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Dust emission from Chinese desert sources linked to variations in atmospheric circulation

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Abstract. Estimates of atmospheric dust deposition to five Asian/Pacific regions indicate that ~800 Tg of Chinese desert dust is injected into the atmosphere annually; about 30% of this is redeposited onto the deserts, 20% is transported over regional scales, primarily within continental China. The remaining 50% of the dust is subject to long-range transport to the Pacific Ocean and beyond. Elemental tracers based on several dust-derived elements (Al, Fe, Mg, and Sc) reveal high-frequency variability in the contributions of the western desert sources versus northern high-dust and low-dust desert sources to eolian deposits from the center of the Loess Plateau. Comparisons of the patterns uncovered with climate signals from the remote North Atlantic region for the last glaciation show that shifts in source areas of Asian dust were synchronous with large-scale variations in atmospheric circulation.

1. Introduction

The major sources for Asian dust lie in deserts of northern and northwestern China [*Zhang*, 1984; *Merrill et al.*, 1989, 1994; *Zhang et al.*, 1996a, b] where high winds mobilize tremendous quantities of Chinese desert dust into the atmosphere. Much of the raised dust is redeposited onto the deserts (region A, see Figure 1), but a substantial amount is transported over the Loess Plateau, especially along the valley of the Yellow River, that is, region B in Figure 1 [*Liu et al.*, 1985]. Historically, "dust showers" have been recorded over extensive areas of northeastern and southeastern China, that is, regions C and D in Figure 1 [*Zhang*, 1984]. Westerly winds also carry the dust thousands of kilometers out over the Pacific Ocean, that is, region E in Figure 1 [*Duce et al.*, 1980; *Shaw*, 1980; *GESAMP*, 1989; *Merrill et al.*, 1989; *Prospero et al.*, 1989].

Despite a growing awareness of the geochemical and atmospheric importance of Chinese desert dust, little information is available on the quantity of dust produced or the distribution of the source regions. Although dust pulses are evident in Chinese loess [*Liu et al.*, 1985; *An et al.*, 1991; *Zhang et al.*, 1993, 1994; *Porter and An*, 1995], it has not been possible to apportion the contributions among source regions or even pinpoint the source areas. In this paper, elemental concentrations (Al, Fe, Mg, and Sc) for mineral dust from 12 desert sites in China (Figure 1) were used to estimate the total amount of Chinese desert dust injected into the atmosphere and to evaluate the sources for the dust input to the Loess Plateau during the last glaciation.

2. Chemical Analyses

A total of 120 size-separated desert-aerosol samples collected during both dust storm and nondust storm periods in spring of 1994 were analyzed directly using proton-induced X ray emission (PIXE). The PIXE analyses were performed using the 2.5 MeV protons with a 50 nA beam current produced by the 2 × 1.7 MV tandem accelerator at Beijing Normal University. Using these procedures, we were able to determine the concentrations of 17 elements, but in this paper we only consider the data for Al, Fe, Mg, and Sc. The concentrations of these four elements also were determined by PIXE in eight aliquots of a standard reference material from National Bureau of Chemical Exploration Analysis, China [*GSS*, 1984]. The results showed that the precision (<10%) and accuracy (<20%) were satisfactory.

3. Results and Discussion

3.1. Atmospheric Emission of Chinese Desert Dust

As a first approximation, all of the Chinese desert dust produced is assumed to deposit in regions A, B, C, D, and E (i.e., deposition equal to production, Figure 1). The total dust deposited is calculated as the sum of the dry plus the wet deposition averaged over each region (Tables 1 and 2). The dry deposition of dust (F_d) was parameterized as the product of a dry deposition velocity (V_d) and the concentration of Al in air (C_{Al}) which is about 8% of the mineral dust by weight [*Taylor*, 1964].

$$F_d = V_d \times C_{Al}/0.08 \tag{1}$$

Values for V_d used in (1) were calculated by fitting a lognormal distribution to the mass-particle size distribution of Al obtained with single-orifice, eight-stage, Battelle-type cascade impactors (PIXE International Corporation, Tallahassee, Florida) and calculating deposition rates for 100 particle-size intervals using a two-layer deposition model [*Slinn and Slinn*, 1981; *Arimoto et al.*, 1985; *Dulac et al.*, 1989; *Zhang et al.*, 1993]. Not included in the flux calculations was the wet deposition term (F_w) for region A because of the extremely low precipitation in this hyperarid region [*Liu et al.*, 1985]. Methods for estimating the wet deposition fluxes for regions B, C, D, and E have been described elsewhere [*Zhang et al.*, 1993; *Gao*, 1994; *Hashimoto et al.*, 1994; *GESAMP*, 1989], respectively.

