# 25. MASS-ACCUMULATION RATES OF THE NON-AUTHIGENIC INORGANIC CRYSTALLINE (EOLIAN) COMPONENT OF DEEP-SEA SEDIMENTS FROM THE WESTERN MID-PACIFIC MOUNTAINS, DEEP SEA DRILLING PROJECT SITE 463<sup>1</sup>

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### ABSTRACT

Elevated regions in the central parts of ocean basins are excellent for study of accumulation of eolian material. The mass-accumulation rates of this sediment component appear to reflect changes in the influx of volcanic materials through the Early Cretaceous to Recent history of Deep Sea Drilling Project Site 463, on the Mid-Pacific Mountains. Four distinct episodes of eolian accumulation occurred during the Cretaceous: two periods of moderate accumulation, averaging about 0.2 to  $0.3 \text{ g/cm}^2/10^3 \text{ yr}$ , 67 to 70.5 m.y. ago and 91 to 108 m.y. ago; a period of low accumulation, approximately  $0.03 \text{ g/cm}^2/10^3 \text{ yr}$ , 70.5 to 90 m.y. ago; and a period of high accumulation, about  $0.9 \text{ g/cm}^2/10^3 \text{ yr}$ , 109 to 117 m.y. ago (bottom of the hole). Much of the Cenozoic section is missing from Site 463. Upper Miocene to Recent sediments record an upward increase in accumulation rates, from less than 0.01 to about  $0.044 \text{ g/cm}^2/10^3 \text{ yr}$ . The late Pliocene-Pleistocene peak may reflect the change to glacial-wind regimes, as well as an increase in volcanic source materials.

## INTRODUCTION

# **DSDP Site 463**

Site 463 is in 2525 meters of water on the northwest Mid-Pacific Mountains, at 21°21.01'N, 174°40.07'E (see Site 463 report, this volume). The Mid-Pacific mountains (Fig. 1), one of the large aseismic rises in the central and northern Pacific Ocean, consists of several segments which are thought to be part of an ancient fracture-zone and ridge-crest system which separated the North and South Pacific plates during the Early Cretaceous (Larson, 1976).

Four major lithologic units are recognized in the 822.5 meters of sediment and rock cored at Site 463 (Fig. 2). Unit I is divided into two sub-units: Sub-unit IA is 46.8 meters of highly disturbed to soupy nannofossil ooze (Pleistocene to late Maastrichtian); Sub-unit IB consists of foraminifer-nannofossil chalk and nannofossil-foraminifer chalk from 46.8 to 452 meters subbottom (late Maastrichtian to late Albian). Chert is a common component of Sub-unit IB. Unit II, from 452 to 587 meters, comprises multicolored limestone and silicified limestone (middle Albian to early Aptian). Chert is also common in Unit II. Unit III is characterized by chert layers, ash beds, tuffaceous limestone, and carbonaceous limestone, from 587.7 to 632.5 meters (early Aptian) age. Unit IV consists of interbedded pelagic and clastic limestone from 632.5 to 822.5 meters at the bottom of the hole (early Aptian to late Barremian). Within this section, there are two major hiatuses, representing the middle Miocene to late Oligocene (~16 m.y.) and early Eocene to late Maastrichtian (~17 m.y.).

### BACKGROUND AND OBJECTIVES

The main objective of DSDP Leg 62 was to investigate the late Mesozoic and Cenozoic paleoenvironment of the north-central Pacific Ocean. The sites drilled during Leg 62 are on structural highs that have remained above the CCD throughout their existence, thus presumably allowing preservation of calcareous sediments. In contrast, calcareous sediments have been dissolved over wide regions of the deeper North Pacific basins. The late Mesozoic and Cenozoic calcareous sediments capping the Mid-Pacific Mountains could provide insight into changing plankton communities, biological productivity, and ocean circulation during the past 118 m.y. Furthermore, we hoped that the continuously cored sediments at Site 463 would reveal information on eolian contributions to the North Pacific sediments during the late Mesozoic and Cenozoic.

