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LATE CRETACEOUS HISTORY OF EOLIAN DEPOSITION IN THE MID-PACIFIC MOUNTAINS, CENTRAL NORTH PACIFIC OCEAN

DAVID K. REA and THOMAS R. JANECEK

Oceanography Program, Department of Atmospheric and Oceanic Science, The University of Michigan, Ann Arbor, Mich. 48109 (U.S.A.)

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ABSTRACT

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Eolian dust preserved in deep-sea sediment can provide a direct historical record of global atmospheric circulation. Data from a reasonably complete Upper Cretaceous section of pelagic sediments recovered at DSDP Site 463 in the central North Pacific provides a good record of eolian activity during the time period between about 112 and 66 m.y. ago. We have isolated the eolian component from these sediments, determined its mass accumulation rate and combined these data with the mineralogy of the inorganic fraction determined by others to construct a record of eolian deposition. Volcanic input is significant during Aptian—Albian and Maastrichtian times, otherwise continentally derived minerals dominate. Mass accumulation rates of the continental eolian component range from over 500 mg/cm^2/10^{3}yr during the late Albian to a low of 5 mg/cm^2/10^{3}yr during Coniacian time. (For comparison, the upper Miocene to Pleistocene rate averages about 20 mg/cm^2/10^{3}yr.) The temporal pattern of Late Cretaceous eolian accumulation of Site 463 generally matches known changes in sea level, suggesting that source availability is the dominant control of eolian sedimentation during that time.

INTRODUCTION

In the central portions of ocean basins, far from the influence of continental-margin sedimentary processes, the dominant source of the non-authigenic [not formed in place by chemical processes] inorganic component of deep-sea sediments is wind-borne dust. Recent work showing that the bottom nepheloid layer is a result of local entrainment of bottom sediments rather than long-distance transport of materials (I. N. McCave, oral communication) suggests that eolian debris dominates inorganic material in the deep ocean (Leinen et al., 1979) as well as the inorganic component of sediments recovered from undersea ridges and rises (Rea and Janecek, 1981; Rea and Harrsch, 1981). These sediments, therefore, all retain some record of the eolian process. We have been attempting to decipher this record in quantitative terms to determine the history of the global wind systems. Preliminary results for the late Cenozoic (Rea and Janecek, 1980) and entire Cenozoic (Janecek et al., 1980) have been presented elsewhere. This paper concerns results of our work on a nearly-continuous Aptian to Maastrichtian section of pelagic sediments recovered at DSDP Site 463 in the western Mid-Pacific Mountains (Fig.1).

Eolian deposition

The deposition of eolian materials is controlled, as are all sediments, by some combination of a source/supply function and the transport process. Wind-borne sediments may come to the middle of the ocean from local, often volcanic, sources or from distant continents. At times, the supply can be quite large. Clouds of Saharan dust in the central Atlantic have been observed by mariners for centuries. They were encountered by Darwin aboard the *Beagle* (Darwin, 1846) and Maury (1855) devoted a chapter in his book to the phenomenon.

Greater input of eolian materials to the deep oceans would be expected during times of increased global aridity and times of increased volcanic activity. Climatic variations are becoming reasonably well known (Frakes, 1979), and the history of Pacific area volcanic activity is well documented for Neogene time (Kennett et al., 1977; Rea and Scheidegger, 1979). There also are indications of older widespread volcanic pulses during the Eocene (Hein and Vanek, 1981; Rea and Thiede, 1981) and in the Cretaceous during Campanian—Maastrichtian and Aptian—Albian times (Watts et al., 1980; Vallier and Rea, 1980).

