equatorial cores. An interval of low variability in eolian grain size occurs about 250,000 to 300,000 years ago and separates the younger portion of the record characterized by lower amplitude size fluctuations from an older portion where the grain size variability is much higher in amplitude and perhaps longer in wavelength (Janecek and Rea, 1985). The eolian-flux data from KK75-02 are less detailed because of the limited LSR and DBD information. Flux values range from about 100–500 mg(cm$^2 \times 10^3$ yr)$^{-1}$ and average 300–400 mg(cm$^2 \times 10^3$ yr)$^{-1}$ (Janecek, 1983; Fig.4). Significant flux minima occur at about 100,000 to 140,000 years ago and at 530,000 to 570,000 years ago.

Late Pleistocene paleoclimatology

Orbital control of atmospheric circulation

Spectral analyses have been conducted on the eolian grain-size records from both cores KK75-02 and RC11-210. The spectrum calculated for KK75-02 displays three discrete peaks which correspond to periods of 104,000, 41,000 and 23,000 years (Janecek, 1983; Janecek and Rea, 1985), which are the periods associated with fluctuations in the eccentricity, obliquity (or tilt), and precession of the Earth's orbit (Hays et al., 1976; Berger, 1978; Imbrie et al., 1984). The length- and resolution-limited record at DSDP Hole 503B exhibits variance at 140,000- and 41,000-year periods (Janecek and Rea, 1985).

At core RC11-210, the interruption in the ongoing variation of the grain-size record at about 300,000 years ago functions as a statistical discontinuity in the long record. Because the timescale constructed for that core is better in the older portion of the record, the interval from 402,000 to 774,000 years ago was selected by Pisias and Rea (1988) for spectral analysis. In this time interval the oxygen isotope record has strong spectral peaks at 100,000- and 23,000-year periods, the expected periodicities during this time when the effect of obliquity is less than during the late Brunhes (Imbrie et al., 1984). The eolian grain-size spectrum shows some power at 100,000 years, but the predominant period in this data set is 31,000 years. Radiolarian-assemblage data that record the strength of equatorial divergence also were subjected to spectral analysis, and these results also display a predominant spectral peak at 31,000 years with lesser power at 100,000 and 23,000 years. The grain size and radiolarian records are coherent and in phase at the predominant periodicity, and provide for the first time a quantitative link between atmospheric and oceanic circulation in the geologic record. Thus the paleoclimatic record of core RC11-210 indicates that stronger winds have resulted in more intense upwelling (Pisias and Rea, 1988).

The 31,000-year spectral peak appears to be a real characteristic of equatorial Pacific phenomena (Pisias and Rea, 1988) and also occurs in paleoclimatic data sets from the Atlantic ocean (Ruddiman et al., 1986a). To explain the predominance of a peak other than that of the primary orbital variability one must call upon both nonlinear interaction of the climate systems to produce the periodicity and some resonance phenomenon to amplify it so it is the predominant peak (Ruddiman et al., 1986a; Pisias and Rea, 1988).

Variability of eolian processes

An important concept to be obtained from the eolian grain-size records of atmospheric-circulation intensity is that wind intensity varies more rapidly than does ice volume. On the time scales being considered in this review, therefore, there is not a direct connection between climate, as depicted by an oxygen-isotope record derived from benthic foraminifera, and intensity of atmospheric circulation. The normal assumption that atmospheric circulation is always stronger during times of increased ice volume and weaker during times of decreased ice volume appears to be incorrect. Ice volume responds to a forcing function over a much longer term than does the atmosphere.

The variability in the flux of dust does seem to change on a time scale that is more close to that of the 100,000-year variation of the oxygen-isotope curve. The two equatorial cores have flux maxima associated with interglacial Stages 5 and 7, and earlier maxima associated with glacial Stages 10 and 12. The proximity of these cores to the ITCZ makes determination of the source of the dust in this area difficult, so the interpretation of this shifting flux pattern is equivocal.

In fact, provenance of the dust in equatorial regions is a major uncertainty in our interpretation of those flux records. If a southern hemisphere source (Prospero and Bonatti, 1969; Raemdonck et al., 1986) is inferred, then the flux record may be related to the paleoclimatology of the northern Andes. Pollen data from that region show the Andean lakes to be full during glacial periods and dry salars or playas during interglacial times (Hooghiemstra, 1984; Van der Hammen, 1985) so the Andean region would supply more dust to the ocean during interglacial times. Alternatively, if the dust distribution processes in the Southern Hemisphere are similar to those of the Northern Hemisphere, then material from an Australian source may dominate dust flux in the entire South Pacific in the same way that material from an Asian source dominates the North Pacific all the way south to the ITCZ (Shaw, 1980; Uematsu et al., 1983).

In the central North Pacific the dust source is obvious, the great arid regions of central and China and Mongolia. The two well-defined flux minima in the KK75-02 record centered at about 120,000 and 550,000 years ago (Fig.4) may correspond to important times of soil formation within the China loess sequences, soils S-1 and S-5 (Liu, 1985; Kukla, 1987). These data are consistent with preliminary information from core V21-146 in the northwestern Pacific that show dust flux maxima associated with the glacial stages (Hovan et al., 1988).

The downcore change in pattern from interglacial-flux maxima to glacial-aged flux maxima in core RC11-210 at about 300,000 years (Fig.2) ago may represent a change in the source area. If the change is from Southern to Northern Hemisphere significant changes in the latitude of the Intertropical Convergence Zone are implied, a change of at least 5° of latitude at RC11-210. It is also possible, but judged less likely, that the shift is from an Andean to an Australian source. The relative timing of Andean wet and dry climatic regimes remains unchanged during the Quaternary (Hooghiemstra, 1984).