Oceanographers have long known that terrigenous dusts are blown over the oceans from continents (Maury, 1855; Radczewski, 1939; Rex and Goldberg, 1962; Prospero and Bonatti, 1969; Windom, 1975; Windom and Chamberlain, 1978). Fine-grained quartz and clays, important components of these dusts (Prospero and Bonatti, 1969; Parkin, 1974), can be mapped in pelagic surface sediments; they reflect the major zonal wind regimes (Rex and Goldberg, 1962; Lisitzin, 1972; Molina-Cruz, 1977; Thiede, 1979).

The North Pacific aseismic rises (Hess, Shatsky, and the plateau level of the Mid-Pacific Mountains) are among the best locations for studying the history of eolian deposition, for several reasons. First, these regions probably have always been above the bottom nepheloid layer, which is generally confined to depths in excess of 4500 meters. Second, these sites are far from continental margins, and thus beyond the realm of hemipelagic deposition. Finally, these rises lay south of

<sup>&</sup>lt;sup>1</sup> Initial Reports of the Deep Sea Drilling Project, Volume 62.



Figure 1. Map of the central Pacific, showing the Leg 62 track line and DSDP drill sites.

the polar front (subpolar convergence) during both the nonglacial and glacial ages of the Pleistocene (CLIMAP Project Members, 1976), and thus were not subject to deposition by ice-rafting. Therefore, the only likely non-authigenic, crystalline, inorganic materials are dust grains carried by the winds. Eolian dust may come from distant continents or local volcanoes. The mass-accumulation rates given hereafter record changes in both supply rate and transport efficiency; we are unable to determine which is the dominant control on accumulation rate from the rate data alone.

This paper is part of an initial attempt to establish the late Mesozoic and Cenozoic record of eolian deposition in the central North Pacific, determined from the mass



Figure 2. Stratigraphic column for Site 463 (Lithologic symbols are explained elsewhere in this volume.)

accumulation rates of the non-authigenic, inorganic, crystalline material isolated from the deep-sea sediments. Fluctuations in the mass-accumulation rates of this component should record variations of wind intensity and (or) periods of volcanic activity in the central Pacific.

### METHODS

Mass-Accumulation-Rate Calculations. Linear sedimentation rates (LSR) and mass-accumulation rates (MAR) both give temporal measures of sediment accumulation on a substrate. Mass-accumulation rates are expressed as mass of sedimentary material accumulating per unit area and unit time; they indicate the flux of material across the sediment/water interface. The LSR, which measures only thickness of sediment deposited per unit time, cannot be used to determine mass flux (see discussion in van Andel et al., 1975). Furthermore, the MAR is not affected by variable compaction, which can cause a significant change in the LSR. Thus, using MAR it is possible to compare sediment accumulation on both a temporal and areal basis. Calculation of the MAR requires data on the LSR and dry-bulk density (DBD) for each interval of the core. MAR is then the product of the LSR and DBD. Furthermore, if the weight percent of any component of the sediment is known, the MAR of that component is the product of the total MAR and the weight percent of that component.

LSR's were obtained from Leg 62 revised shipboard paleontological information (see Site 463 report, this volume). Core intervals were assigned ages in million-year intervals by combining sediment ages from planktonic-microfossil zonations and the appropriate time scales (Berggren and van Couvering, 1974; van Hinte, 1976; Hardenbol and Berggren, 1978) and interpolating when necessary. Core intervals were then assigned the appropriate LSR.

The dry-bulk density of each sediment sample was calculated by determining the porosity of the sediment and solving an equation of the form:

$$DBD = (1 - P/100)\varrho$$

where

P is the porosity of the sediment in percent

 $\varrho$  is the density of the sediment (assumed to be 2.65 g/cm<sup>3</sup>).