The other significant control of eolian deposition is the vigor of the global wind system with stronger winds carrying more and larger particles (Parkin,

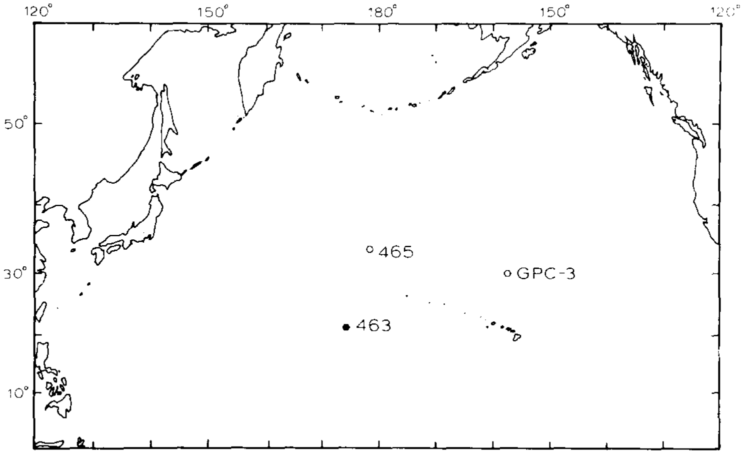

Fig.1. Index map of the North Pacific Ocean showing localities discussed in text: DSDP drill sites 463 and 465, and piston core GPC-3.

1974). Wind strength is related to the earth's pole to equator temperature gradient (Lamb and Woodrolfe, 1970; Newell, 1974). During times of glacial advance this gradient and the resulting winds may become more intense in the temperate zone than elsewhere, as the tropics cool only slightly (CLIMAP, 1976; Gates, 1976; Manabe and Hahn, 1977). When the earth is characterized by expanded warm climates, such as during the Cretaceous (Frakes, 1979), the vigor of atmospheric circulation should be reduced.

Over a decade ago Windom (1969) suggested that up to 75% of the non-biogenic component of deep-sea sediments may be atmospheric dust fallout. Recent estimates of this flux of continental dust to the world's oceans are in the range of 0.5 to 1.0×10^{15} g/yr (Savoie and Prospero, 1980), corresponding to an average mass accumulation rate on the ocean floor of about 150 to 300 $\mathrm{mg/cm^2/10^3yr}$. Particulate accumulation measured in snowfields is of a similar order of magnitude, although generally less than 100 $\mathrm{mg/cm^2/10^3yr}$ (Windom, 1969). The eolian contribution to surface sediments is commonly defined on the basis of the amount of quartz present (Rex and Goldberg, 1958), with patterns of prevailing winds reflected by decreasing quartz concentrations away from the continental source (Lisitzin, 1972; Windom, 1975; Thiede, 1979; Leinen and Heath, 1981). Illite and kaolinite concentrations in surface sediments also reflect eolian transport (Windom, 1975).

Downcore studies of eolian materials have been relatively few. Parkin and co-workers (Parkin, 1974; Parkin and Shackleton, 1973; Parkin and Padgham, 1975) studied two cores from the African continental margin and found that both the abundance and size of quartz grains larger than 7 μm increased during glacial times. Their data extend back through Brunhes time and suggests that the tradewinds were stronger during glacial times. Thiede (1979) has shown that the region of high quartz concentration east of Australia and New Zealand was much more extensive during the last glaciation than it is now, strongly implying that the Southern Hemisphere westerlies were much stronger during glacial times. Piston core LL44-GPC3 recovered from the abyssal Pacific 1000 km north Hawaii (Fig.1) appears to contain a complete Cenozoic record of eolian deposition (Leinen and Heath, 1981). In that core, the mass accumulation rate of quartz increased by almost an order of magnitude from mid-Pliocene to late Quaternary time, again reflecting the intensification of wind associated with the onset of the present ice age in the late Pliocene.

The basic data set for our effort is the mass accumulation rate (MAR) of the *total* eolian load, not just of quartz which may be only 20—30% (Windom, 1969) or even less (Prospero and Bonatti, 1969) of the wind-transported sediment. We isolated the eolian component by a series of selective extractions that remove the biogenic and authigenic sedimentary components. If the sample site is far from land or on some topographic elevation, we assume that the non-authigenic, inorganic sedimentary component isolated represents eolian material, and that changes in its accumulation rate reflect fluctuations in the eolian process.

