Earth-Science Reviews 54 (2001) 81–114
www.elsevier.com/locate/earscirev

DIRTMAP: the geological record of dust

Karen E. Kohfeld *, Sandy P. Harrison

Max Planck Institute for Biogeochemistry, Postfach 10 01 64, D-07701 Jena, Germany

Received 19 April 2000; accepted 15 December 2000

Abstract

Atmospheric dust is an important feedback in the climate system, potentially affecting the radiative balance and chemical composition of the atmosphere and providing nutrients to terrestrial and marine ecosystems. Yet the potential impact of dust on the climate system, both in the anthropogenically disturbed future and the naturally varying past, remains to be quantified. The geologic record of dust provides the opportunity to test earth system models designed to simulate dust. Records of dust can be obtained from ice cores, marine sediments, and terrestrial (loess) deposits. Although rarely unequivocal, these records document a variety of processes (source, transport and deposition) in the dust cycle, stored in each archive as changes in clay mineralogy, isotopes, grain size, and concentration of terrigenous materials. Although the extraction of information from each type of archive is slightly different, the basic controls on these dust indicators are the same. Changes in the dust flux and particle size might be controlled by a combination of (a) source area extent, (b) dust emission efficiency (wind speed) and atmospheric transport, (c) atmospheric residence time of dust, and/or (d) relative contributions of dry settling and rainout of dust. Similarly, changes in mineralogy reflect (a) source area mineralogy and weathering and (b) shifts in atmospheric transport. The combination of these geological data with process-based, forward-modelling schemes in global earth system models provides an excellent means of achieving a comprehensive picture of the global pattern of dust accumulation rates, their controlling mechanisms, and how those mechanisms may vary regionally. The Dust Indicators and Records of Terrestrial and MArine Palaeoenvironments (DIRTMAP) data base has been established to provide a global palaeoenvironmental data set that can be used to validate earth system model simulations of the dust cycle over the past 150,000 years.

Keywords: Dust; Palaeoclimate; Quaternary; Accumulation rates

1. Introduction: why is dust important?

The concentration of dust in the atmosphere (i.e. the atmospheric dust loading) influences the climate system through affecting radiative forcing, through chemical reactions with other atmospheric constituents, and through acting as a source of nutrients to biological systems (Rahn et al., 1979; Swap et al., 1992; Duce, 1995; Lacis and Mishenko, 1995; Dentener et al., 1996; Li-Jones and Prospero, 1998; Zhang and Carmichael, 1999; Harrison et al., in press). Although the role of dust in the climate system is poorly understood in quantitative terms, it is clear that changes in atmospheric dust loading could potentially have a significant impact on future climate changes (Andreae, 1995; Tegen and Fung, 1995; Tegen et al., 1996; Shine and Foster, 1999).

* Corresponding author. Fax: +49-3641-643-775.
E-mail address: kek@bgc-jena.mpg.de (K.E. Kohfeld).

0012-8252/01/$ - see front matter
PII: S0012-8252(01)00042-3

Thus, the role of dust in climate change has become a major focus of earth system modelling research (see e.g. Sokolik and Toon, 1996, 1999; Tegen et al., 1996; Tegen and Lacis, 1996; Andersen et al., 1998; Miller and Tegen, 1998, 1999; Mahowald et al., 1999; Reader et al., 1999).

Because dust is highly variable in space and time, and quantitative data on global dust distributions are not available, the climate impact of dust aerosols is usually estimated using dust distributions derived from global transport models. Simulations of the magnitude and spatial patterns of the atmospheric dust loading under modern climate conditions are evaluated using satellite data (specifically measurements of short-wave spectral reflectance, e.g. King et al., 1999). Simulated dust deposition rates can be similarly evaluated using modern surface sediment trap data (e.g. Ratmeyer et al., 1999). Despite a number of problems in the interpretation of these data (see e.g. Tegen and Miller, 1998), comparisons of simulations of the modern dust cycle show that atmospheric tracer transport models coupled with atmospheric general circulation models are capable of reproducing the first-order patterns of dust transport and deposition under modern climate conditions reasonably well (Wefers and Jaenicke, 1990; Genthon, 1992; Tegen et al., 1996; Mahowald et al., 1999). However, evaluation under modern climate conditions is not sufficient to show that the current generation of dust-cycle models is capable of correctly simulating the impact of changing dust loadings on future climates. Comparisons of the climate simulated by a range of different atmospheric general circulation models (AGCMs) under radically different conditions in the past, made as part of the Palaeoclimate Modelling Intercomparison Project (PMIP; Joussaume and Taylor, 1995), with palaeoclimate observations show that the models that are best at reproducing the modern climate are not necessarily the best at reproducing changed climates (Harrison et al., 1998; Joussaume et al., 1999; Kageyama et al., 1999; Pinot et al., 1999). The concept that the performance of earth system models under both modern and palaeoclimate conditions should be compared with observational evidence is now widely accepted as fundamental to the strategy for evaluation of earth system models (Kutzbach and Webb, 1993; Kohfeld and Harrison, 2000).

The past provides many opportunities to evaluate dust-cycle models under radically different conditions from today, and global Quaternary records of dust suggest that different climatic periods are accompanied by a large variation in dust fluxes (Fig. 1). Ice core records from Greenland and Antarctica, spanning several climatic cycles, show that aeolian

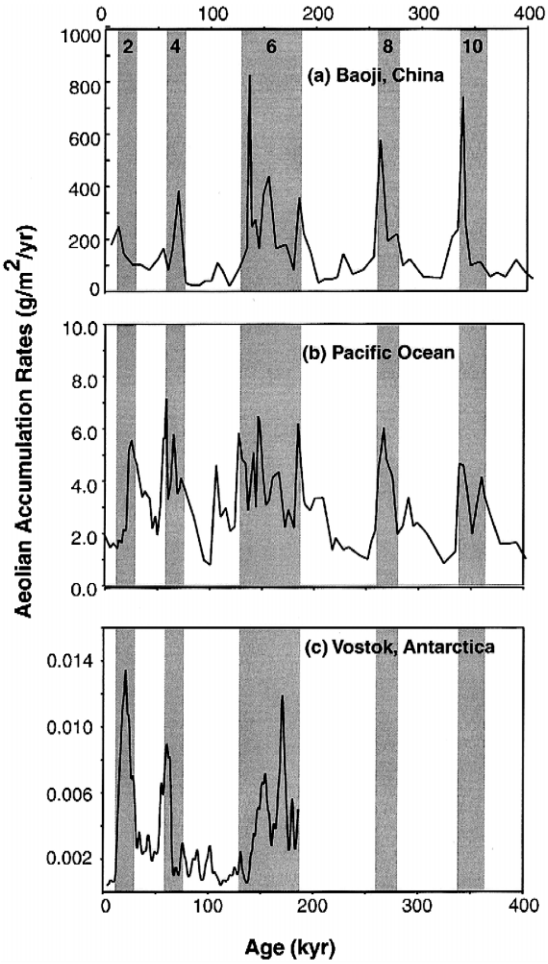


Fig. 1. Quaternary changes in aeolian accumulation rates from (a) the Chinese Loess Plateau (Ding et al., 1994), (b) a marine sediment record of V21-146 in the Pacific Ocean (Hovan et al., 1991), and (c) the Vostok ice core (Petit et al., 1990). Grey shaded regions represent the cold, glacial Marine Isotope Stages. Although records from all three sources show a similar pattern, with increased dust loadings during glacial stages, the absolute magnitudes of dust accumulation decrease by more than an order of magnitude between terrestrial and marine, and marine and polar ice core sites.

deposition rates at high latitudes were as much as 20 times greater during glacial than interglacial periods (Petit et al., 1981, 1990, 1999; Hammer et al., 1985; Taylor et al., 1993; Steffensen, 1997). Marine and terrestrial records from the last glacial maximum (LGM, ca. 21,000 calendar year B.P.) imply that the increase in atmospheric dust loading was not spatially uniform or ubiquitous, but increased dust deposition rates at the LGM of up to 10 times more than present downwind of major source areas resulted in a much dustier state globally. On shorter time scales, the ice core records suggest that variations in the atmospheric burden of dust may have been even larger than the glacial–interglacial changes (Steffensen, 1997; Fuhrer et al., 1999). For example, the transition between the Bölling/Alleröd and Younger Dryas periods (ca. 13,000 calendar year B.P.) was marked by a rapid shift in dust deposition with significant increases at both poles. Dust concentrations in the Greenland ice core within the cold stages show short-term (annual to decadal) fluctuations that are up to two orders of magnitude greater than levels recorded in Holocene ice (Steffensen, 1997; Fuhrer et al., 1999). The cause and implications of these rapid variations in atmospheric dust loading are still not known, but can be investigated using models.

In order to use geologic data for the evaluation of earth system models, including dust cycle models, the data must be made available to the modelling community in a useful form. Ideally, the geologic data should be available as a spatially extensive data set, where the data at individual sites are expressed in a form that is directly comparable with the output(s) of the model (Kohfeld and Harrison, 2000). There is a number of global palaeoclimate data bases available for the evaluation of various aspects of earth system models (e.g. BIOME 6000, Prentice and Webb, 1998; the Global Lake Status Data Base, Qin et al., 1998; Kohfeld and Harrison, 2000). The purpose of this paper is to present the Dust Indicators and Records of Terrestrial and Marine Paleoenvironments (DIRTMAP) data base, which has been specifically designed to provide data relevant to the palaeo-dust cycle. We begin by discussing the kinds of information provided by the different geologic archives of dust, before showing how these records are preserved and transformed in the DIRTMAP data base. We illustrate how DIRTMAP can be used (a) to document the state of the dust cycle at key times in the past and (b) for model evaluation. Finally, we conclude by considering how future improvements to DIRTMAP could be used to improve our understanding of both the record and role of dust in the past.

2. What information does the geologic record of dust provide?

Models of the dust cycle simulate the processes by which dust is generated in source regions, lifted from the surface into the atmosphere, transported by winds, and then deposited either by settling (dry deposition) or by being washed out (wet deposition) of the atmosphere (see Harrison et al., this volume). Ideally, data are required to evaluate how well the model simulates each process. Thus data are required that provide information on the location and extent of dust sources, the magnitude of dust emissions, the pathways of dust transport, and the location and magnitude of dust deposition. Although the geological archives of dust cannot provide all of this information, they do provide a remarkable amount of information about different aspects of the dust cycle. Here, we describe (a) the different ways in which different processes are documented within the geological record and (b) how information can be extracted from each type of archive using several techniques and tracers, including mineralogy, isotopes, grain size, and concentration of terrigenous materials.