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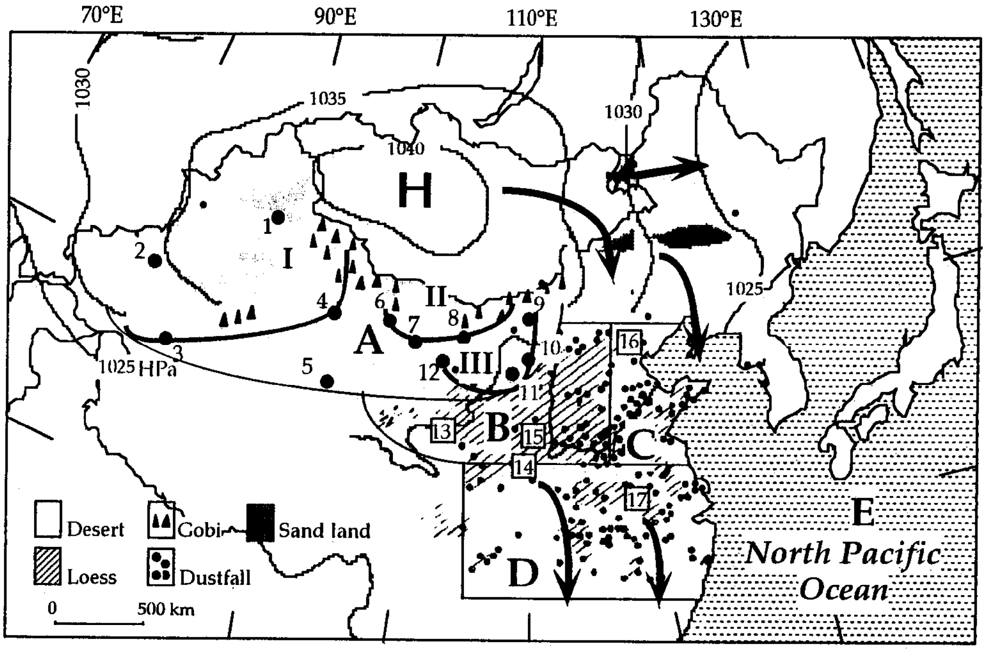

Figure 1. Map showing the winter monsoon regimes of eastern Asia [*Zhang and Lin*, 1992], aerosol sampling locations (1–12) in Chinese deserts, the sites (open squares with numbers) for the data cited (13–17), source regions (sources I, II, and III), and depositional regions (regions A, B, C, D, and E) for Asian dust. The prevailing northwesterly winds (arrows) associated with the Siberian High (HPa) and westerly winds from central Asia entrain the bulk of the Chinese desert dust delivered to the depositional areas: 1, Fukang (44°17′N, 88°7′E); 2, Aksu (41°22′N, 80°43′E); 3, Qira (37°6′N, 82°34′E); 4, Dunhuang (40°16′N, 94°10′E); 5, Golmud (36°52′N, 95°54′E); 6, Jiayuguan (40°38′N, 98°31′E); 7, Heiquan (40°26′N, 100°16′E); 8, Jartai (40°34′N, 106°34′E); 9, Dalad Qi (40°53′N, 110°5′E); 10, Yulin (38°37′N, 109°46′E); 11, Dingbian (37°37′N, 107°34′E); 12, Minqin (39°17′N, 103°10′E); 13, Lanzhou; 14, Xian; 15, Luochuan; 16, Beijing; and 17, Hefei. Chinese deserts with representative sites in parentheses: Gurbantunggut Desert (1), Taklimakan Desert (2, 3), Kumutage Desert (4), Desert in the Tsaidam Basin (5), Badain Juran Desert (6, 7), Ulan Buh Desert (8), Hobq Desert (9), Mu Us Desert (10, 11), and Tengger Desert (12). Source regions: I, western deserts (2, 3, 4); II, northern high-dust deserts (6, 7, 8); and III, northern low-dust deserts (9–12). Depositional regions: A, Chinese desert regions (1–12), excluding three sandy lands in northeastern China [*Zhang*, 1984]; B, Chinese Loess Plateau (13, 14) [*Liu et al.*, 1985]); C, historical northeastern dustfall region, 34.3°–41°N, to the east of 114°E ((16) [*Zhang*, 1984]); D, historical southeastern dustfall region, 27.3°–34.3°N, to the east of 104.7°E ((17) [*Zhang*, 1984]); E, North Pacific Ocean [*GESAMP*, 1989].