EOLIAN RECORD IN THE CENTRAL PACIFIC

Sediments and biostratigraphy

DSDP Site 463 is located in the western Mid-Pacific Mountains in 2525 m of water at 21°21.01'N, 174°40.07'E (Fig.1). Four lithologic units of Pleistocene to late Barremian age are recognized. Unit I is divided into two subunits: subunit IA is 46.8 m of highly disturbed to soupy nannofossil ooze of Pleistocene to Maastrichtian age; IB consists of foraminifer—nannofossil chalk and nannofossil—foraminifer chalk from 46.8 m to 452 m subbottom and from Maastrichtian to late Albian age. Chert is a common component of Unit IB. Unit II, from 452 to 587 m, is comprised of multicolored limestone and silicified limestone of middle Albian to early Aptian age. Chert is also common in Unit II. Unit III is characterized by chert layers, ash beds, tuffaceous limestone, and carbonaceous limestone 587.7 to 632.5 m subbottom, of early Aptian age. Unit IV consists of interbedded pelagic and clastic limestone of early Aptian to late Barremian age from 632.5 to 822.5 m at the bottom of the hole. Within this section hiatuses occur during late Miocene to Oligocene, Oligocene to middle Eocene, early Eocene to late Maastrichtian, and early Campanian to Coniacian time (Thiede, Vallier et al., 1981).

The results presented below are from the late Maastrichtian to Aptian part of the section, Units IB, II, and III. Much of Unit IV was emplaced by downslope transport which masks the eolian input, thus rendering those sediments unsuitable for the purposes of this study.

Sediment ages are based upon the biostratigraphy given by Thiede, Vallier et al. (1981). Absolute ages of various horizons in the sedimentary column were obtained by comparing the foraminifer and nannofossil zones to the time scale of Van Hinte (1976). Linear sedimentation rates (LSR) were then calculated between depth horizons of known age.

Mass accumulation rates of the eolian component

In order to quantify the mass flux of any component to the sediment column it is necessary to determine the mass accumulation rate (MAR) of the total sediments and the weight percent of the sediment component of interest and find their product. The MAR is the product of the LSR and the dry bulk density (DBD) of the sediment. This is determined by weighing, freeze-drying, and reweighing a fresh sediment sample. Then MAR $(g/cm^2/10^3yr)$ = LSR $(cm/10^3yr)$ × DBD (g/cm^3). The weight percent of the inorganic, non-authigenic, presumably eolian fraction is measured by treating the freeze-dried sample successively with acetic acid to remove calcium carbonate, a buffered sodium dithionite—sodium citrate solution to remove oxides, hydroxides and zeolites, and sodium carbonate to remove opal. The residue of these processes is again freeze-dried and weighed to give the weight percent of the eolian component. Authigenic clays and feldspars would survive this procedure, but examination of SEM photographs taken during various stages of the extraction process and the general coherency of the data suggest that they do not present a problem. The process is described in detail by Rea and Janecek (1981).

We have determined the total MAR, weight percent of eolian material, and eolian MAR for sixty samples from the Cretaceous pelagic sediments of DSDP Site 463 (Table I, Fig.2). The precision of the weight-percent data is good, replicate analyses are commonly with 2 or 3%, with a few within only 10% of each other. The accuracy of the resulting eolian MAR data is more likely limited by the time scales and assumptions by which the LSR is calculated.

Mineralogy of the eolian component

The clay mineralogy of the inorganic fraction of the sediments can be used as a source indicator. Illite-dominated assemblages indicate a general continental source of material while those dominated by smectite or mixed-layer clays suggest a volcanic source. Several authors have determined the mineralogy of the Site 463 sediments (Nagel and Schumann, 1981; Hein and Vanek, 1981; Rateev et al., 1981; and Vallier and Jefferson, 1981). Their combined results give 59 analyses in the Late Cretaceous section of Site 463 and clearly define the times when continental and volcanic materials dominate the eolian input (Table II, Fig.2).

HISTORY OF EOLIAN DEPOSITION

The information compiled here and presented on Fig.2 is a record of eolian deposition in the central Pacific Ocean during Aptian through Maastrichtian time. A backtrack plot for Site 463 (according to Lancelot, 1978) shows that it lay beneath the zone of southeast tradewinds during Cretaceous times, crossing beneath the equator during the Maastrichtian. The position of Site 463 beneath the southeast tradewinds requires that the source of continentally derived dust be the South American and perhaps African continental masses 10,000 to 15,000 km to the east (90 m.y. reconstruction of Firstbrook et al., 1979).