2.1. Ice cores

The continental particulate material deposited on ice sheets is derived by atmospheric transport, and is thus of purely aeolian origin. Of all the dust archives, ice cores are therefore the most straightforward to interpret and can provide useful records of dust deposition from regions that are currently ice-covered (Table 1). The total amount of dust present in the core provides a record of dust deposition rates through time. The mineralogy, chemistry and isotopic composition of the particulate matter can be used to identify the source areas from which the aeolian material is derived or to reconstruct the atmospheric circulation regime during deposition.

Table 1
Ice cores containing published dust measurements

Site	Latitude	Longitude	Measurement type	Time span (ka)	References
Camp Century, Greenland	77°N	61°W	C_M	0.63–112	Cragin et al. (1977)
			C_N	0–26	Thompson and Mosley-Thompson (1981)
			PS	0–26	Thompson (1977)
			GC	0.63–112	Cragin et al. (1977)
Crête, Greenland	71°N	37°W	C_N	0–0.05	Hammer (1977)
			C_M	0–0.1	Steffensen (1988)
			GC	0–0.1	Steffensen (1988)
Dye 3, Greenland	65°N	44°W	C_M	0–0.4	Hammer et al. (1985)
			C_N	0–0.4	Hammer (1977)
			PS	0–0.4	Steffensen (1995)
Renland, Greenland	71°N	27°W	$C_{M,}$	0–120	Hansson (1994)
			GC	0–120	Hansson (1994)
GRIP, Summit, Greenland	73°N	38°W	C_M	0–120	Steffensen (1997)
			C_N	snowpit	Steffensen et al. (1996)
			PS	0–120	Steffensen (1997)
			MIN	0–135	Maggi (1997)
			GC	8–100; 115–130; snowpit	Fuhrer et al. (1993, 1999) and Steffensen et al. (1996, 1997)
GISP2, Summit, Greenland	73°N	38°W	$C_{N,}$ C_M	10.5–14.0	Zielinski and Mershon (1997)
			PS	10.5–14.0	Zielinski and Mershon (1997)
			MIN	23.34–26.18	Biscaye et al. (1997)
			ISO	23.34–26.18	Biscaye et al. (1997)
			GC	0–110	Mayewski et al. (1997)
Milcent, Greenland	70°N	44°W	C_N	0–0.05	Hammer (1977)
Byrd, Antarctica	80°S	120°W	C_M	2.5–90	Cragin et al. (1977)
			C_N	0–26	Thompson and Mosley-Thompson (1981)
			GC	2.5–90	Cragin et al. (1977)
Dome C, Antarctica	75°S	124°E	C_M	0–30	Petit et al. (1981) and Royer et al. (1983)
			C_N	0–30	Petit et al. (1981), Thompson and Mosley-Thompson (1981) and Royer et al. (1983)
			PS	0–30	Thompson and Mosley-Thompson (1981)
			MIN	0–30	Gaudichet et al. (1986, 1992)
			ISO	0–30 s	Grousset et al. (1992a) and Basile et al. (1997)
Vostok, Antarctica	78°S	107°E	C_M	0–400	Petit et al. (1981, 1990, 1999) and De Angelis et al. (1984)
			MIN	0–150	Gaudichet et al. (1988, 1992)
			ISO	18, 60, 160	Basile et al. (1997)
			PS	0–50	De Angelis et al. (1984)
			GC	0–50	De Angelis et al. (1984)
Siple Station, Antarctica	76°S	84°W	C_N	0–0.55	Mosley-Thompson et al. (1990)
Penny Ice Cap, Canada	67°N	66°W	C_N, C_M, PS	0–11.55	Zdanowicz et al. (2000)
Devon Island, Canada	75°N	82°W	C_N	0–120	Fisher (1979)
Dunde, China	38°N	97°E	C_N	0–35.5	Thompson et al. (1989)
Guliya, China	35°N	81°E	C_N, GC	0–132	Thompson et al. (1997)
Huascaràn, Peru	9°S	78°W	C_N	0–25	Thompson et al. (1995)
Sajama, Bolivia	18°S	69°W	C_N, PS	0–25	Thompson et al. (1989)
Dasuopu Glacier, China	28°N	85°E	C_N, GC	0–0.02	Thompson (2000)

Types of analysis: C_M = Mass concentrations of insoluble particles; C_N = Number concentrations of insoluble particles; ISO = Radiogenic isotope fingerprinting; MIN = Clay Mineralogy; GC = Glacio-chemistry (Ca, Mg, Na, K, NH_4, SO_4, NO_3, and Cl); PS = Particle Size.

Changes in the grain size of the particulate matter can be interpreted as an indicator of changes in source area proximity, wind strength and/or changes in the type of deposition processes.

Changes in the total amount of dust (i.e. the dust concentration) in the ice cores provide the most direct measure of changes in atmospheric dust loading through time. Changes in the dust concentration are derived in two ways. In the first approach, changes in the flux of aeolian material to the ice core are estimated by combining measurements of the dust particle concentration with ice accumulation rates (e.g. Petit et al., 1981). The dust flux estimate is sensitive to the accuracy of the estimation of changes in ice accumulation rate (Mahowald et al., 1999). The second approach assumes that changes in Ca^{2+} concentration within the ice cores directly reflect the overlying atmospheric dust loading (e.g. Alley et al., 1995; Mayewski et al., 1997). However, the concentration of Ca^{2+} within the ice is affected by changes in the relative fractions of dust removed by rainout and dry settling (Alley et al., 1995), which in turn is affected by the changes in local precipitation rates. Furthermore, recent studies have suggested that the relationship between the amount of dust removed by precipitation and precipitation rate is highly variable (Davidson et al., 1996).

The mineralogy, chemistry and isotopic composition of the particulate matter can be used to identify the source areas from which the aeolian material is derived. The approach is based on characterising the mineralogical, chemical or isotopic "fingerprints" of potential source areas, and matching them to the observed mineral, chemical, and/or isotope composition of the particulate matter in the ice cores (e.g. Gaudichet et al., 1988, 1992; Grousset et al., 1992a; Fuhrer et al., 1996; Basile et al., 1997; Biscaye et al., 1997; Maggi, 1997; Steffensen et al., 1997). Since the mineral matter reaching the ice cores is extremely fine-grained (< 6 μm, Steffensen, 1997), mineralogical characterisation relies heavily on the characterisation of a suite of clay minerals, relative abundance of clay minerals (e.g. kaolinite/illite ratios), or the presence of rare earth elements (REE) that can be compared with that of potential source areas (Gaudichet et al., 1986, 1988; Biscaye et al., 1997). Chemical characterisation is generally based on measurement of the major chemical species found in glacial ice and snow (e.g. sodium, potassium, ammonium, calcium, magnesium, sulphate, nitrate, and chloride). These chemical species can be introduced to the atmosphere as primary aerosols, such as sea salt (sodium, chloride, and to a lesser extent magnesium, calcium, sulphate, and potassium) or continental dust (magnesium, calcium, carbonate, sulphate, and aluminosilicates) (e.g. Delmas and Legrand, 1989; Shaw, 1989; Whitlow et al., 1992; Legrand and Mayewski, 1997; Mayewski et al., 1997). However, they can also be secondarily introduced to the atmosphere via other sources including oxidation pathways involving several atmospheric trace gases (e.g. Legrand et al., 1993; Legrand and Mayewski, 1997), making chemical data difficult to interpret uniquely. Isotope fingerprinting has been undertaken using radiogenic isotopes of rubidium, strontium, and neodymium (Biscaye et al., 1974, 1997; Goldstein et al., 1984; Grousset et al., 1988, 1992a; Basile et al., 1997).

The mineral and chemical characteristics of source areas are generally less regionally specific than the radiogenic isotope fingerprint, but have been used, e.g. to rule out the Sahara (characterised by high smectite and high kaolinite/chlorite (K/C) ratios) as a potential source for dust reaching the Greenland ice cores (characterised by high illite, no smectite, and low K/C ratios) during the LGM (Biscaye et al., 1997). Radiogenic isotope fingerprinting has been used to identify the East Asian deserts as the most likely source of dust in the Greenland ice core both at the LGM and during the Holocene (Biscaye et al., 1997; Svensson et al., 2000). Similar studies have established that the dust recorded in the Antarctic ice cores during the LGM was probably derived from Patagonia (Grousset et al., 1992a; Basile et al., 1997). Once the source of dust reaching a specific core site is established, the dust transport trajectory, and hence, changes in atmospheric circulation regimes can be inferred. In the case of the Greenland and Antarctic records, for example, there does not appear to be any indication that the LGM high-latitude circulation regimes were different from today. The reliability of mineralogical, chemical, or isotopic analyses to identify the sources of dust in ice cores clearly requires the availability of measurements characterising all potential source areas. Such studies of the isotope composition of potential source areas have been carried out at relatively few sites (Fig. 2) and so there are still some uncertainties inherent in the attribution technique. Nevertheless, those studies that have been carried out demonstrate the importance of establishing the mineralogical, chemical and isotopic characteristics of dust both for establishing provenance and for the reconstruction of atmospheric transport patterns.

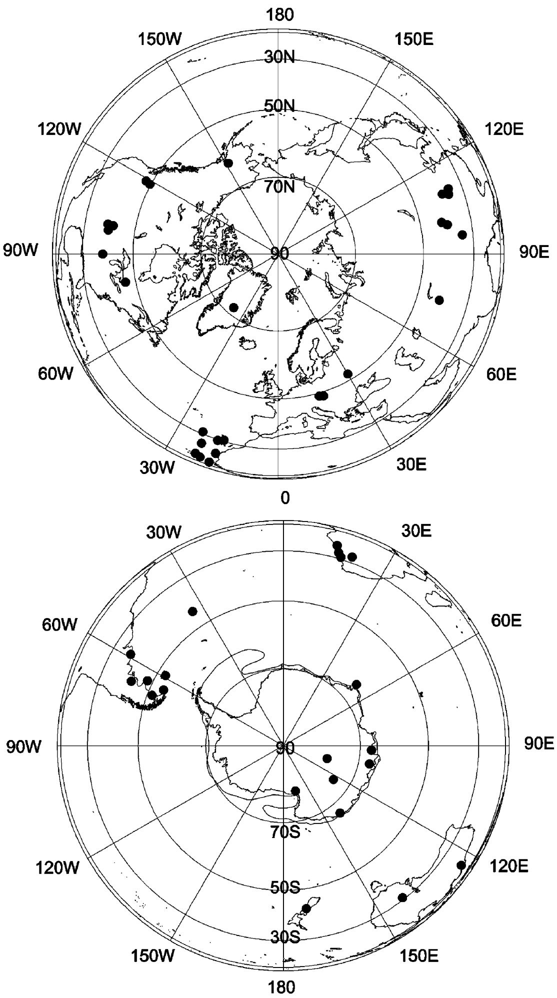


Fig. 2. Location of potential source areas of dust reaching the polar ice cores that have been sampled for mineralogy and radiogenic (Nd, Sr, and Pb isotope) analysis. Northern Hemisphere sites (Biscaye et al., 1997; Svensson et al., 2000); Southern Hemisphere sites (DePaolo et al., 1982; Grousset et al., 1992b; Basile et al., 1997).