Annual fluxes were estimated by assuming that the springtime fluxes represented 34% of annual dust activity in northwestern China [*Watts*, 1969]. We used a corresponding value of 13% for southeastern China where dust storms are infrequent [*Gao*, 1994]. Monthly depositional fluxes during the spring were estimated by combining dry deposition and wet deposition. As a means of evaluating the representativeness of the 1994 data, we compared the daily mean dry deposition fluxes for samples collected from Minqin (MQ) in the spring of 1994 versus samples from Shapuotou (SPT) collected in April 1990 [*Zhang et al.*, 1993]. Both the MQ and SPT sites are located along the southern margin of the Tengger Desert. The daily mean fluxes of Al at MQ and SPT were essentially the same (MQ: 88 mg m^{-2} d^{-1}; SPT: 85 mg m^{-2} d^{-1}), even though the samples were collected in different years; this demonstrates that our results for the spring of 1994 are not extraordinary.

The annual flux to the Taklimakan Desert was obtained by averaging the yearly data for two representative sites, that is, Qira and Aksu, and adding the contributions from dust storms observed at the two sites (Table 1). The annual fluxes for the other regions were obtained in the same way, except at Xi'an where we have a complete set of yearly data [*Zhang et al.*, 1993]. The fluxes in nondust storm (NDS) and dust storm (DS) conditions were calculated from (2) and (3), respectively. Of the 120 sets of aerosol samples collected in the desert regions, only nine were collected under DS conditions. Since we do not have dry deposition velocity data for Al at Hefei, we averaged the V_d (3.2 cm s^{-1}) calculated for Xi'an [*Zhang et al.*, 1993] with that for Xiamen [1.4 cm s^{-1}, *Gao*, 1994] to obtain a value (2.3 cm s^{-1}) which should be reasonably applicable to the Hefei region. This average V_d is comparable to a value of 2 cm s^{-1} which has been widely accepted as a representative value for nearshore areas [*Uematsu et al.*, 1985; *Duce et al.*, 1991]. The Al concentrations at Hefei were from *Hu et al.* [1990].

As an example, consider the mineral dust flux calculations for the Taklimakan Desert (Table 1). The concentration of Al (C_{Al}) in one sample from Qira was 28 μg m^{-3}, and the dry deposition velocity (V_d) calculated for one of the 29 NDS days of April was 4.8 cm s^{-1}. On the basis of these values the annual dry deposition flux (F_d) of dust during the NDS periods, assuming 34% of the flux occurs in spring, would be

Table 1. Total Atmospheric Deposition of Mineral Aerosol to Chinese Deserts

Chinese Deserts (Sampled From 1991 to 1994)	Representative Sites	n[a]	Total Deposition,[b] g m^{-2} yr^{-1}: Individual Site	Total Deposition,[b] g m^{-2} yr^{-1}: Region	Deposition,[b] Tg yr^{-1}
Taklimakan Desert (337,600 km^2)[c]				450 (110–1900)	150 (37–630)
	Qira	9	290 (47–1000)		
	Qira DS[d]	4	77 (31–170)		
	Aksu	9	520 (94–2300)		
	Aksu DS	3	12 (7.7–17)		
Gurbantunggut Desert (48,800 km^2)				130 (37–270)	6.2 (1.8–13)
	Fukang	15	130 (37–270)		
Desert in the Tsaidam Basin (34,900 km^2)				230 (68–480)	7.9 (2.4–17)
	Golmud	15	230 (68–480)		
Kumutage Desert (19,500 km^2)				320 (40–1100)	6.2 (0.8–22)
	Dunhuang	7	320 (40–1100)		
Badain Juran Desert (44,300 km^2)				310 (99–750)	14 (4.4–33)
	Jiayuguan	6	380 (120–760)		
	Jiayuguan DS	1	4.3		
	Heiquan	7	240 (78–740)		
Ulan Buh Desert (9,970 km^2)				670 (14–2100)	6.7 (0.1–21)
	Jartai	6	630 (14–2100)		
	Jartai DS	1	37		
Hobq Desert (16,100 km^2)				420 (73–570)	6.7 (1.2–9.2)
	Dalad Qi	8	420 (73–570)		
Mu Us Desert (32,100 km^2)				380 (66–1300)	12 (2.1–42)
	Yulin	10	390 (95–1700)		
	Dingbian	10	370 (37–860)		
Tengger Desert (42,700 km^2)				290 (15–1200)	12 (0.6–52)
	Minqin	9	290 (15–1200)		