Porosity is defined as the ratio in percent of the volume of voids to the total volume of the wet sediment mass and can be calculated according to the following equation:

$$P = \frac{V_{\rm sw}}{(V_{\rm sw} + V_{\rm sed})} \times 100$$

where

 $V_{sw}$  is the volume of seawater  $V_{sed}$  is the volume of dry sediment.

To determine the above volume terms, the water content of each sediment sample was calculated by weighing the wet sediment, freeze drying, then reweighing the sediment. The difference between the initial and the final weights equals the weight of pure water lost; the final weight being the weight of the dry sediment plus salt. These values were substituted into the following equations:

$$V_{sw} = \frac{(Wt. water lost) \times 1.025 \text{ } g_{sw}/g_{water}}{1.025 \text{ } g_{sw}/cm^3} = \frac{Wt. water lost}{1.00 \text{ g/cm}^3}$$
$$V_{sed} = \frac{Wt. dry \text{ sediment}}{2.65 \text{ g/cm}^3} = \frac{Wt. \text{ wet sediment} - 1.025 (Wt. water)}{2.65 \text{ g/cm}^3}$$

Result of these calculations are given in Table 1.

**Chemical Methods.** A split of 1 to 10 grams of the freeze-dried sediment was used for the isolation of the non-authigenic, inorganic, crystalline component. The carbonate portion was removed by placing the pre-weighed, dried sediment in an Ehrlenmeyer flask with 100 ml of 25% (vol.) acetic acid. The flask was shaken on an automatic

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Table 1. Data for calculation of mass-accumulation rates (MAR) for the non-authigenic, inorganic, crystalline, eolian (NICE) component of sediments recovered at DSDP Site 463.