Ice core chemistry has been used to infer the nature of the atmospheric circulation regime responsible for bringing dust to the polar ice cores. Using EOF analysis, Mayewski et al. (1994, 1997) have demonstrated that 76% of the variability in measured chemical species (derived from both continental dust and marine aerosols) can be explained by a single factor. Arguing that the abundance of material reaching the ice core is largely determined by the strength of the polar circulation cell, they suggest that the dominant EOF mode represents changes in size and intensity of the polar circulation cell, with high concentrations of chemical species occurring when the polar cell is most intense. Although a useful first step for understanding changes in dust transport to the poles, the polar circulation cell encompasses several potential source areas and processes, and therefore, this analysis does not provide specific evidence of what causes changes in dust characteristics in the polar regions.

The grain size distribution of aeolian material in ice cores is likely to be affected by the proximity of the source, wind intensity during transport, and the nature of the deposition process (i.e. the relative importance of wet or dry deposition). The polar ice cores are generally distant from potential source regions and thus usually contain only the fine-grained material reflecting long-distance transport of dust. Studies have demonstrated that the particle size distribution of dust in polar ice cores shows relatively small changes between climatic periods, in contrast to the total dust content which varies by an order of magnitude over the same periods (Steffensen, 1997; Zielinski and Mershon, 1997). Dust within the Greenland Ice Core Project (GRIP) ice core has a log-normal particle size distribution between 0.4 and 2.0 μm, and the modal particle radius decreases from 0.97–1.01 μm during the LGM to 0.87 μm during the late Holocene (Steffensen, 1997). Similarly, the mean diameter of dust in the GISP2 ice core decreases from 1.25 μm during the Younger Dryas period to 1.1 μm during the early Holocene (Zielinski and Mershon, 1997). In both cases, the small increases in grain size are correlated with increases in dust deposition, suggesting that some increase in wind intensity and/or intensity of the polar circulation cell may have accompanied increased Northern Hemisphere aridity during cold climate periods (Zielinski and Mershon, 1997). The dust deposited in Antarctica is also fine (particle radius size range 0.4–3.0 μm at Dome C and modal particle radius = 1–1.25 μm at Dome C and Vostok). Glacial dust deposited in the Antarctic ice cores is somewhat coarser than that characteristic of the Holocene, with modal radii of approximately 1.5 μm for the glacial samples (Thompson, 1977; Petit et al., 1981; De Angelis et al., 1984). Given that isotope fingerprinting suggests that the source region (Patagonia) did not change, these slight increases in grain size with increasing dust volume may suggest that winds were somewhat stronger during the LGM compared to the Holocene.

Changes in particle size distribution of dust could also reflect changes in the relative amounts of deposition due to rainout and dry settling of particles. The particle size distributions of dust deposited via dry and wet deposition processes on the Australian continent are distinctly different (Hesse and McTainsh, 1999). Dust deposited during rainfall events has a bimodal distribution (modal radii at 3 and 10.5 μm). Particles deposited by dry settling at the same location are coarser (modal radii at 12–13 μm) and the distribution is unimodal. This suggests that wet deposition removes fine particles more efficiently and that larger particles are more susceptible to dry deposition due to their faster gravitational settling. This study was completed near to a source region where particle sizes are distinctly larger than those found at great distances from source areas. Furthermore, some studies have concluded that the influence of depositional processes on particle size distribution occurs within 1000–2000 km of the source regions (Gillette et al., 1974; Johnson, 1979; Schultz, 1979). Thus, it is to be expected that the impact of depositional mode on the particle size distribution of dust found in distal locations such as the polar ice cores will be minimal.

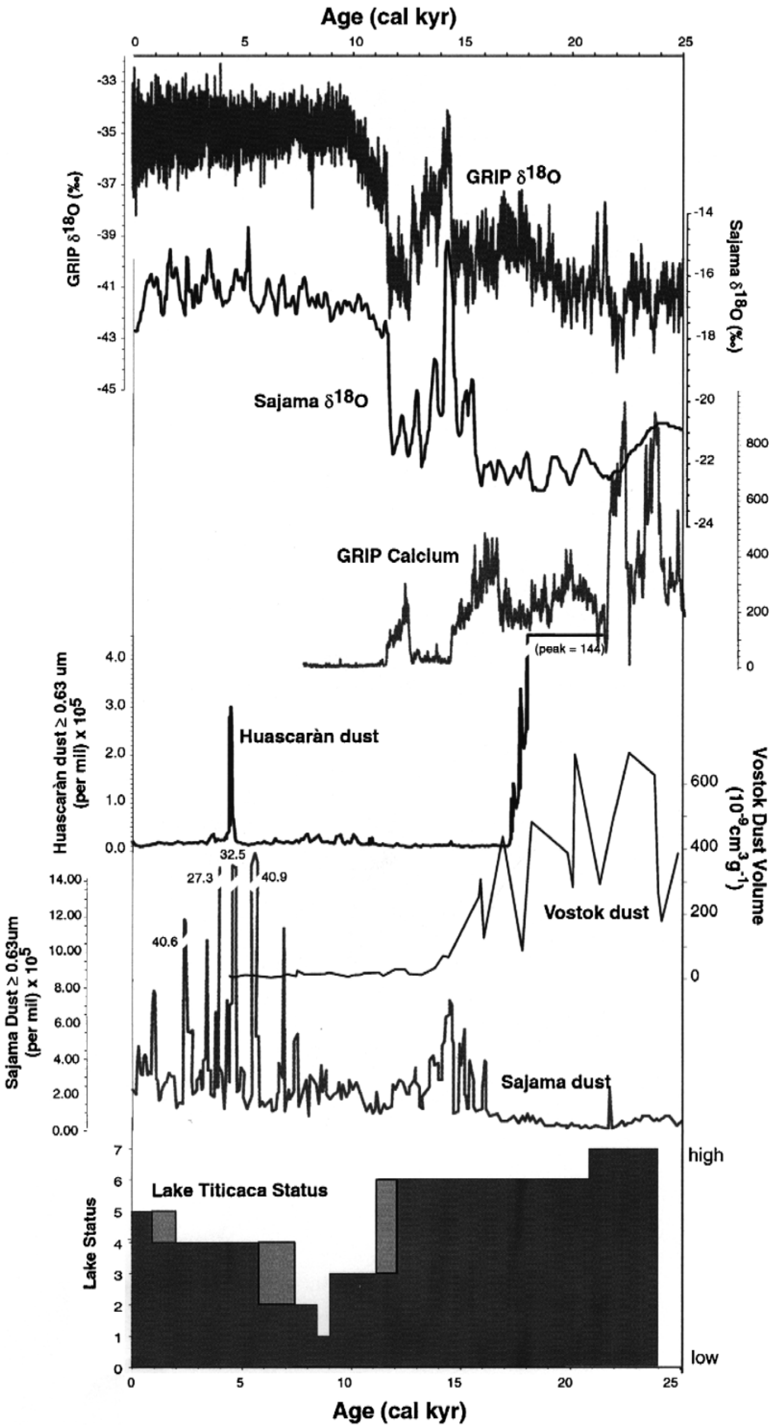
Potential source areas are closer to the locations of tropical ice cores. The dust record from tropical cores is therefore more likely to reflect the changing balance between long-distance transport of dust and dust transport from local sources than appears to be the case for the polar ice cores. The dust records from Huascaràn, Peru (6048 m) show a glacial–interglacial pattern that is similar to the pattern in the dust volume record at Vostok, suggesting that the dust reaching Huascaràn may reflect the global signal of increased aridity during the LGM (Thompson et al., 1995). However, the dust record from the Sajama ice core, Bolivia (6542 m), is not like the records from Huascaràn or Vostok (Fig. 3): the lowest concentrations of dust are found during the glacial period and increase dramatically during the Holocene period. Thompson et al. (1998) have suggested that the Sajama dust record most likely responds to local changes in dust source and is modulated by changes in the extent of Lake Titicaca (Thompson et al., 1998). The different glacial–interglacial patterns observed at the two Tibetan ice cores, Dunde and Guliya, are also likely to be the result of differences in local conditions (Thompson et al., 1997).

The ice core records provide a record of dust source areas (and by inference, dust trajectories), the regional atmospheric dust loading, and the effectiveness of dust depositional processes. Unfortunately, the ice core data provide this information for only a limited part of the world. There are only 18 ice cores with dust records (Table 1), 12 of which are located at the poles, with the remaining cores sampling South America and the Tibetan Plateau. Thus, other data sources must be used to acquire a global picture of changes in dust sources, distribution, and transport through time.

2.2. Marine sediments

Large regions of the ocean floor, in the deep basins, contain almost entirely atmospherically derived clay minerals. These regions may thus seem to be ideal locations for sampling dust records, but they contain no datable material because the biogenic components dissolve before deposition, and the basins are characterised by such low sedimentation rates that even isolating glacial–interglacial changes in dust deposition is challenging. Thus, most of the dust records from marine sediment cores come from open-ocean sites with higher sedimentation rates and biogenic deposition. The best dust records come from cores located long distances away from continental source regions: like the ice cores, therefore, they provide estimates of long-distance dust transport.

However, the interpretation of material in marine sediments is more complicated than interpreting the ice core records, because there are non-aeolian sources of terrestrial material in marine sediments.



Rea (1994) estimates that only 5% of the terrigenous material entering the oceans are atmospherically derived. In addition to the contribution of exported biogenic materials, marine sediments are affected by hemipelagic input, which includes biogenic debris and mud brought in from river plumes, storms, ice rafting, or low concentration turbidity currents bringing in material from continental shelves. Sites close to continents, on continental shelves, and near major river basins are most likely to be affected by these non-aeolian inputs. The width of the region affected has been estimated as ranging from 200 (Sarnthein et al., 1982) to 1000 km (Rea, 1994). In hemipelagic regions, terrigenous fluxes are usually of the order of 18–30 $g/m^2/year$, 1–2 orders of magnitude higher than typical aeolian fluxes found in open-ocean sediments. Hemipelagic contamination can be inferred from the presence of turbidites (Sirocko et al., 1991), from grain size characteristics (Sarnthein et al., 1982; Rea and Hovan, 1995; Joseph et al., 1998), or from other regional palaeoclimatic evidence documenting the climatic conditions of the neighbouring land-mass (Rea, 1994).

High-latitude sites (e.g. ~45° or higher) are likely to contain terrigenous material that has been entrained and subsequently melted out of icebergs. Ice-rafted detritus has been found as far south as 40°N in the North Atlantic Ocean (e.g. Ruddiman et al., 1981). Ice-rafted detritus can be readily identified because it is poorly sorted and contains material with a wide range of grain sizes.