[a]Number of aerosol samples.
[b]Mean value with the range in parentheses.
[c]Institutes of Glaciology, Geocryology, and Desert, Academia Sinica.
[d]DS, dust storm.

$$F_{d(\mathrm{NDS})} = (V_d \times C_{\mathrm{Al}}/0.08)(2.51 \times 10^6 \text{ s month}^{-1})(3 \text{ months})/34\%$$
$$= (4.8 \text{ cm s}^{-1})(28\ \mu\text{g m}^{-3})(2.77 \times 10^8)$$
$$= 370 \text{ g m}^{-2} \text{ yr}^{-1} \qquad (2)$$

During the DS period (2 days during the sampling period) the Al concentration (C_{Al}) in one sample was 41 μg m^{-3}, and the calculated V_d was 68 cm s^{-1}. Thus the dry deposition flux (F_d) of dust under DS conditions based on those values would be

$$F_{d(\mathrm{DS})} = V_d \times C_{\mathrm{Al}}(8.64 \times 10^4 \text{ s d}^{-1})(2 \text{ d yr}^{-1})/0.08$$
$$= (68 \text{ cm s}^{-1})(41\ \mu\text{g m}^{-3})(2.16 \times 10^6)$$
$$= 60 \text{ g m}^{-2} \text{ yr}^{-1} \qquad (3)$$

In Table 1 the mean flux (290 g m^{-2} yr^{-1}) at Qira during NDS periods was calculated by averaging the nine individual

Table 2. Atmospheric Deposition of Mineral Aerosol to Five Regions

Region	Description	Representative Sites	n[a]	Total Flux,[b] g m^{-2} yr^{-1}: Individual Site	Total Flux,[b] g m^{-2} yr^{-1}: Region	Deposition,[b] Tg yr^{-1}
A	Chinese Deserts (585,970 km^2)	see Table 1				220 (50–840)
B	Chinese Loess Plateau (291,600 km^2)	Xian[c]	15	170 (10–600)	250 (10–600)	73 (3.0–180)
		Xian DS	2	3.9 (3.2–4.7)		
		Lanzhou[d]	1	330		
C	Historical NE dustfall region (360,000 km^2)	Beijing[e]	3	94 (78–119)	95 (78–120)	34 (28–43)
		Beijing DS	1	1.0		
D	Historical SE dustfall region (1,250,000 km^2)	Hefei[f]	17	27 (14–89)	27 (14–89)	34 (18–110)
E	North Pacific[g]					420 (400–500)
				Total deposition (equal to emission) ≅ 800 (500–1100) Tg yr^{-1}		

[a]Number of aerosol samples.
[b]Mean value with the range in parentheses.
[c]Wet deposition was 7.1% of the yearly data [*Zhang et al.*, 1993].
[d]Wet deposition was 1.6% of the spring flux [*Hashimoto et al.*, 1994].
[e]Dry deposition data were from Zhang (1993), and wet deposition was 2.2% of the spring data [*Gao*, 1994].
[f]Wet deposition was 24% of the spring data [*Gao*, 1994].
[g]Data from *GESAMP* [1989] and adjusted for non-Chinese desert sources [*Zhang et al.*, 1996a].