The mineralogical data reveal two periods of dominant volcanogenic input, one during Aptian and Albian time and the second during the early Maastrichtian. The older event is far more intense at Site 463, resulting in the ash layers of sedimentary Unit III and eolian MAR's in excess of 1000 mg/cm^2/10^{3}yr (Table I, Fig.2). During early Maastrichtian time smectites again dominate the clay minerals although the volcanogenic component remains diffuse within the pelagic carbonates. MAR's of the eolian fraction reach over 300 mg/cm^2/10^{3}yr during this time (Table I). These horizons record the two episodes of significant Cretaceous volcanic activity in the western Pacific (Vallier and Rea, 1980; Watts et al., 1980).

TABLE I

Accumulation of eolian sediments; data determinations explained in text; samples closely spaced in time are bracketed, averaged, and plotted on Fig. 2 as a single point

Depth in core (m)	Age (m.y.)	DBD (g/cm^3)	LSR (cm/10^{3}yr)	MAR (g/cm^2/10^{3}yr)	Eolian fraction (wt. %)	Eolian MAR (mg/cm^2/10^{3}yr)	Eolian MAR plotted on Fig.2
48.5	66.0	1.19	5.42	6.45	1.68	108	106
56.3	66.2	1.13	5.42	6.12	1.14	70	
64.9	66.3	1.13	5.42	6.12	1.14	70	
76.2	66.5	1.20	5.42	6.50	2.75	177	
79.2	66.6	1.34	5.42	7.26	10.30	748	425
84.7	66.7	1.35	5.42	7.32	1.42	104	
91.4	66.8	1.37	5.42	7.43	6.75	502	
95.9	66.9	1.30	5.42	7.05	4.89	345	
102.5	67.1	1.23	1.10	1.35	2.50	34	41
107.0	67.5	1.27	1.10	1.40	4.96	69	
112.0	68.0	1.31	1.10	1.44	1.35	19	
115.2	68.3	1.19	1.10	1.31	1.36	18	26
119.9	68.7	1.21	1.10	1.33	2.64	35	
126.4	69.1	1.25	5.45	6.81	1.47	100	158
131.4	69.2	1.27	5.45	6.92	2.66	184	
135.9	69.2	1.39	5.45	7.58	1.86	141	
145.6	69.4	1.29	5.45	7.03	2.95	207	
159.7	69.7	1.27	5.45	6.92	5.74	397	294
164.2	69.8	1.25	5.45	6.81	5.14	350	
167.7	69.8	1.28	5.45	6.98	1.94	135	
178.5	70.1	1.25	1.22	1.52	1.76	27	30
183.1	70.5	1.22	1.22	1.49	2.20	33	
188.6	70.9	1.34	1.22	1.63	1.93	31	
191.4	71.1	1.29	1.22	1.57	1.39	22	30
196.3	71.5	1.16	1.22	1.42	2.74	39	
201.1	71.9	1.44	1.22	1.76	1.57	28	
202.7	72.2	1.35	0.47	0.63	1.52	10	10
206.8	73.0	1.17	0.47	0.56	1.27	7	7
217.1	82.4	1.22	0.30	0.37	1.42	5	5
221.4	83.8	1.21	0.30	0.36	2.20	8	8
225.7	84.4	1.17	0.97	1.13	3.96	45	45
243.3	86.2	1.21	0.97	1.17	5.40	63	63
254.3	87.3	1.53	0.97	1.48	6.23	92	92
262.1	87.9	1.43	1.52	2.17	6.49	141	141
282.3	89.2	1.25	1.52	1.90	3.13	59	59
292.3	89.9	1.34	1.52	2.04	2.93	60	60
309.5	90.7	1.70	2.25	3.82	7.65	292	292
329.2	91.6	1.61	2.25	3.62	5.56	201	201
378.3	95.7	1.52	1.06	1.61	17.36	279	279
382.6	96.1	1.64	1.06	1.74	7.08	123	123
426.6	100.4	1.48	0.60	0.89	3.69	33	33
433.5	101.1	1.62	2.50	4.05	11.37	460	460
462.8	102.6	1.81	1.40	2.53	20.16	510	510
481.2	103.9	1.46	1.40	2.04	27.77	566	566
490.9	104.6	1.79	1.40	2.51	7.95	200	200
501.0	105.3	1.93	1.40	2.70	1.16	31	31
510.2	105.9	2.09	1.40	2.93	11.81	346	350
513.3	106.2	2.04	1.40	2.86	12.36	353	
522.5	106.8	2.20	1.40	3.08	12.51	385	385
535.8	107.8	2.37	1.40	3.32	16.56	550	489
538.1	107.9	2.13	1.40	2.98	19.25	574	
541.0	108.1	2.09	1.40	2.93	11.68	342	
547.5	108.6	2.13	1.40	2.98	13.97	416	416
558.5	109.4	1.96	1.40	2.74	3.29	90	90
567.1	110.0	2.00	2.50	5.00	9.75	487	487
585.9	110.8	2.10	2.50	5.25	14.02	736	736
606.0	111.6	1.99	2.50	4.98	5.86	292	292
615.0	111.9	1.93	2.50	4.82	32.73	1578	1578
622.6	112.2	1.63	2.50	4.08	56.97	2324	1412
626.3	112.4	1.79	2.50	4.48	11.16	500	