Sample (interval in cm)	Sub-bottom Depth (m)	Age (m.y.)	Porosity (%)	DBD (g/cm <sup>3</sup> )	LSR (cm/10 <sup>3</sup> yr)	MAR (g/cm <sup>2</sup> /10 <sup>3</sup> yr)	NICE (%)	NICE MAR (g/cm <sup>2</sup> /10 <sup>3</sup> yr)
463-1-1, 20-24	0.2	0.2	56.84	1.14	0.35	0.400	10.25	0.041
1-2, 70-72	2.2	1.0	59.74	1.07	0.35	0.374	12.95	0.048
2-1, 20-24	5.7	2.0	58.38	1.10	0.35	0.386	6.25	0.024
2-2, 94-96	7.9	2.5	64.22	0.95	0.35	0.332	3.20	0.011
2-4, 92-94	10.9	3.0	68.02	0.85	0.35	0.297	3.81	0.011 7
2-5, 100-101	12.5	3.0	60.55*	1.05*	0.35	0.368	6.24	0.023*J
3-1, 10-13	14.0	4.5	58.37	1.24	0.35	0.386	4.20	0.032
3-2, 72-74	17.2	5.0	62.01	1.01	0.35	0.353	3.64	0.013
3-4, 15-19	19.8	5.5	58.37	1.10	0.35	0.384	4.09	0.016
3-5, 72-74	21.7	6.0	58.20	1.10	0.35	0.385	2.05	0.008
4-2, 102-104	24.6	8.5	58.77	1.13	0.35	0.394	3.64	0.014
4-4, 20-23	29.2	9.5	62.11	1.00	0.35	0.350	0.64	0.002
4-5, 102-104	31.5	10.0	60.96	1.03	0.35	0.360	1.65	0.006
5-2, 58-60	36.1	25.0	54.75	1.20	0.05	0.060	0.54	
6-5, 68-70	40.2	32.0	49.24	1.34	0.05	0.067	0.36	
7-2, 70-72	45.7	67.5	49.91	1.33	5.20	6.916	0.65	0.045 7
7-4, 46-48	48.5	67.5	55.20	1.19	5.20	6.188	1.68	0.102
8-3, 30-32	56.3	67.5	57.51	1.13	5.20	5.876	1.14	0.067
9-2, 90-92	76.2	68.0	54.84	1.13	5.20	5.876	2 75	0.085
10-5, 118-120	79.2	68.5	49.51	1.34	5.20	6.968	10.30	0.717
11-3, 16-17	84.7	68.5	48.91*	1.35*	5.20	7.020	1.42	0.100*
12-1, 40-41	91.4	68.5	48.32	1.37	5.20	7.124	6.75	0.481
12-4, 41-42	95.9	68.5	53.69	1.30	5.20	6.760	4.89	0.330 ]
13-5, 50-52	107.0	69.0	52.05*	1.27*	5.20	6.610	4.96	0.328*
14-2, 72-74	112.0	69.0	50.41	1.31	5.20	6.812	1.35	0.095
14-4, 72-74	115.2	69.0	54.85	1.19	5.20	6.188	1.36	0.084
15-2, 86-88	119.9	69.5	54.42	1.21	5.20	6.292	2.64	0.166
16-2, 88-90	131.4	69.5	51.88	1.27	5.20	6.604	2.66	0.176
16-5, 88-90	135.9	69.5	47.68	1.39	5.20	7.228	1.86	0.134
17-5, 110-112	145.6	70.0	51.26	1.29	5.20	6.708	2.95	0.198
19-2, 70-72	159.7	70.0	52.18	1.27	5.20	6.604	5.74	0.348
20-1, 75-77	167.7	70.0	51.53	1.23	5.20	6.656	1.94	0.129
21-2, 51-53	178.5	70.5	52.76	1.25	0.50	0.625	1.76	0.011
21-5, 61-63	183.1	70.5	53.96	1.22	0.50	0.610	2.20	0.013 ]
22-2, 112-114	188.6	71.0	49.57	1.34	0.50	0.670	1.93	0.013
23-1, 78-80	191.4	72.0	56.07	1.16	0.50	0.580	2.74	0.016
24-2, 9-11	201.1	73.0	45.73	1.44	0.50	0.720	1.57	0.011
24-3, 22-24	202.7	73.0	48.90	1.35	0.50	0.675	1.52	0.010
25-2, 32-34	206.8	74.0	55.65	1.17	0.50	0.585	1.27	0.007
26-2, 106-108	217.1	80.5	54.27	1.22	0.60	0.732	2.20	0.010
27-2, 18-20	225.7	81.0	55.96	1.17	0.60	0.702	3.96	0.028 J
29-1, 24-26	243.3	81.0	54.34	1.21	1.00	1.210	5.40	0.065 -
30-2, 25-27	254.3	82.0	42.35	1.53	0.75	1.147	6.23	0.071
33-1, 123-125	282.1	80.0	40.00	1.43	0.50	0.625	3.13	0.045
34-2, 30-32	292.3	89.0	49.41	1.34	1.80	2.412	2.93	0.071
36-1, 1-3	309.5	91.0	35.97	1.70	1.80	3.060	7.65	0.234
38-1, 65-67	329.2	92.0	39.24*	1.61*	1.80	2.900	5.56	0.161*
43-5, 61-62	382.6	99.0	37.88	1.64	1.80	2.952	7.08	0.209
48-2, 49-51	426.6	100.0	43.99	1.48	1.80	2.664	3.69	0.098
50-1, 49-51	433.5	101.0	38.68	1.62	1.20	1.944	11.37	0.221
53-1, 130-131	462.8	104.0	31.86	1.81	1.20	2.172	20.16	0.438
56-1, 89-90	490.9	106.0	32.38	1.40	1.20	2.148	7.95	0.171
57-2, 45-46	501.0	107.0	27.02	1.93	1.20	2.316	1.16	0.027
58-1, 119-120	510.2	107.0	21.00	2.09	1.20	2.508	11.81	0.296
58-3, 129-130	513.3	107.0	22.98	2.04	1.20	2.448	12.36	0.303
61-1, 28-29	535.8	109.0	10.46	2.37	2.70	6.399	16.56	1.060
62-1, 55-56	538.1	110.0	19.38	2.13	2.70	5.751	19.25	1.107 J
62-3, 46-48	541.0	110.0	21.13	2.09	2.70	6.643	11.68	0.659
63-1, 50-51	547.5	111.0	19.71	2.13	2.70	5.751	13.97	0.803
65-1, 105-107	567.1	112.0	23.90	2.00	2.70	5.400	9.75	0.526
67-1, 84-86	585.9	112.0	20.90	2.10	2.70	5.670	14.02	0.795
69-2, 50-52	606.0	112.0	24.80	1.99	2.70	5.373	5.86	0.315
70-2, 1-2	615.0	112.0	27.03	1.93	2.70	5.211	32.73	1.705
71-3 31-33	626.3	113.0	32 33*	1.03	2.70	4,401	11.16	0.540*
72-2, 106-107	635.1	113.0	26.12	1.95	3.10	6.045	31.92	1.930
72-3, 127-128	636.8	113.0	32.76	1.78	3.10	5.518	15.40	0.850
73-1, 20-21	642.2	113.0	17.35	2.19	3.10	6.789	11.56	0.785
74-2, 11-12	661.6	114.0	24 20	2.21	3.10	6.372	4.98	0.327
76-1, 10-11	670.6	114.0	26.97	1.93	3.10	5.983	22.16	1.326
77-1, 0-1	680.0	114.0	16.73	2.20	3.10	6.820	14.41	0.982
78-1, 8-9	689.6	115.0	15.91	2.23	3.10	6.913	8.43	0.583
81-2, 50-51	712.2	115.0	12.08	2.33	3.60	8.388	7.09	0.595
84-2, 45-46	739.0	116.0	16.73	2.18	3.60	7.956	13.55	1.291
85-1, 128-129	747.8	116.0	13.77	2.28	3.60	8.208	4.60	0.378
88-1, 6-8	775.1	117.0	22.97	2.04	3.60	7.344	9.90	0.727