Table 3. Concentrations of Atmospheric Aluminum and Elemental Signatures for Chinese Desert Regions

Individual Desert Site	n[a]	Al, $\mu g\ m^{-3}$	Fe/Al	Mg/Al	Sc/Al, (Modal Value) $\times 10^{-3}$
Aksu	12	29	0.84	0.25	0.26
Qira	13	36	1.2	0.16	0.29
Dunhuang	7	20	0.57	0.16	0.25
Jiayuguan	7	18	0.65	0.30	0.34
Jartai	7	22	0.60	0.32	0.35
Heiquan	7	18	0.70	0.31	0.32
Minqin	9	16	0.35	0.22	0.35
Dingbian	10	30	0.78	0.20	0.31
Yulin	10	24	0.25	0.17	0.37
Dalad Qi	8	17	0.53	0.23	0.34
Desert Sources[b]	n[a]	Al, $\mu g\ m^{-3}$	Fe/Al	Mg/Al	Sc/Al, (Geometric Mean) $\times 10^{-3}$
Source I	32	28	0.83	0.19	0.27
Source II	21	19	0.65	0.31	0.34
Source III	37	21	0.44	0.20	0.34

[a]Number of samples.
[b]Source I, Aksu, Qira, and Dunhuang; source II, Jiayuguan, Jartai, and Heiquan; source III, Minqin, Dingbian, Yulin, and Dalad Qi.

NDS fluxes, one of which is shown in detail above. In the same way the mean flux for DS conditions (77 g $m^{-2}\ yr^{-1}$) was the average of the fluxes for the four DS events, assuming each lasted for 2 days. At Aksu the corresponding NDS and DS values were 520 and 12 g $m^{-2}\ yr^{-1}$, respectively.

Assuming negligible input by wet deposition, the total dust flux (D_T) to the Taklimakan Desert (with the area of 337,600 km^2, Table 1) is given by

$$D_T = F_{d(\mathrm{NDS})}$$

$$+\ F_{d(\mathrm{DS})}\ \text{(averaged using data from Qira and Aksu)}$$

$$= [(290 + 520)/2 + (77 + 12)/2]\ \mathrm{g\ m^{-2}\ yr^{-1}}$$

$$\times\ 337{,}600 \times 10^6\ \mathrm{m^2} = 150\ \mathrm{Tg\ yr^{-1}}$$

The total dust production from Chinese deserts, obtained by summing the amounts of dust deposited on the five study regions, is ~800 Tg yr^{-1} with an estimated range from 500 to 1100 Tg yr^{-1} (Table 2); this amounts to about half of the global production of dust which is variously estimated to be ~1500 Tg yr^{-1} [*Andreae*, 1995], or 1000–2000 Tg yr^{-1} [*Duce*, 1995].

For region A the deposition (220 Tg yr^{-1}) is ~30% of the total emissions from the Chinese deserts (Table 2). The estimated dust deposition per unit area in some desert regions, for example, the Gurbantunggut and the desert in the Tsaidam Basin, actually were lower than that estimated for Lanzhou, which borders the western margin of the Loess Plateau. The Loess Plateau (region B) receives far more desert dust per unit area (250 g $m^{-2}\ yr^{-1}$) than the continental regions C and D (Table 2). The total dust input to continental China (the sum of regions B, C, and D) is equivalent to ~20% of the total desert production. About half of the dust produced is transported to the remote North Pacific, region E [*GESAMP*, 1989; *Prospero et al.*, 1989]. However, ~5% of the material deposited is thought to be from non-Chinese desert sources [*Zhang et al.*, 1996a].

3.2. Elemental Tracers and Source Distributions of Asian Dust

No single elemental tracer is likely to be unique to a specific desert source, but some variations in the proportions of elements could occur if different combinations of minerals are present in the desert source areas [*Zhang et al.*, 1996b]. Weighted multiple-linear regression techniques are useful for estimating the contributions of various sources to a given sample [*Rahn and Douglas*, 1984], and therefore a chemical element balance (CEB) model [*Williamson and DuBose*, 1983; *Lowenthal et al.*, 1988] was used to evaluate the source distributions for Asian dust. The signatures were constructed from three elemental ratios (Fe, Mg, Sc) normalized to Al [*Zhang et al.*, 1996b]. Ratios were used to account for any processes, such as changes in source strengths or the dispersal or the removal of dust, that would affect the concentrations but not the composition of dust. The elemental data from the two low-deposition desert regions mentioned above were not used in the development of this source tracer system [*Zhang et al.*, 1996b].

The signatures for three putative dust sources have been shown to differ significantly (Fisher's Protected Least Significant Difference test [*Zhang et al.*, 1996b]). The signatures for Qira, Aksu, and Dunhuang were grouped together as a cluster, and these sites were considered to be representative of the western deserts, referred to hereinafter as source I. Similarly, Jiayuguan, Heiquan, and Jartai were combined into a northern high-dust desert signature, source II; and Dalad Qi, Yulin, Dingbian, and Minqin defined a northern low-dust desert signature, source III (Figure 1 and Table 3). To compensate for the effects of transport among the desert sites, the signatures were based on modal values.