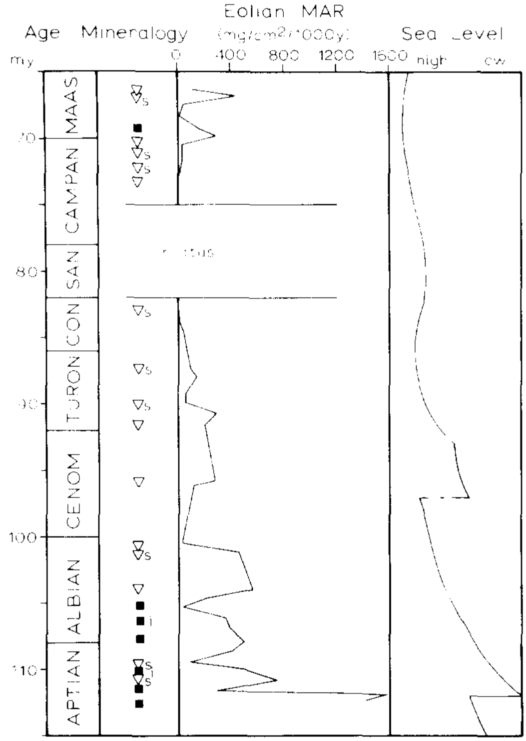

Fig.2. Age, mass accumulation rate, and clay mineralogy of eolian sediments at DSDP Site 463. Relative sea-level curve of Vail et al. (1977) is at right. Open triangles and *i*'s show dominant and secondary illite; closed squares and *s*'s represent dominant and secondary smectite, respectively.

Most of the Late Cretaceous section (see Fig.2 for absolute ages) records eolian deposition from a continental source. The rate of eolian deposition fluctuates during the Maastrichtian. For about 20 m.y., Campanian through mid-Turonian, the eolian MAR is low, reaching values as low as 5 mg/cm^2/10^{3}yr. During early Turonian and late Cenomanian time the eolian MAR exceeds 200 mg/cm^2/10^{3}yr. Early Cenomanian eolian sediment has an apparently low MAR, although the data for this interval are sparse. The influx of continental dust to the Pacific reached a maximum during middle to late Albian time with MAR values exceeding 500 mg/cm^2/10^{3}yr and may also have been high during the late Aptian, although those rates are complicated by the Aptian volcanic input (Table I, Fig.2).