Note: Asterisks indicate samples which had dried before arrival in our laboratory; for these samples the average of the porosity values of overlying and underlying samples was assigned. Plus sign indicates arbitrary depth at bottom of cored interval for Core 6 (recorded as coring 5.5 m but recovering 8.23 m of sediment). Brackets enclose values averaged and plotted as a single point. shaker for 2 to 4 hours; it was then removed, and the suspension was rinsed into a centrifuge tube. The suspension was centrifuged for 10 minutes at 2000 rpm, and the clear supernatant liquid was decanted. The centrifuge tube with the residual sediment was filled with candle-filtered distilled water; this was stirred into a suspension and centrifuged for 10 minutes at 2000 rpm. The clear supernatant liquid was decanted, and the procedure was repeated as necessary.

Amorphorus Fe and Mn oxides and hydroxides were removed with a sodium dithionite-sodium citrate solution buffered with sodium bicarbonate. The procedure has been detailed elsewhere (Mehra and Jackson, 1960) and is modified in part as follows: 40 ml of 0.3M sodium citrate and 5 ml of 1M sodium bicarbonate were added to the rinsed sediment in the centrifuge tube following the carbonate removal. The centrifuge tube was then placed in a 75 to 80°C water bath, and the solution was allowed to come to temperature. One gram of sodium dithionite was added, and the suspension was stirred continuously for 1 minute, then occasionally for 15 minutes; 40 ml of a hot solution of equal parts water and 0.3M sodium citrate were stirred into the suspension, and the centrifuge tube was removed from the water bath and centrifuged for 10 minutes at 2000 rpm. The clear supernatant liquid was then decanted, and 100 ml of full strength 0.3M sodium citrate was added to the centrifuge tube, stirred, and centrifuged as described above. The supernatant liquid was decanted, and the sediment was rinsed, as described earlier in the carbonateremoval section.