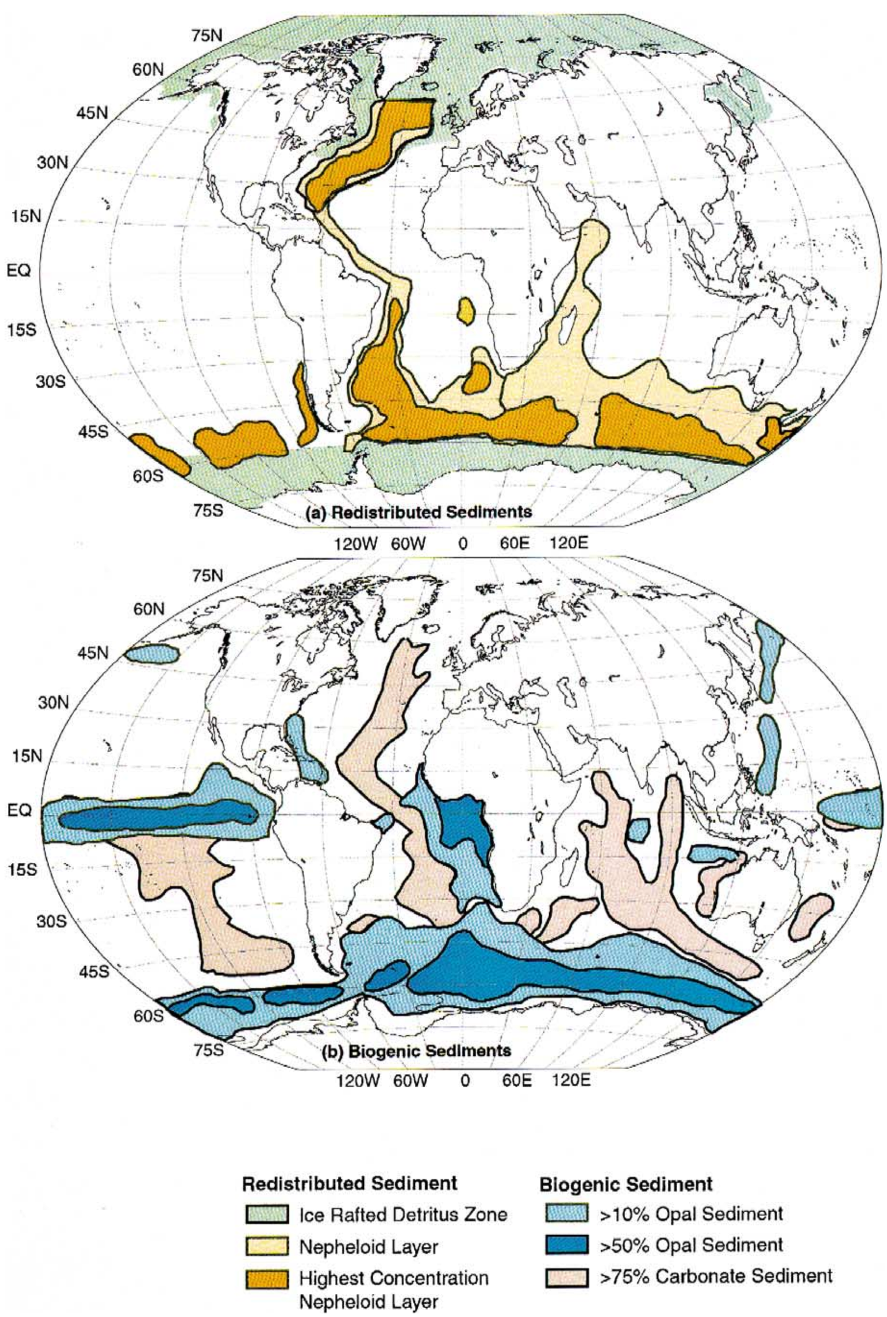
The marine sediment record can also be affected by redistribution processes within the ocean, specifically, sediment resuspension and focussing that results in the conflation of extra-local and local signals of accumulation. Resuspension and focussing occurs when material from shelves and ridges is entrained in strong western boundary or bottom water currents and then is transported within a 'nepheloid layer' along ocean basin floors (Biscaye and Eittreim, 1977; Damuth et al., 1983; McCave, 1986). The nepheloid layer usually contains particles that are < 2 μm in size, and ranges in thickness from 500 to 1500 m, depending on the energy imparted by the bottom water currents (McCave, 1986). The thickness of the nepheloid layer has been mapped extensively using light-scattering techniques to estimate particle concentration (see McCave, 1986). Changes in the grain size distribution between 10 and 63 μm has also been used to indicate the effect of bottom-water currents on winnowing and redistribution processes (e.g. Manighetti and McCave, 1995; McCave et al., 1995; Hall and McCave, 1998; Joseph et al., 1998).

It is possible to obtain reliable estimates of dust flux to the ocean only by careful site selection, and specifically by excluding records from the continental margins, from high latitudes, or from regions where the nepheloid layer is thick (Fig. 4). The regions that are likely to yield the best record of dust accumulation could change on glacial–interglacial timescales, as a result of changes in riverine input and/or ocean circulation patterns. The marine sediment records must therefore be evaluated individually, and additional measurements may be required to verify the aeolian nature of the materials analysed.

Even in ideal locations, it is necessary to isolate the aeolian component of marine sediments by removing biogenic material. Biogenic components of the sediments include calcium carbonate, silica and organic carbon. Calcium carbonate (largely tests of foraminifera and coccoliths) usually makes up 50–90% of the marine sediments. Biogenic silica (diatom and radiolaria frustules) usually comprises < 5% of the sediments, but higher concentrations are found in equatorial upwelling regions, in the high-latitude North Pacific, and in the Southern Ocean (Fig. 4). Carbonate and biogenic silica can be removed by leaching (Clemens and Prell, 1990) or the

Fig. 3. Ice core records reflect both global and local signals. The $\delta^{18}O$ of ice at GRIP (Dansgaard et al., 1989, 1993) and Sajama (Thompson et al., 1998) reflect global changes in the $\delta^{18}O$ of precipitation. The dust records from Greenland (Ca^{++} record, (Fuhrer et al., 1993), Huascaràn (insoluble particles, Thompson et al., 1995; Thompson, 2000), Vostok (insoluble particles, Petit et al., 1981), and Sajama (insoluble particles, Thompson et al., 1998). The GRIP Ca and Vostok dust concentrations are expected to reflect long-distance transport of dust from Asian (Biscaye et al., 1997) and Patagonian (Basile et al., 1997) deserts, respectively. The tropical sites of Huascaràn and Sajama can reflect a combination of short- and long-distance transport of dust. The dust record at Huascaran shows a glacial–interglacial pattern more similar to that of the polar ice cores, the dust record at Sajama seems to reflect changes in local source regions (Thompson et al., 1995). Dust concentrations at Sajama increase as lake status at nearby Lake Titicaca (Street-Perrott et al., 1989) decreases.

percentages of each may be measured by extracting these components from separate aliquots of the same sample (Ruddiman, 1997). Organic carbon makes up only a small percentage of marine sediments in most locations (< 1%) and therefore is not always eliminated from the aeolian component. Marine sediments may also contain authigenic oxides and hydroxides. Since these are a significant component of marine sediments from the Pacific and Indian Oceans, they have been removed prior to estimating the aeolian



fraction in some studies (e.g. Rea and Leinen, 1988; Leinen, 1989; Clemens and Prell, 1990; Hovan et al., 1991). The amount of authigenic material in Atlantic Ocean sediments has been considered negligible (Biscaye, 1965) and no attempt to remove it has been made in studies of dust accumulation rates from this region (Ruddiman, 1997).

The assumption that the calcium carbonate in marine sediments is biogenic in origin could lead to an underestimation of aeolian fluxes downwind of regions such as the Sahara. The carbonate content of dust from the southern Sahara and Sahel ranges from 4% to 8% and is between 20% and 50% of dust in the central and northern Saharan regions (Sarnthein et al., 1982). The carbonate component can be accounted for using isotope mass balance calculations. The oxygen isotope compositions of biogenic carbonate (e.g. foraminifera), the wind-blown carbonate (measured in the source regions), and the bulk carbonate composition in the marine sediments can be used together to isolate what fraction of the carbonate sediment might be wind-blown (e.g. Sirocko et al., 1991). Using this approach, Sirocko et al. (1991) found that 0–15% of the carbonate in marine cores from the Gulf of Arabia was aeolian. The largest amounts of lithogenic carbonate were found directly east of the Arabian Peninsula. In the future, estimates of the aeolian fraction of marine sediments should determine what fraction of the calcium carbonate is derived from continental instead of marine biogenic sources.

When sites have been carefully selected and appropriate techniques have been used to isolate the aeolian from the biogenic and/or authigenic components of the sediments, it is possible to derive a record of aeolian flux to the ocean at individual sites from the tropics, sub-tropics, and mid-latitudes. It is possible to characterise this material in terms of, e.g. particle size distribution and mineralogical and/or isotopic composition. These records can be interpreted in a similar fashion to the dust records from ice cores. However, the interpretation is not as straightforward as it appears to be for ice core records of dust because of (a) the possibility that material can be diagenetically modified after deposition, and (b) the closer proximity of most marine cores to dust source areas.

The total dust content of marine sediments is primarily a function of the extent of the source regions, wind strength, and distance from the source regions. The particle size distribution is a function of wind strength, distance from source region, and the process by which the dust is deposited (i.e. by dry settling or rainout). Since changes in both dust content and particle size can be caused by multiple factors, additional knowledge or corroborating data concerning atmospheric transport and/or conditions at the source regions must be used in order to interpret these data.