One would assume that coarse dust particles would deposit faster than small particles, and this could lead to a fractionation of the elements. However, we found that the elemental ratios remained stable at least when the dust particles were transported directly from the source regions to the areas in which loess accumulates. This conclusion is based on cascade impactor data collected during a dust storm event that was first sampled at a source site in the Tengger Desert (Shapuotou), and then 400 km downwind at Xi'an, on the southern margin of the Loess Plateau [*Zhang et al.*, 1991, 1993]. The elemental ratios remained recognizable, at least until the majority of the particles larger than 2 μm in diameter were removed during transport [*Zhang et al.*, 1996b]. This suggests that the elemental signatures are fairly stable and provide a means for tracing the long-range transport of Chinese desert dust. Furthermore, using the signatures for the three presumptive sources (Table 3) together with the dustfall data from four dust storm events, the estimated contributions of each source calculated from CEBs were shown to be generally consistent with the meteorological analyses for the origins of the dust storm events [*Zhang and Wang*, 1995]. This further attests to the validity of the approach [*Zhang et al.*, 1996b].

3.3. Past Atmospheric Circulation Inferred From the Elemental Tracers

Since the main structures of Chinese deserts formed before the late Pleistocene [*Dong et al.*, 1990, 1991, 1995], one can assume that the source signatures have changed little since the