The sample was then sieved at  $63 \,\mu\text{m}$  to remove larger diatoms and radiolarians, and the remaining opal was removed by repeated treatments with sodium carbonate. A 100-ml centrifuge tube containing the sample was filled to three-fourths with 0.4N sodium carbonate and placed in a boiling water bath for 20 minutes. The suspension was cooled and centrifuged at 2000 rpm for 10 minutes; the supernatant liquid was decanted, and the sediment was freeze dried and weighed to determine the weight percent of the non-authigenic, inorganic, crystalline (eolian) component (Table 1).

Examination of the dried residue by scanning electron microscopy indicated that it contained no biogenic carbonate or opal, and no authigenic zeolite. Any authigenic clays would presumably remain in the residue and be included in our totals for the eolian component. The effect of not including zeolites, which have a preponderantly volcanic origin, and including smectites (montmorillonite), which result almost entirely from the diagenetic alteration of volcanic debris (Elderfield, 1976), in our totals is not clear. However, this does not appear to be a major problem in these sediments, as zeolites are much less abundant than the clay minerals (Vallier and Jefferson, this volume).

The precision of these chemical techniques was determined by replicate analyses and is commonly within 2 or 3%, a small number of replicate analyses being only within 10% of each other. The accuracy of our final values for the mass-accumulation rates of the eolian component is determined largely by the accuracy of the biostratigraphy. Placement of zonal boundaries, assuming a constant LSR between boundaries, and assignment of absolute ages to those boundaries almost certainly result in errors larger than the few percent that are the result of our laboratory work. We hesitate, therefore, to attach geological significance to changes in MAR of less than a few tens of percent of the values.

# MASS-ACCUMULATION RATES OF THE EOLIAN COMPONENT

Atmospherically transported materials include continental dust and volcanic debris. The accumulation rate of the eolian component varies with distance from source (Parkin, 1974), among other things. Although volcanic contributions are sporadic compared to the relatively steady eolian influx from continents, the proximity of various volcanic edifices to the study site suggests that the continental influx probably was masked during periods of volcanic activity.

In light of this and the mineralogical data on the nonbiogenic sediment component at Site 463 (Vallier and Jefferson, this volume), the following discussion will focus on climatic and volcanic episodes in the interpretation of the fluctuating mass-accumulation rates seen in Figure 3.

The largest average MARs (0.93 g/cm<sup>-2</sup>/10<sup>-3</sup> yr: see Table 1) of the inorganic, crystalline component, for sediments 109 to 117 m.v. old, are approximately higher by a factor of 3 than those of younger sediments at Site 463. Initial shipboard descriptions of the cores indicated a volcanic-ash layer approximately 112 to 113 m.y. old. This ash and the other inorganic materials, which accumulated rapidly from 117 to 109 m.y. ago (Fig. 3) may reflect the waning stages of the Barremian to Aptian volcanism that resulted in the formation of the Mid-Pacific Mountains (Winterer, 1976). Basement ages of the Ontong-Java Plateau (Site 289, DSDP Leg 30; Andrews, Packham, et al., 1975), the Manihiki Plateau (Site 317, DSDP Leg 33; Schlanger, Jackson, et al., 1976), and the northern Hess Rise (Site 464, report this volume) are all Aptian to early Albian and, together with the Site 463 data, provide evidence for a major mid-Cretaceous volcanic event that apparently formed several of the central Pacific aseismic rises.

Moderately high MARs  $(0.27 \text{ g/cm}^2/10^3 \text{ yr}; \text{see Table 1})$  correspond to early Albian to early Turonian time (91–108 m.y. ago). General sediment composition and mineralogy from DSDP Sites 169, 170, and 171 (Winterer, Ewing, et al., 1973) near the Line Islands in the central Pacific (Winterer, 1976) indicate moderate midplate volcanism 85 to 105 m.y. ago. Volcanic MARs from DSDP Site 462 in the Nauru Basin also are higher than those of immediately overlying sediments for this time (91–103 m.y. ago; Rea and Thiede, in press).