When the distance from the source area is relatively constant, the particle size reaches equilibrium with transport velocity, and the median grain size of the aeolian component of marine sediments can be interpreted as reflecting the intensity of transporting winds (Rea, 1994). This simplified relationship has been documented from the patterns in median grain size recorded over one annual cycle in modern marine sediment traps in the Arabian Sea, where dust

Fig. 4. Schematic maps of the distribution of non-aeolian material in marine sediments, either by (a) redistributed sediments or (b) biogenic materials. In (a) green-shaded area indicates areas potentially contaminated by poorly sorted detritus transported by icebergs and sea ice, as estimated from the maximum LGM sea-ice extent (CLIMAP, 1981). Orange and yellow shaded regions show areas in which the concentration of particles within the nepheloid layer is highest, and therefore where terrigenous material is most diluted by sediment redistribution in bottom currents (Biscaye and Eittreim, 1977; McCave, 1986). Orange regions denote areas of greatest contamination, and beige areas suggest regions where caution must be exercised in interpreting dust records. In the Atlantic regions, orange and yellow shading represents the regions where the concentrations of particles that are mixed upwards into the nepheloid layer are >500 and 100 $\mu g/cm^3$, respectively (taken from Biscaye and Eittreim, 1977). In the Indian and Pacific Oceans, orange and yellow shading represent regions of 'excess turbidity' within the nepheloid layer >0.6 and 0.2, respectively. Here excess turbidity is defined as $\log E/E_c$, where E is the maximum light scattering near the bed and E_c is value at the clear water minimum (McCave, 1986). A value of 1 represents a factor of 10 increase from the clear water value. In (b), light blue shading indicates areas where opal concentrations in marine sediments are $>10\%$; opal concentrations in sediments are $>50\%$ in dark blue regions (Broecker and Peng, 1982). Pink regions indicates areas where carbonate is $>75\%$ of marine sediments (Broecker and Peng, 1982). Special care must be taken to separate the biogenic and terrigenous materials in regions with large opal and carbonate concentrations indicated.

deposition occurred primarily by gravitational settling (Clemens, 1998). Here, high correlations ($r = 0.84–0.93$) are found between median grain size diameters and overlying wind speed, barometric pressure, and regional pressure gradients. Furthermore, the similarity between particle size distributions in surface marine sediments and sediment trap samples collected during the summer months illustrates that the aeolian material in surface sediments primarily reflects the aeolian characteristics of the southwest summer monsoon, when most of the dust is delivered to the sediments. These high correlations indicate that under modern conditions, the grain size record of dust in the Arabian Sea predominantly reflects regional changes in atmospheric transport strength during the southwest summer monsoon.

Sites in the central part of the ocean may be sufficiently remote from dust source areas that even quite large changes in the extent of source areas has little impact on their proximity to the sampling site and hence on grain size (see Rea, 1994). However, the expansion of source areas may still affect the total amount of aeolian material reaching these sites. In North Pacific records, periods of maximum dust fluxes coincide with maximum glacial periods, but grain size records vary at higher frequencies than those observed in the global ice volume record (Janacek, 1984; Hovan et al., 1991). This change in frequency distribution has been interpreted as suggesting that the changes in grain size reflect changes in wind intensity.

At sediment sites closer to dust source regions such as the eastern equatorial Atlantic, the combined effects of both source area changes and wind speed are likely to affect the median grain size of aeolian material, and thus other proxies are necessary to decipher these records. Increases in both median grain size and aeolian accumulation rates in the equatorial Atlantic Ocean at the LGM can be interpreted as indicating enhanced wind speeds (see Sarnthein, 1978; Sarnthein and Koopman, 1980; Ruddiman, 1997), largely because supplemental data from foraminiferal assemblages and productivity proxies suggest intensification of local winds at the LGM (e.g. Ravelo et al., 1990; Ruddiman, 1997).

The clay mineralogy and tracer isotope composition of the terrigenous component of marine sediments has been used to determine the potential source areas of dust reaching the ocean (see, e.g. Biscaye, 1965; Biscaye et al., 1974; Goldberg and Griffin, 1970; Kolla et al., 1976, 1981; Grousset et al., 1988, 1998). For example, palygorskite clay in the Arabian Sea and Indian Ocean is derived only from Saudi Arabia and therefore serves as a discrete tracer of winds from this region (Kolla et al., 1981; Sarnthein et al., 1982; Sirocko and Lange, 1991). The interpretation of clay mineralogy in marine cores is complicated because the clay fraction of terrigeneous material is most easily redistributed by oceanic bottom currents. However, even in regions where fluvial and hemipelagic contamination is ubiquitous, information about marine sediment mineralogy provides a useful, first-order smoothed estimate of the mineralogy of the nearby continental source areas (Biscaye et al., 1997), which may prove useful in defining potential transport paths of dust. Coupled with geochemical isotope measurements, these data may be used to determine changes in the relative proportion of various sources of terrigenous input (e.g. Sirocko and Sarnthein, 1989; Nakai et al., 1993; Grousset et al., 1998).

Isotopic data have also been used to identify the sources and trajectories of dust found in marine cores (Sirocko and Sarnthein, 1989; Nakai et al., 1993; Grousset et al., 1998). Between the Equator and the Canary Islands, carbonate-free modern surface marine sediment and aerosol samples (< 30 μm) have $^{143}Nd/^{144}Nd$ and $^{87}Sr/^{86}Sr$ isotope signatures resembling radiogenic dust sources in Morocco, Mauritania, and Mali (Grousset et al., 1998). The distribution is consistent with atmospheric transport of aerosols along the NE–SW axis of the Saharan Air Layer. The pattern and latitudinal distribution of dust at the LGM is unchanged, although glacial marine sediments south of the Cape Verde Islands ($\sim 16°$N) are less radiogenic than modern sediments. This change could be the result of either an increase in dust from the Sahel region (which is relatively less radiogenic than the dust sources in the western Sahara), or simple dilution of the highly radiogenic background signal of Atlantic sediments by increased dust input from the western Sahara. Similar analyses have been conducted in Pacific sediments. Nakai et al. (1993) used Nd and Sr isotope compositions and Rb, Sr, and rare earth element (REE) concentrations to determine the provenance of surface sediments in the North Pacific Ocean. The isotope data suggest that Central Pacific sediments share the same isotope composition as Asian loess deposits, North Pacific sediments are dominated by radiogenic signatures of island arc volcanism, and the eastern equatorial Pacific sediments reflect the signatures of South America, with coastal sites of North America dominated by hemipelagic deposition.

On longer (tectonic) time scales, the clay mineralogy of marine sediments have been used to determine changes in the overall weather regime of continents (Arnold et al., 1995). Climate modelling studies have suggested that the uplift of the Tibetan Plateau may have resulted in an overall cooling of northern Asia and an aridification of the Eurasian interior (Kutzbach et al., 1989; Ruddiman et al., 1989). Arnold et al. (1995) hypothesise that this climate shift is responsible for radical changes in the weathering regimes across the entire Asian continent, impacting the mineralogy of North Pacific sediments, i.e. because of uplift and cooling, physical weathering products (i.e. chlorite, quartz, illite, and plagioclase) increased at the expense of chemical weathering proxies (i.e. kaolinite, smectite). Analysing changes in weathering regimes on shorter time scales requires detailed mineralogical studies of the potential source regions.

2.3. Terrestrial (loess) deposits

Although the aeolian origin of loess is usually unequivocal, the interpretation of loess deposits is complicated because they are proximal to dust source regions, and can serve as both sources and sinks of dust. Furthermore, loess deposits contain a mixture of aeolian material derived from local aeolian reworking, short-distance transport from nearby continental sources, and longer-distance (hemispheric) transport. The importance of local sources and reworking is reflected in the fact that loess is generally much coarser than aeolian material in marine cores. Only ca. 30% of the dust from eastern Colorado is finer than 8 μm (Muhs et al., 1999)—a size fraction comparable to 95% of the aeolian materials found in marine sediments (Rea, 1994). These large particle sizes indicate close proximity to dust source regions.

Particle size analyses of loess deposits have been made along transects in NW USA (Busacca and McDonald, 1994), mid-continental USA (Muhs et al., 1999) and the Loess Plateau (Liu et al., 1985; Eden et al., 1994), and used to identify source areas and the trajectory of aeolian transport from the source region. When it can be assumed that the source area has been constant through time, then changes in particle size at a single site through time can be used to infer changing wind strength. Such reconstructions have been made at key sites from China (e.g. Porter and Zhisheng, 1995; Chen et al., 1997). The interpretation of particle size data from loess is complicated by the fact that post-depositional processes under different climatic conditions can preferentially enhance the amount of fine material within a loess deposit (Derbyshire et al., 1995). However, micromorphological studies (Kemp et al., 1995) can be used to identify the presence of translocated fine material in the form of, e.g. clay skins. It should therefore be possible (though time-consuming) to estimate the proportion of fine material associated with the primary aeolian deposit relative to that associated with secondary fabric features, and thus due to post-depositional processes.

The thickness of loess deposits has also been used as an indicator of palaeowind vectors (e.g. Frazee et al., 1970; Handy, 1976; Ruhe, 1983). In the mid-continental USA, thickness trends in the last glacial loess have been used to indicate that the predominant surface winds came from the west and northwest at the LGM (e.g. Thorp and Smith, 1952; Ruhe, 1983). Loess thicknesses are greatest near dust sources along rivers, at the southern edge of the Late Wisconsin ice sheet, and directly downwind of the Rocky Mountains, and rapidly decrease to the east of these source regions (Fig. 5). On the Columbia Plateau in the Northwest USA, maps of loess thickness suggest that source sediments for the late Quaternary loess deposits were in the southern part of the plateau, and that winds came from the south and southwest, similar to today (Busacca and McDonald, 1994).

Just as mineralogical and radiogenic isotope information can be used to identify potential source areas for dust reaching the polar ice cores or marine sites, so a similar approach can be used to identify the source areas for loess deposits (e.g. Aleinikoff et al., 1999; Unruh et al., 1999). In the mid-continental