The general nature of Cretaceous climate was one characterized by high temperatures and high aridity (Frakes, 1979). Evaporite deposits were formed to 55° latitude and coals to 70°; reefs and deserts were extensive. The pole to equator temperature gradient is estimated to have been about half of the present value, and remained constant throughout the period (Douglas and Savin, 1975). The Cretaceous period experienced the warmest

TABLE II

Mineralogy of Eolian sediments

Depth in core (m)	Age (m.y.)	Mineralogy, dominant clay	Reference	Mineralogy plotted on Fig.2
54.9	66.2	Is	1	I
64.5	66.3	I	1	
77.4	66.6	I	1	
96.1	66.9	I	1	Is
99.4	67.0	Is	2	
139.2	69.3	Si	2	Si
142.3	69.4	Is	1	
180.8	70.3	I	1	I
190.1	71.0	Is	1	Is
203.1	72.2	Is	1	Is
207.8	73.2	I	2	I
218.8	82.9	Is	1	Is
253.6	87.2	Is	1	Is
294.2	90.0	Is	1	Is
328.6	91.5	I	2	I
328.8	91.5	I	1	
379.4	95.8	I	1	I
427.0	100.5	I	1	I
433.4	101.1	Is	1	Is
481.5	103.9	I	1	I
500.2	105.2	S	1	S
511.5	106.0	Is	2	Si
512.1	106.1	Si	1	
518.6	106.5	Si	2	
528.7	107.3	S	3	S
535.7	107.8	S	3	
536.6	107.8	S	1	
539.2	108.0	S	1	
539.3	108.0	S	2	
539.7	108.0	S	3	
540.0	108.1	S	3	
540.2	108.1	S	3	
559.6	109.5	Is	3	Is
567.7	110.0	Si	2	Si
567.8	110.0	Si	2	
577.8	110.4	Is	1	Is
586.5	110.8	Is	2	
587.2	110.8	S	3	
587.6	110.8	I	1	
596.7	111.2	S	3	S
605.4	111.5	S	1	
605.5	111.5	S	4	
607.5	111.6	Si	3	
607.6	111.6	S	2	
614.0	111.9	S	2	
614.5	111.9	S	3	

TABLE II (*continued*)

Depth in core (m)	Age (m.y.)	Mineralogy, dominant clay	Reference	Mineralogy plotted on Fig. 2
614.5	111.9	Is	1	
617.6	112.0	S	2	
620.6	112.2	S	2	
621.4	112.2	S	2	
622.8	112.2	S	2	
622.8	112.2	S	1	
623.0	112.3	S	3	S
623.2	112.3	S	3	
623.2	112.3	S	2	
624.0	112.3	Si	2	
625.2	112.3	Si	2	
625.2	112.3	Si	3	
627.2	112.4	Si	1	

Combined data on the dominant and secondary clay minerals, I, i = illite, S, s = smectite. Samples closely spaced in time are bracketed, averaged and plotted on Fig.2 as a single point. References are: 1 = Nagel and Schumann (1981); 2 = Rateev et al. (1981); 3 = Vallier and Jefferson (1981); and 4 = Hein and Vanek (1981).

and driest global climate during all of Phanerozoic time, reaching a climatic optimum sometime during Albian to Santonian time with cooling since then and fluctuations in the Maastrichtian (Frakes, 1979).

During a period of unchanging temperature gradients, the eolian transport process would be expected to remain uniform. Changing eolian accumulation rates, therefore, should reflect the changing supply of wind-blown dust. The widespread Cretaceous transgressions and regressions would have significantly altered the source availability of eolian material with times of maximum transgression corresponding to lowest eolian MAR. Vail et al. (1977) have presented sea-level curves illustrating the general timing and degree of transgressions and regressions during Mesozoic and Cenozoic time. Their data show a Hauterivian to mid-Aptian transgression followed by a sharp, brief regression, then a second transgression continuing to Cenomanian time. Another regression occurs during the latter Cenomanian followed by a transgression that continued until the middle Turonian. The ensuing 25 m.y. from mid-Turonian through Maastrichtian time was a period of constant, widespread epicontinental seas that remained until the late Maastrichtian—Paleocene regression (Vail et al., 1977, their fig.2; Pitman, 1978).

The temporal pattern of land-derived eolian sedimentation at Site 463 (Fig.2) matches these sea-level fluctuations reasonably well. The early to mid-Aptian horizon of illite-dominated material occurs at the time of the sharp mid-Aptian regression. The late Albian sharp decrease in eolian deposition may reflect the transgression of Albian to early Cenomanian age. The mid-Cenomanian regression apparently corresponds to an increase in eolian MAR and the ensuing reduction in accumulation from that time to the

Coniacian and continuing low values correspond to the Cenomanian—Turonian transgression and continuing highstand (Fig.2).