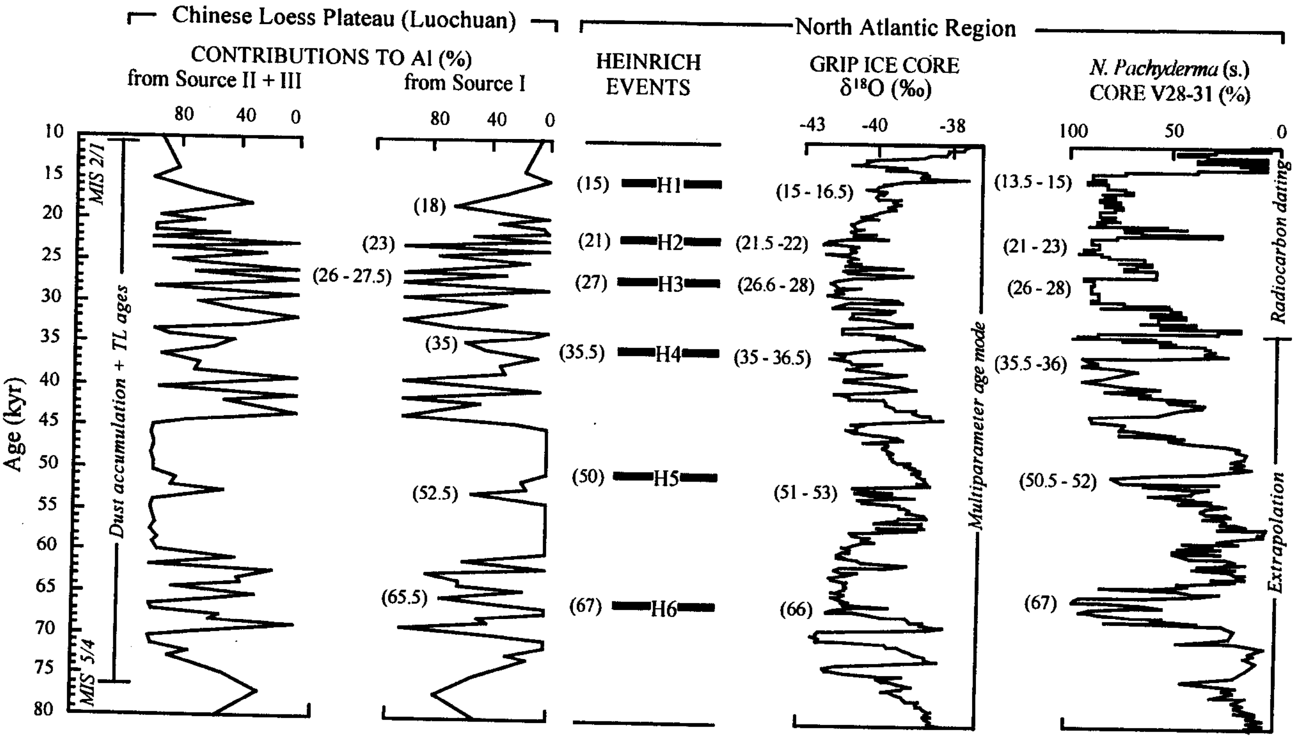

Figure 2. Time series comparing variations in the contribution of desert sources to Al concentrations in the Luochuan loess/paleosol succession with Greenland ice core $\delta^{18}O$ variations [*Dansgaard*, 1993], North Atlantic Heinrich events [*Heinrich*, 1988], and *N. pachyderma* (s) variations [*Bond et al.*, 1993] during the last glaciation (80–10 kyr). Loess stratigraphic unit [*Zhang et al.*, 1994] was sampled at 10-cm intervals; elemental concentrations were from *Zhang et al.* [1994]. The loess time series was dated by a sedimentation model [*Porter and An*, 1995]. Ages of Heinrich events H1–H3 were based on accelerator mass spectrometry (AMS) date bracketing layers of ice-rafted detritus [*Bond et al.*, 1992]; the age of H4 was based on one AMS age [*Bond et al.*, 1993], whereas ages of H5 and H6 were obtained by extrapolation [*Bond et al.*, 1993]. The upper part of the *N. pachyderma* (s) record was controlled by AMS ages, but ages of the lower part were based on extrapolation of ^{14}C-based sedimentation rates [*Bond et al.*, 1992]. The chronology of the upper 8.6 kyr of the GRIP core was controlled by annual layer counting and volcanic-acid reference horizons, whereas the lower part was based on a multiparameter method [*Johnsen et al.*, 1992]; estimated error limits ranged from 0.05 kyr at 8.6 kyr to 2 kyr at 40 kyr. Ages of peaks (in parentheses) are rounded to the nearest (0.5 kyr) in all time.

last glaciation. Accordingly, the dust contributions to the well-known Luochuan loess/paleosol sequence [*Liu et al.*, 1985; *An et al.*, 1991; *Zhang et al.*, 1994] can be apportioned between source I versus sources II and III using the CEB model. Each tracer element showed similar variations, and therefore only the curves for Al showing the source apportionments vs. age are presented in Figure 2.

The contribution of source I exhibits high-frequency variability that is synchronous with paleoclimatic proxy records from the North Atlantic dating back to the last glaciation (Figure 2). Maxima in the source I contributions are generally coherent with minima in $\delta^{18}O$ of Greenland Ice-Core Project (GRIP) (Greenland) ice cores, that is, Dansgaard-Oeschger cycles [*Dansgaard et al.*, 1993]; maxima in the percentages of *Neogloboquadrina pachyderma* (s), that is, Bond cycles [*Bond et al.*, 1993]; and Heinrich layers (H1–H6) in a core in North Atlantic [*Heinrich*, 1988]. Therefore, the increased dust exports from the western desert sources evidently coincided with the cooling of Atlantic surface waters [*Bond et al.*, 1993; *Heinrich*, 1988] and atmospheric cooling over Greenland [*Dansgaard et al.*, 1993]. These cooling episodes are thought to have effected sites outside the North Atlantic region [*Broecker*, 1994], including regions directly downwind in Europe [*Rind et al.*, 1986] and China [*Porter and An*, 1995]. Furthermore, it is noteworthy that cooling episodes in the northern hemisphere have been linked with strengthening of the westerlies (X. D. Liu, Institute of Plateau Atmospheric Physics Research, Chinese Academy of Sciences, Lanzhou, People's Republic of China, personal communication, 1997).

The regional-scale transport of Chinese desert dust is dominated by the surface-level winds of the winter monsoon [*An et al.*, 1990; *Zhang and An*, 1996c]. Surges of cold air associated with the monsoonal circulation historically increased in strength and frequency from the west as the westerlies intensified and latitudinal circulation developed [*Zhang and Wang*, 1995]. Similarly, the increased dust inputs from source I observed in Chinese loess can be attributable to shifts in the western fraction of the cold surge. Conversely, the periods during which the dust from sources II and III became increasingly important correspond to the transport of cold air from the north and the associated breakdown of the westerlies. These shifts in Asian dust sources apparently were synchronous with climatic events in the North Atlantic region, a strong indication that changes in the dust fluxes from the Asian source regions were linked with large-scale variations in atmospheric circulation.

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