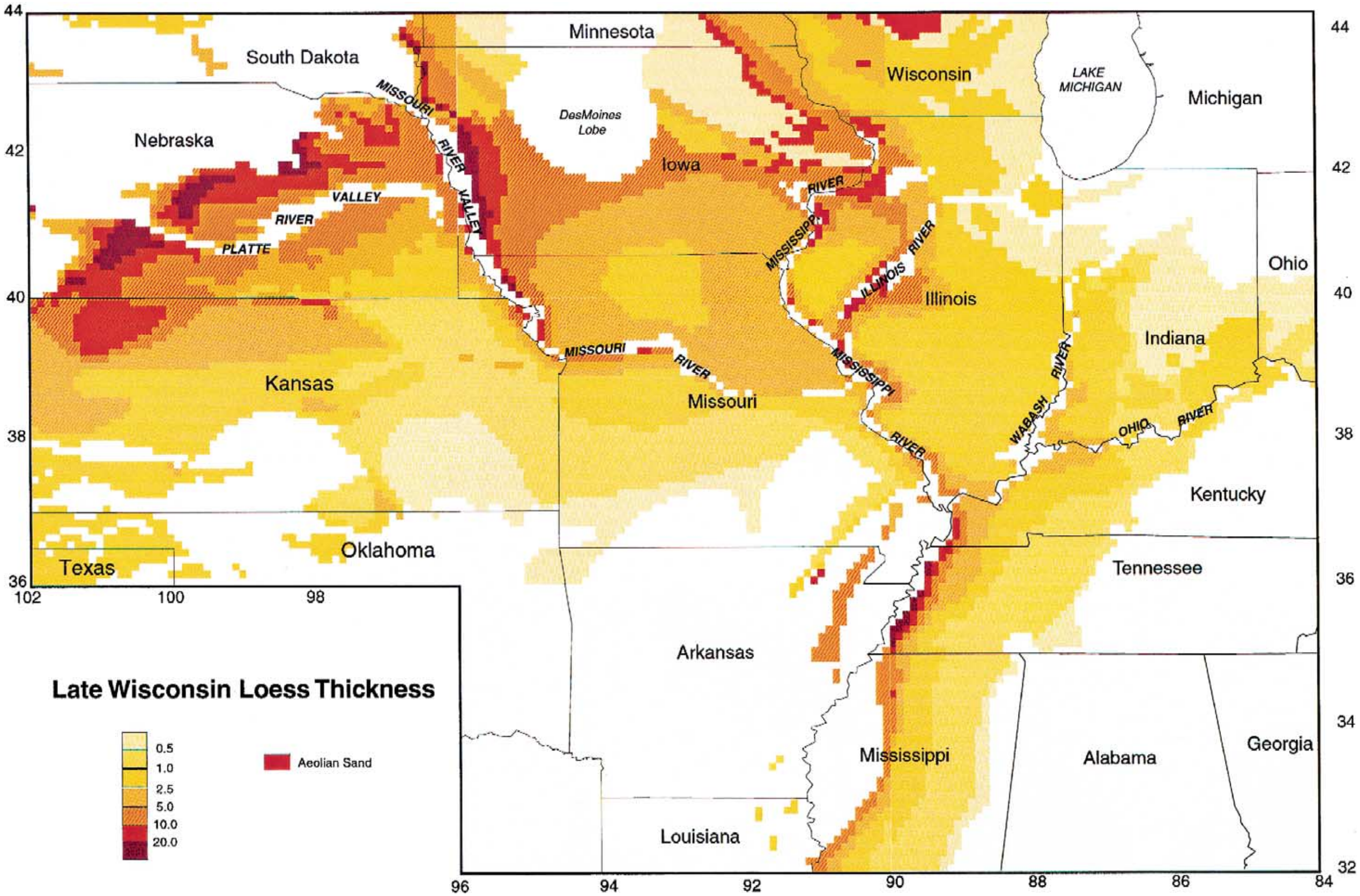


Fig. 5. Thickness of the Late Wisconsin loess in the mid-continental USA, demonstrating the thinning of loess deposits downwind of source regions. Data are gridded at $0.1 \times 0.1°$ from maps from the US Geological Survey (taken from (Lineback et al., 1983; Miller et al., 1988; Holbrook et al., 1990; Gray et al., 1991; Hallberg et al., 1991; Denne et al., 1993; Whitfield et al., 1993; Swinehart et al., 1994).

USA, Pb isotope measurements on potassium feldspar grains found in the Peoria Loess (Colorado) show systematic variability throughout the period of deposition between 20.0 and 11.8 ka (Aleinikoff et al., 1999). During warmer periods, loess deposits are derived from glaciogenic silts downwind of the valley glaciers in the Front Range. During times when these glaciers were at their maximum extent, the Pb isotope signatures suggest that dust was derived from a more southerly source region which became active as a result of strong winds and reduced vegetation cover. Such studies, although requiring careful sampling to avoid reworked loess, are a promising avenue of future research for reconstructing changes in source region and wind direction.

Mineralogical analyses of loess deposits have been used to characterise the mineralogy of potential source regions and to assess their relative contributions to ice cores (Biscaye et al., 1997), aerosol samples (Leinen et al., 1994; Arnold et al., 1998), and marine sediments (Leinen, 1989). However, information about the mineralogy of loess deposits may be crucial for reasons other than sourcing. Recent modeling studies have suggested that the radiative properties of airborne dust are strongly influenced by the relative proportions of gypsum, calcite, quartz and iron oxides (hematite in particular) present (Claquin et al., 1998). The use of mineralogical data from loess deposits to characterise the mineralogical properties of airborne dust relies on the assumption that there is no significant sorting. Analysis of aerosol samples from the North Pacific suggests that the transport-related mineral fractionation drives the aerosol mineralogy towards a clay mineral enriched and primary mineral depleted composition with increasing transport time (Arnold et al., 1998). Thus, to study the potential radiative impact of dust mineralogy, it is more useful to determine the mineralogical properties of the clay fraction (< 5 μm) of dust in loess deposits. In order to use the mineralogical analyses of loess sediments in studies of the radiative properties of dust, it is important that measurements encompass the full range of radiatively important minerals. For example, in the mineral analyses from the Loess Plateau, the iron oxides, which are perhaps most important to radiative calculations, are routinely eliminated (Eden et al., 1994). Future mineralogical analyses from potential source regions should take into account which components of dust are most likely to influence the radiative properties.

3. The role of earth system models

Changes in dust properties (i.e. total flux, particle size, mineralogy) at a particular location may be explained in several different ways. Thus, changes in dust flux might be caused by a combination of factors including (a) changes in the areal extent of source regions, (b) changes in dust emission efficiency (wind speed over source regions) or atmospheric transport of dust, (c) changes in the residence time of dust in the atmosphere, and/or (d) variations in the relative contributions of dry settling or rainout of dust over a given region. Similarly, changes in mineralogy could reflect (a) changes in the source region (b) shifts in atmospheric transport from the source regions to the dust deposition site, or (c) changes in the mineralogic composition of the source region resulting from changes in weathering. While it may be possible to eliminate some explanations, by drawing on supplementary evidence, interpretation of the causes of observed changes in dust properties rarely yields an unequivocal answer.

One way to examine the relative contributions of the possible factors contributing to changes in dust deposition is to use a forward modelling approach, in which process-based models can be used to predict the response of palaeoenvironmental variables (e.g. vegetation, dust deposition, lake distributions) to simulated climate changes. The predicted response can then be compared directly with palaeo-observations. One such forward simulation of the dust cycle (Mahowald et al., 1999) has been used to assess the relative contributions of climate (e.g. winds and hydrological cycle) and changes in potential source areas of dust due to changes in vegetation and soil moisture. This study showed that the changes in precipitation and wind intensity were sufficient to reproduce the 3- to 5-fold increase in dust deposition off western Africa at the LGM. However, an expansion of central Asia and high latitude source areas was required to explain the overall global dust deposition, and the 20-fold increase observed in high latitudes. The fact that the relative contributions of source area aridity and climatological factors apparently vary from region to region reinforces the need for a global perspective in interpreting the geological record of dust. Furthermore, while regional studies of dust deposition have been completed, a global overview of aeolian distribution and accumulation is still needed. Available data still represent only discrete points in a spatial array. In order to achieve a comprehensive picture of global patterns of dust accumulation rates, their controlling mechanisms, and how those mechanisms may vary regionally, global earth system models are required.

4. The DIRTMAP data base

Global validation data sets should be developed and used in parallel with earth system models of the dust cycle. Validation datasets help to quantify changes in potential dust source areas (e.g. using maps of vegetation and loess accumulation), as well as identifying the magnitude and extent of dust during past climatic periods. The Dust Indicators and Records of Terrestrial and Marine Palaeoenvironments (DIRTMAP) database was originally developed to serve as a validation data set.

The database contains the aeolian information from ice cores, marine sediments, and terrestrial (e.g. loess and lake sediment) deposits, focussing on information from the last glacial–interglacial cycle (i.e. 30 ka). Types of information incorporated in the database include (1) aeolian accumulation rates and the information required to derive them (i.e. age models to derive sediment accumulation rates, fraction of sediments considered aeolian, and sediment bulk density), (2) radiogenic isotopes and clay mineralogy important to interpreting dust provenance and composition, (3) grain size information for evaluating wind direction, intensity, and potential contamination, and (4) chronological data. As with other palaeoenvironmental data sets, these data are accompanied by adequate documentation and "metadata" (e.g. site type, depositional environment, chronological control, etc.) to allow users to choose data appropriate for specific analyses.

5. Estimating aeolian accumulation rates: methods and problems

Aeolian accumulation rates can be derived from ice cores, marine sediments, and loess deposits. The mass accumulation rate of aeolian material (MAR_{eol}) is estimated as

$$MAR_{eol}(g/m^2/year) = AR(m/year)\, f_{eol} BD(g/m^3)$$

where AR is the bulk accumulation rate, f is the mass concentration of aeolian materials within the sample, and BD is the bulk density (Table 2).

Thus the estimation of MAR_{eol} requires that (a) it is possible to isolate the aeolian component, (b) the

Table 2
Calculation of eolian accumulation rates in different regimes: $MAR_{eol} = AR\ (m/year)\ f_{eol}\ \rho\ (g/m^3)$

Environment	Sedimentation rate	Eolian fraction	Density
Marine sediments	• $\delta^{18}O$ stratigraphy • ^{14}C dates	eliminate: • $CaCO_3$ • opal • organic C • authigenics	• direct measure • salinity model • correlation with $CaCO_3$, opal
Ice core	• ice flow models • layer counting • stratigraphic correlation	• dust concentration (mass/vol)	
Loess	• TL • ^{14}C • MS/PSA correlation • tephrochronology	$f = 1$	• 1.65 g/cm^3 (Pye, 1987) • sand/silt/clay loess classification

chronological control is good, and (c) estimates of bulk density are available from the sampled sediment. Estimates of these variables are made differently in the different types of geological archive.

5.1. Ice cores

In ice cores, the bulk accumulation rate is equivalent to the ice accumulation rate. Ice accumulation rate is calculated based on the age model developed for the ice core. Several different methods are commonly used to establish the age of the ice, including ice flow models (e.g. Jouzel et al., 1993), counting of annual layers (e.g. Taylor et al., 1992; Alley et al., 1997), and stratigraphic correlation using CH_4 and $\delta^{18}O_{air}$ records (e.g. Sowers et al., 1993; Blunier et al., 1997).

Ice chronology can be estimated by directly modelling ice sheet dynamics and computing the ice-flow lines (e.g. Lorius et al., 1985; Jouzel et al., 1993). However, this strategy introduces significant uncertainties (± 20 ka) at depths of 2500 m, resulting from lack of information about the current ice accumulation rates upstream from the ice core site (Jouzel et al., 1993). Stratigraphic correlation using both oxygen isotope and trace gas measurements has also been used to provide a common temporal framework for comparison of ocean records and ice core records from both hemispheres. Some ice cores with multiple annual indicators have reduced the estimated uncertainty of ice-core ages to about $\pm 1\%$ at the end of the Younger Dryas Period (about 11.5 ka; see Alley, 2000), but can be used only in the most recent parts of ice cores. Correlation between oxygen isotope measurements on air and oxygen isotope chronology developed for marine sediments (e.g. Martinson et al., 1987) can introduce several errors. These errors include uncertainties from tuning the records (2.5–4 ka), uncertainties in the marine isotope chronology itself (as much as ± 5 ka; Martinson et al., 1987), and the assumption that the variables measured in the ocean and in ice are contemporaneous. Finally, when using measurements on the air in ice cores, one must estimate the ice age–gas age difference: a difference by which the age of the air is younger than the surrounding ice, reflecting the gradual process by which gases are trapped in the ice. The total uncertainties in stratigraphic correlation have been estimated to be within the range of ± 3–6 ka (Salamatin et al., 1998).