The mid-Brunhes climate event

The change in the eolian grain-size records of both the tradewinds and the westerlies between 250,000 and 300,000 years ago is one of the best examples of the mid-Brunhes climatic event (MBCE) available from the deep ocean. This change in the nature of variability of various paleoclimatic proxy indicators was first described by Janecek (1983) and Janecek and Rea (1985) for the eolian grain-size data and by Sancetta and Silvestri (1984) and Schramm (1985) for siliceous plankton data. Pisias and Leinen (1984) described this event in opal flux information from the Northwest Pacific, and Pisias and Rea (1988) have demonstrated the MBCE in radiolarian assemblage data in the equatorial Pacific. Krissek (1988) reported a change in the North Pacific records of ice rafting at 320,000 years ago. This period during the mid-Brunhes also may have been the time when the Arctic Ocean finally acquired a permanent ice cover (Scott et al., 1989). Jansen et al. (1986) have compiled an extensive listing of changes in paleoenvironmental indicators that occurred some time during the middle to lower part of the Brunhes normal chron.

The MBCE, although not well understood, may be an important mechanism for learning more about the interaction among the various climatic subsystems that were described in the "Introduction" section. Evidence of it exists in proxy-indicators of the atmosphere, the ocean surface waters, and perhaps on land, but not in the deep ocean waters or in the ice-volume records. The observation that the MBCE may be limited to oceanic phenomena that occur above the main thermocline emphasizes the importance of that boundary between climatic subsystems.

Climatic transition from early to late Pleistocene

Evidence of an increase in the amplitude of climatic variability about 875,000 years ago has been described from the carbonate and isotope records of long cores obtained from the western Pacific (Shackleton and Opdyke, 1973, 1976) and the Atlantic (Van Donk, 1976; Prell, 1982); this evidence has been interpreted as a rather sudden increase in the nature/severity of glaciations at that time. Investigators attribute this change in the amplitude of glacial–interglacial cycles to changes in the nature of the orbital forcing, particularly an increase in the importance of the 100,000-year cycle about 900,000 years ago (Pisias and Moore, 1981; Ruddiman et al., 1986a, b), but the connection between cause and effect is not yet clear (Pisias and Moore, 1981; Ruddiman et al., 1986a, b).

Core RC11-210 provides the first information about the changes in atmospheric circulation that occurred in conjunction with this change in climatic variability. The increase in amplitude of variability of paleoclimatic proxy indicators from core RC11-210 occurs near the boundary of oxygen-isotope Stages 22 and 23, as seen elsewhere (Shackleton and Opdyke, 1976). At the ages determined for the RC11-210 sediments the increase in $\delta^{18}O$ amplitude begins at 848,000 years ago (Fig.2) and the increase in $CaCO_3$ variability at 855,000 years ago, which is one sample interval earlier (Chuey et al., 1987). The change in eolian grain size, however, begins at 875,000 years ago (Fig.2), which is three sample intervals prior to the increase in the amplitude of the carbonate record and four before the change in the oxygen-isotope record.

These data indicate that at the equatorial location of core RC11-210, the early to late Pleistocene change in climatic variability occurred in the following sequence. Initially, the Southern Hemisphere tradewinds intensified. This was followed 20,000 years (or three sample intervals) later by an increase in the amplitude of the bottom-water $CaCO_3$-dissolution signal. Finally, an increase in the amplitude of the ice volume $\delta^{18}O$ variations occurred. The atmospheric and sea-surface climatic subsystems are the first to respond to this 875,000-year-old change in climatic forcing (Pisias and Rea, 1988), followed many thousands of years later by changes in the deep-water and cryosphere systems.

Summary

Examination of the eolian record at locations in the North Pacific Ocean enables the following insights into the nature of late Pleistocene atmospheric circulation:

(1) Atmospheric circulation responds to orbital forcing. The periodicities corresponding to eccentricity, obliquity, and precession occur in the eolian component of pelagic sediments that accumulate beneath the westerlies. The dominant periodicity in eolian material recovered from beneath the tradewinds is about 31,000 years and is coherent with the radiolarian record of divergence at this frequency.

(2) The variability in the intensity of atmospheric circulation appears to be on higher frequencies than that of the variation in ice volume. There is not an obvious correspondence between greater ice volume (glacial stages) and enhanced wind intensity.

(3) An important change in the variability of the grain-size record of wind intensity occurs about 300,000 years ago, from higher amplitude–longer period variations in older sediment to lower amplitude–shorter period variations in younger material (Figs.2 and 4). This change is the atmospheric expression of the mid-Brunhes climatic event. The MBCE is clearly portrayed in records of atmospheric and sea-surface climatic subsystems and does not seem to occur in deep ocean carbonate records or in the oxygen-isotope ice-volume records. Because this event apparently affects some, but not all, of the climatic subsystems, further examination of it may reveal more about the behavior and interaction of those subsystems.

(4) The increase in amplitude of variation of climatic proxy indicators that delineates the early/late Pleistocene transition near the 22/23 Stage boundary occurs 875,000 years ago in the eolian grain-size record, 20,000 to 25,000 years earlier than the transition in the carbonate-dissolution record of deep-water corrosivity and the $\delta^{18}O$ record of ice volume both of which change at about 850,000 years ago.

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