The eolian MARs from middle Turonian to upper Campanian time (70.5-89 m.y. ago) record fluctuating but generally low values (0.03 g/cm<sup>2</sup>/10<sup>3</sup> yr; see Table 1). Shallow-water and reef fossils from guyots in the eastern Mid-Pacific Mountains area give an approximate age of 90 m.y. for cessation of volcanic activity (Heezen, MacGregor, et al., 1973). The decreasing eolian MARs at Site 463 and the subsequent low values continuing until 71 m.y. ago correlate well with this regional reduction in volcanic activity about 90 m.y. ago. The MAR values during the interval 70.5 to 81 m.y. ago are very low, about 0.01 g/cm<sup>2</sup>/10<sup>3</sup> yr, and may be an accurate reflection of continental eolian influx to the region.

A significant increase in the eolian accumulation rate to an average of 0.21 g/cm<sup>2</sup>/10<sup>3</sup> yr (see Table 1) occurred during the Maastrichtian (67–70 m.y. ago) and during the time corresponding to sediments immediately underlying the lower to middle Tertiary hiatus. Several DSDP sites in the central Pacific record increased volcanic activity about this same time, or perhaps somewhat earlier (Sites 165, 169, 170, and 171, Leg 17, Winterer, Ewing, et al., 1973; Sites 198 and 199, Leg 20, Heezen, MacGregor, et al., 1973; Sites 315 and 316, Leg 33, Schlanger, Jackson, et al., 1976; Site 462, Leg 61, Larson, Schlanger, et al., in press). These increases in the amount of volcanic debris during the late Campanian and Maastrichtian and the ages of volcanic rock at DSDP Site 171 (91–100 m.y.) (Winterer, Ewing, et al.,



Figure 3. Mass accumulation rates of non-authigenic inorganic crystalline, (eolian) (NICE) material at Site 463. Circled points represent averages of two or more samples.

1973) and Site 313 (72–76 m.y.) (Larson, Moberly, et al., 1975), 150 km apart, suggest that repeated mid-plate volcanism dominated the eolian influx during those times.

The late Cenozoic MARs shown in Figure 3, and at an expanded scale in Figure 4, increase by a factor of 5 from late Miocene to Quaternary time. Superimposed on this general increase are significant peaks in accumulation at 4.5 and 1.0 m.y. ago. Sites 464 and 466, on Hess Rise, also reveal accumulation peaks for this same component about 4 to 6 m.y. ago (Rea and Harrsch, this volume). Regional studies of circum-Pacific (Kennett et al., 1977) and mid-plate (Rea and Scheidegger, 1979) volcanism have shown a maximum in volcanic activity between 3 and 6 m.y. ago; the lower Pliocene maximum in the non-authigenic, inorganic, crystalline component MAR appears to record this same ocean-wide phenomenon.

Although increased volcanic activity may be one cause of increasing inorganic crystalline MARs during the Quaternary (Fig. 4), the onset of Northern Hemisphere glaciation and concomitant climate deterioration may also be an important factor. General models for atmospheric circulation (Gates, 1976; Manabe and Hahn, 1977) predict that climate changes during glacial periods would result in significant increases in the intensity of atmospheric circulation. The accumulation rate and grain size of eolian sediment increase with wind intensity (Parkin, 1974). Data from a North Pacific long core taken about 1000 km north of Hawaii (LL 44-GPC3; Leinen, et al., 1979) show a sharp increase in the



Figure 4. Details of late Cenozoic MAR of the eolian fraction at Site 463. Circled Points represent averages of two or more samples.

accumulation of quartz, Al, and Si, associated with the onset of Northern Hemisphere glaciation, when the accumulation of eolian material increased significantly in the Northern Hemisphere (Windom, 1975). Thus, the intensification of atmospheric circulation resulting from deteriorating climate may account for increasing eolian MARs over the past 10 m.y., along with the periodic volcanism discussed by Kennett et al. (1977).

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