The dust content in ice cores can be measured and therefore reported in several ways. The number of dust particles can be counted using, e.g. a Coulter counter (e.g. Thompson and Mosley-Thompson, 1989). The total insoluble mass can be measured either via laser light scattering (Petit et al., 1990; Hansson, 1993), or through the concentration of key elements such as calcium and aluminum by atomic adsorption (e.g. Fuhrer et al., 1993). Al concentrations are then converted to mass concentrations by assuming that dust has an 8% Al content. Most models produce mass-based predictions of dust deposition: for this reason, we standardise all measurements in DIRTMAP by converting them to mass accumulation rates or ratios between accumulation during a given period and the modern baseline. In DIRTMAP, we assume that all measurement methods produce broadly comparable results, although there are differences in calculated LGM/Modern ratios depending on whether mass-based, number-based or calcium estimates are used. Dust and calcium content are highly correlated, but the LGM/Modern ratio is smaller for dust than for calcium for unclear reasons (see discussion in Hansson, 1994; Steffensen, 1997). Thus, for mass accumulation rates, only dust particle measurements are used. Mass concentrations of dust are simply converted to mass accumulation rate by multiplication by the ice accumulation rate. Conversion of number concentrations to mass concentrations is not entirely straightforward, in part because of uncertainties in the bulk density of dust (values used range from 2 to 2.8 g cm^{-3} for mineral dust; Taylor and McClennan, 1985; Giorgi, 1988; Petit et al., 1990; Clemens, 1998; Zhang et al., 1999). In ice core studies where only particle number was reported, only the LGM/Modern ratios are estimated, and included in DIRTMAP.

One advantage of dust measurements from ice cores is the possibility of recovering high-resolution changes in dust content. For example, Steffensen (1997) reports maximum dust concentrations over short time intervals (ca. 100–500 years) during the LGM that are as much as 168 times greater than mean Holocene values. However, since this type of resolution is not readily achievable in either marine sediments or loess deposits, the ice core record needs to be averaged in some reasonable way to ensure comparability between the sensors.

5.2. Marine sediments

In marine sediments, BD is determined via the dry bulk density which can be estimated either directly, using salinity models in which the original water content of a dried sample is inferred from the measured amount of dried salt and an assumed paleosalinity, or by correlation with the calcium carbonate or opal content (Rea and Leinen, 1988; Hovan et al., 1991; Ruddiman, 1997). The weight percent of dust is determined by isolating the terrigenous fraction of the sediment by elimination of carbonate, opal, oxides and hydroxides (Rea, 1994).

Non-carbonate accumulation rates are an approximation of the dust flux, except in regions of the ocean where the non-carbonate component is expected to be dominated by non-aeolian components (e.g. in equatorial and upwelling regions of the ocean where the opal content is high, or in high-latitude regions where sediments are influenced by ice-rafted contamination). DIRTMAP currently includes non-carbonate accumulation rate estimates from several marine cores (Catubig et al., 1998).

Sediment age models on marine records are determined using radiocarbon dates where available (e.g. Sirocko et al., 1991; Ruddiman, 1997): where they are not, correlation is effected with oxygen isotope stratigraphies (e.g. Imbrie et al., 1984; Martinson et al., 1987). The temporal resolution of marine sediment records is generally much lower than that of the ice core records, with higher analytical uncertainties associated with the lower resolution cores (Mahowald et al., 1999). Furthermore, because the number of radiocarbon ages is limited, dust deposition rates may be averaged over several thousand years. Time-averaging minimises the error associated with determining linear sedimentation rates on few radiocarbon dates, but results in loss of information and can lead to biases if there are significant climatic trends within the sampling interval. For example, there were significant changes in monsoon intensity most likely associated with changes in dustiness off Africa during the time period normally averaged off northern Africa (3–11 ka B.P.; Ruddiman, 1997). Thus, increasing the number of radiocarbon measurements could significantly improve the existing marine records of dust.

5.3. Loess deposits

As a first approximation, loess deposits are assumed to be entirely aeolian in origin. Bulk density (BD) measurements on individual loess units or sections are rarely reported. Pye (1987) suggests the average BD of loess is 1.65 g/cm^3. We use this value to calculate MAR on loess, if no specific BD measurements are available. However, measurements on last glacial loess from the Chinese Loess Plateau have yielded BD values from 1.281 to 1.632 g/cm^3 (Liu, 1966). BD values on soils and loesses within the Heimugou Loess section (China) range from 1.40 to 1.65 g/cm^3 during the last glacial–interglacial cycle (Liu et al., 1985). Thus BD values vary by $>$ 18% depending on whether or not a soil or a loess deposit is forming. Adopting an average BD for the calculation of MAR could lead to non-negligible errors in the MAR estimates (Fig. 6).

Determining bulk accumulation rates also requires detailed age models for each stratigraphic section analysed. Estimates of dust accumulation rates from loess deposits currently rely upon the interpretation of both dating and stratigraphic tools, including ra-

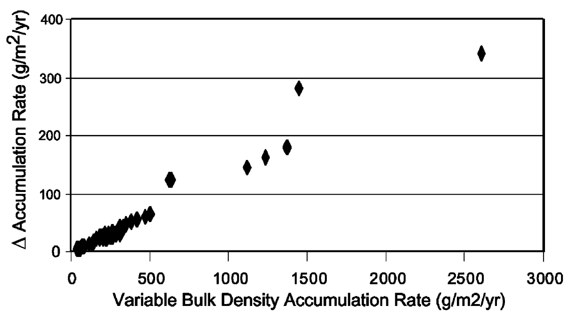


Fig. 6. Comparison accumulation rates of loess from the Chinese Loess Plateau estimated using a constant bulk density (BD) of Pye (1987) and variable BDs taken from (Liu, 1966). The $\triangle$ accumulation rate is the difference between rates calculated using a constant BD of 1.65 g/cm^3 (Pye, 1987) and rates calculated using the appropriate BDs for the type of loess (1.38, 1.46 and 1.5 g/cm^3 for sandy, silty and clay loess, respectively) at each site. The comparison shows that aeolian accumulation rates could be overestimated by as much as 15–20% when the constant BD for loess sediments is used.

diocarbon dating of soil layers, luminescence dating of the loess itself, stratigraphic identification of last glacial loess, and correlation of magnetic susceptibility records with the marine oxygen isotope stratigraphy. The number of reliable dates on loess deposits, using either radiocarbon or luminescence techniques that can be used to determine deposition rates during the LGM is often limited. Measurements of the bulk concentration of magnetic minerals or bulk magnetic susceptibility has been correlated on many occasions with the marine oxygen isotope stratigraphy (Kukla, 1987), and has been used extensively as a chronostratigraphic tool on the Chinese Loess Plateau. These measurements have alternatively been interpreted as an indicator of enhanced soil formation and moisture, as both the concentration and grain size of magnetic minerals appear to be altered under these conditions (e.g. Zhou et al., 1990; Maher and Thompson, 1991). Regional patterns of magnetic susceptibility may be negatively correlated with the marine oxygen isotope record (e.g. in central Europe; Chlachula et al., 1998), or show no enhancement in magnetic susceptibility (e.g. in Alaska; Beget, 1990; Vlag et al., 1999). Thus, while several studies suggest that magnetic susceptibility measurements track climate changes on the Chinese Loess Plateau, their use as a chronostratigraphic tool should be treated with caution because (a) much remains to be learned about the mechanism controlling these changes, and (b) the time lag between the processes controlling changes in climate and the enhancement of magnetic minerals in soils is not yet well understood. Accordingly, the use of the magnetic susceptibility signal in loess as a continental or global stratigraphic tool is still questioned. Clearly, until better radiometrically dated loess chronologies are available, there will be uncertainties of as much as several thousand years inherent in the loess chronologies (Sun et al., 2000). These uncertainties will necessarily affect estimates of accumulation rates.

The lack of dating control on modern and late Holocene loess also complicates the estimate of glacial–interglacial changes in loess accumulation. In many regions, these modern and late Holocene time periods are represented by soil formation, and deposits have been extensively modified by human agriculture. Modern accumulation rates could be estimated using modern trap samples (Zhang et al., 1997), but these estimates will include anthropogenic effects. Furthermore, modern trap measurements must be of sufficient length to provide an average dust flux estimate over several years, and must include sample periods of both high and low dust content. Another solution is to include longer time series of dust deposition, and then to estimate a "glacial" and "interglacial" average.

6. Observed patterns in dust deposition

Currently, the DIRTMAP data base contains a total of 426 sites (Fig. 7). Of these sites, 253 are from marine sediments, 56 from marine sediment traps, 18 from ice cores, and 99 from terrestrial (loess) deposits. The first application of this data base was to document the patterns of dust deposition rates for the late Holocene and LGM periods, to be compared with simulations of these two different climate states (Mahowald et al., 1999). This will be the focus of the discussion here. In general, dust records confirm the regional estimates that the earth was considerably dustier at the LGM than today, although spatial patterns are evident. Records of high-latitude changes in dust deposition are found solely in polar ice cores and suggest the largest glacial–interglacial increases (2–20-fold) in dustiness. Marine sediment records from low- to mid-latitude regions suggest a smaller glacial–interglacial increase of up to five times, with some equatorial regions suggesting reduced dust fluxes at the LGM. The expansion of loess deposits, primarily in midlatitudes, occurred both downwind of desert regions (e.g. on the Chinese Loess Plateau and the Pampa regions of South America) and directly downwind of the edges of major ice sheets (e.g. in North America and Europe).