The actual mass flux of eolian material to the oceans is surprisingly high, averaging about 200 mg/cm^2/10^{3}yr from Albian to Turonian time, 25 mg/cm^2/10^{3}yr from Coniacian to Campanian time, and 175 mg/cm^2/10^{3}yr during the Maastrichtian. Maastrichtian values are influenced by the uppermost Cretaceous volcanic episode (Vallier and Rea, 1980) and are an order of magnitude greater than 65—70 m.y. eolian MAR's on the Hess Rise (DSDP Site 465, Rea and Harrsch, 1981) or in the northeast Pacific (Core GPC-3, 30.3°N, 157.6°W, our unpublished data; Fig.1). Other post-middle Albian data appear to be a true reflection of continental eolian input.

For comparative purposes the eolian MAR values for the late Miocene and Pleistocene sediments recovered at DSDP Site 463 are: for the late Miocene (average of 7 samples, 5—10 m.y. old), 9 mg/cm^2/10^{3}yr; for the Pliocene (7 samples, 2—4.5 m.y. old), 21 mg/cm^2/10^{3}yr; and for the Pleistocene (2 samples, 1.0—0.2 m.y. old), 44 mg/cm^2/10^{3}yr (Rea and Janecek, 1981). These values for Neogene deposition appear to reflect the increasing aridity and the growing intensity of the global wind system as the pole to equator temperature gradient increased during the later Cenozoic (Savin et al., 1975; Frakes, 1979), but are much lower than the values for Late Cretaceous eolian deposition.

Extant data suggest that the Cretaceous temperature gradient, and thus overall wind intensity, was similar to or less than that of the late Miocene (Douglas and Savin, 1975; Savin et al., 1975). The longitudinal distance of Site 463 from the nearest source of eolian dust was 11,500 km 90 m.y. ago (Firstbrook et al., 1979) compared to about 8500 km now, probably not a significant change. The relatively high mid-Cretaceous MAR of eolian material, therefore, is most likely the result of a greatly increased supply of dust, perhaps an order of magnitude more than the late Cenozoic average. This atmospheric dust, a natural result of the previously documented Cretaceous aridity, might reasonably be expected to have occasioned a global cooling trend since the earth's albedo increases with the concentration of atmospheric particulates (Lamb, 1970; Frakes, 1979; Bray, 1979). Various types of data suggest that such a cooling trend may have begun as early as Albian time (Douglas and Savin, 1975; Frakes, 1979) when atmospheric dust loadings were high (Fig.2). We do not yet understand the relative significance of the mid-Cretaceous atmospheric dust concentrations to the onset of global cooling, however, as many other factors such as cloudiness, humidity, CO_2 and O_3 concentrations, continental vegetation, and the area covered by low-reflectivity water also bear upon the determination of the earth-surface temperature (Pollack et al., 1976; Frakes, 1979; Barron et al., 1980).

SUMMARY

Isolation and quantification of the eolian component of pelagic sediments from the north Central Pacific Ocean has enabled a reconstruction of the Late Cretaceous history of eolian sedimentation. Smectite-dominated eolian materials denote periods of volcanic activity during Aptian to early Albian and early Maastrichtian time. At other times illite is the dominant clay mineral, indicative of a continental source of the eolian component. Mass accumulation rates of the illite-dominated component range over two orders of magnitude from over 500 $mg/cm^2/10^3yr$ in late Albian time to a low of 5 $mg/cm^2/10^3yr$ during the Coniacian. In general, the eolian MAR appears to be a reflection of global sea level (Fig.2) with times of low eolian MAR corresponding to the Cretaceous transgressions and highstands. This correspondence, coupled with the suggestion of relatively sluggish and unchanging oceanic and atmospheric circulation, implies that the rate of eolian sedimentation during the Cretaceous was determined by the amount of dust available for transport rather than by the transporting process itself.

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