6.1. Ice cores

Ice core records from Greenland and Antarctica show glacial increases in dust deposition rates from 2 to 20 times compared with the present, although the actual amount of dust deposited during the glacial is very small (0.004–0.34 $g/m^2/year$; Cragin et al., 1977; Fisher, 1979; Petit et al., 1981, 1990; Thomp-

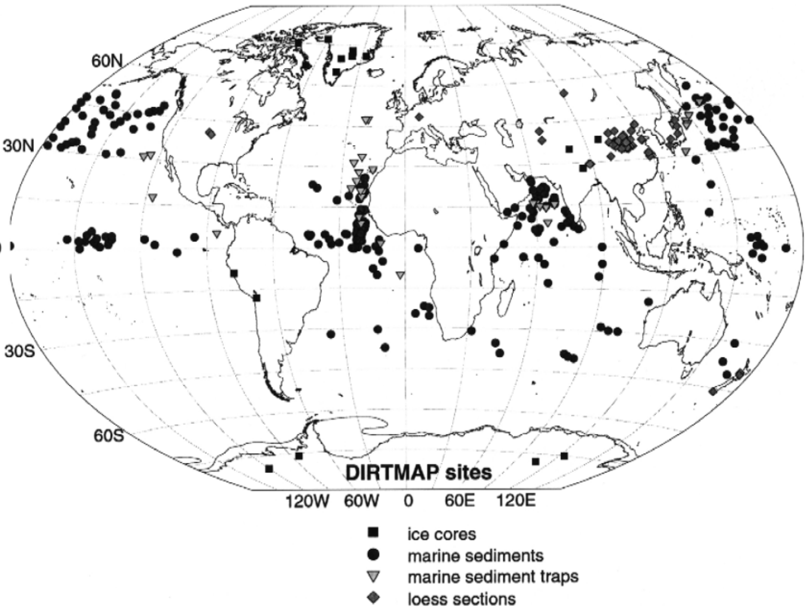


Fig. 7. Distribution of all sites in the current version of the DIRTMAP database, including ice cores, marine sediments, sediment traps, and loess deposits.

son and Mosley-Thompson, 1981; Hammer et al., 1985; Hansson, 1994; Mahowald et al., 1999). Some tropical ice cores show a similar signal, with increased dust deposition rates during the glacial. However, changes in the tropical ice cores are more likely to be influenced by local conditions. For example, dust deposition rates in Huascaràn, Peru (Thompson et al., 1995), were greater at the LGM, but dust deposition at the nearby core in Sajama, Bolivia was less at the LGM, most likely in response to the expansion of local lakes (Fig. 2) (Thompson et al., 1998). A similar situation is seen on the Tibetan Plateau where dust deposition rates at Dunde increased 1.5–4 times at the LGM compared to modern rates, and the LGM deposition rates at Guliya (on the southwest of the Plateau) were less than 1/5 of current values.

6.2. Marine sediments

Glacial increases in dust deposition downwind of different continental dust sources are not spatially uniform (Fig. 8). Furthermore, data from some regions show little or no change in dust deposition, or even decreased dust deposition at the LGM.

6.2.1. North Pacific Ocean

Northeast of Asia in the North Pacific Ocean (~ 28–45°N), LGM dust deposition rates range from no increase to values four times higher than late Holocene dust fluxes, but most LGM values show 1–3-fold increase over interglacial values, with decreasing LGM fluxes towards the tropical regions. The magnitudes of deposition rates range from 0.6 to a maximum value of 10 $g/m^2/year$, with most late Holocene values in the range of 1–5 $g/m^2/year$. Higher deposition rates are recorded in modern sediment traps (10–25 $g/m^2/year$) that were deployed for less than 50 days and most likely are biased towards seasons when dust fluxes are high.

6.2.2. Arabian Sea

The average increase in dust deposition rates observed in the Arabian Sea for the LGM are small (60–80% increase), although the LGM/Late Holocene flux ratios range from 0.6 to 3.6 times greater at the LGM. The slight reductions in LGM dust accumulation rates compared to Late Holocene values occur in the southern regions of the Arabian Sea. Sirocko et al. (1991) hypothesise that the general increase in glacial dust results from both in-

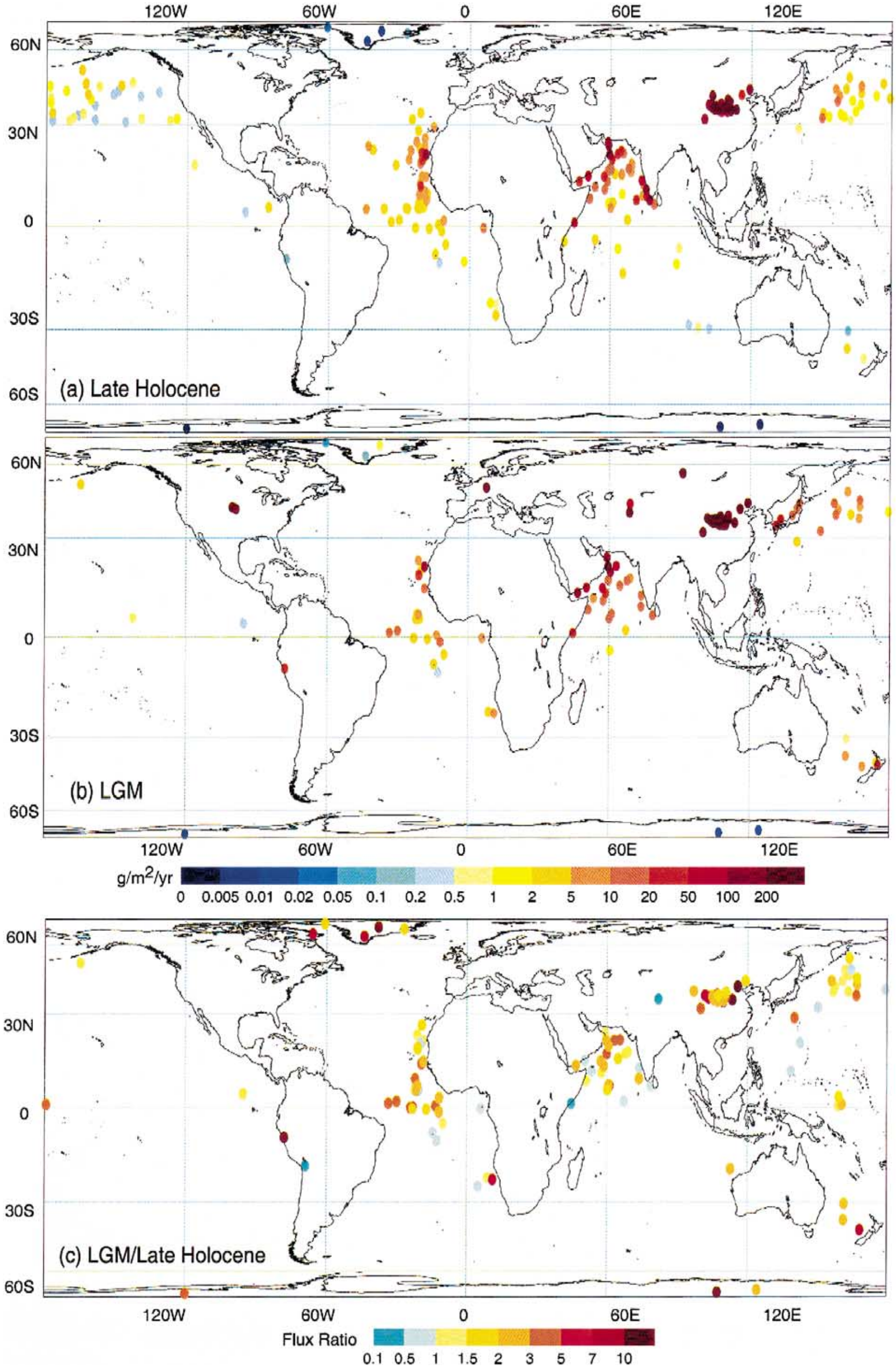


Fig. 8. (a) Late Holocene and (b) LGM dust deposition rates and (c) LGM/Late Holocene flux ratios derived from the DIRTMAP database.

creased entrainment of dust from the Persian Gulf as well as a weakening of the southwest monsoon during glacial summers. The magnitude of dust deposition in the Arabian Sea ranges from 2.7 to 141.6

g/m^2/year in the late Holocene sediments, with highest rates occurring in close proximity to the Arabian and Indian peninsulas. It is possible that these sites are affected by some hemipelagic contamination, although authors have attempted to eliminate fluvial contamination by (1) screening sediment cores for turbidites, and (2) using isotope tracers to identify the proportions of fluvial input to the remaining cores (Clemens and Prell, 1990; Sirocko et al., 1991).

6.2.3. Australia

The three dust records southeast of Australia suggest a 2–9-fold increase in dust deposition rates at the LGM (Hesse, 1994), and a 2-fold increase in quartz deposition rates are observed northwest of Australia in the Indian Ocean (Kolla and Biscaye, 1977). Deposition rates to the Tasman Sea are low (0.2–3 g/m^2/year for the late Holocene). An increase in the coarse fraction of terrigenous material to the sites in the Tasman Sea (Hesse and McTainsh, 1999) is coincident with the glacial increases in aeolian accumulation rates. These authors suggest the change in particle size may reflect a reduction in dust rainout events rather than an increase in wind speeds at the LGM, associated with a more arid glacial climate around Australia (Wasson, 1987).

6.2.4. Africa

Measured glacial dust deposition rates downwind of the Sahara/Sahel region are two to five times larger than late Holocene values, with modern dust deposition rates ranging from 10 to 100 g/m^2/year close to the continent, and values of 2–10 g/m^2/year in regions further downwind. South of the Sahara (i.e. south of the Equator into the Gulf of Guinea), dust deposition rates remain the same or were reduced slightly at the LGM compared with the late Holocene.

6.3. Terrestrial deposits

The DIRTMAP database currently contains only a limited subset of the available data from terrestrial deposits. Nevertheless, the data that have been included show some interesting features. Dust deposition rates are greatest in the LGM loess deposits near the source regions, ranging from approximately 100 g/m^2/year (SE Chinese Loess Plateau) to as much as 1000–4000 g/m^2/year (Late Wisconsinian Loess, Nebraska). High LGM loess deposition rates are also found in the periglacial regions of North America, Central Europe and Asia. LGM deposition rates on the Chinese Loess Plateau are highest in the northwest, decreasing by an order of magnitude to the southeast.

The record of massive loess deposition and palaeosol development on the Chinese Loess Plateau extends back ~2.5 Ma and, at lower rates, as far back as 6–7 Ma (Ding et al., 1998, 1999; Sun et al., 1998a,b). This loess–palaeosol succession provides a unique record of the cycles of dust deposition and Quaternary climate change, with cold, dry glacial periods generally corresponding with periods of high loess deposition. Conversely, the warm moist interglacial periods are marked by periods of soil formation across most of the plateau. In spite of the many studies of the mineralogy, grain size, magnetic characteristics, and sedimentology of loess deposits from the Chinese Loess Plateau, the age control on this region is still limited. Nevertheless, synthesis of these data as part of the DIRTMAP database provides a first assessment of changes in aeolian accumulation rates across the Loess Plateau for the last 130,000 (Fig. 9) (Sun et al., 2000).

Interpretation of Holocene dust accumulation rates on the Chinese Loess Plateau is complicated by the presence of extensive cultivation layers. For this discussion, we examine changes in the aeolian accumulation rates estimated for the last interglacial period (80–130 ka). Comparisons between accumulation rates for the present interglacial period (i.e. Marine Isotope Stage 1, 0–12 ka) and the last interglacial period suggest that, overall, accumulation rates for the current interglacial are greater, and that accumulation rates in the western part of the Loess Plateau are considerably larger (in some cases, an order of magnitude higher) than those recorded for the last interglacial. Interglacial accumulation rates (in particular those for the last interglacial) fall in the range of 20–500 g/m^2/year (Fig. 9a). In general, the range falls between 20 and 200 g/m^2/year, with the very highest accumulation rates occurring in the north and northwest regions of the Loess Plateau. During the LGM, aeolian accumulation rates were significantly higher, ranging from 50 to greater than

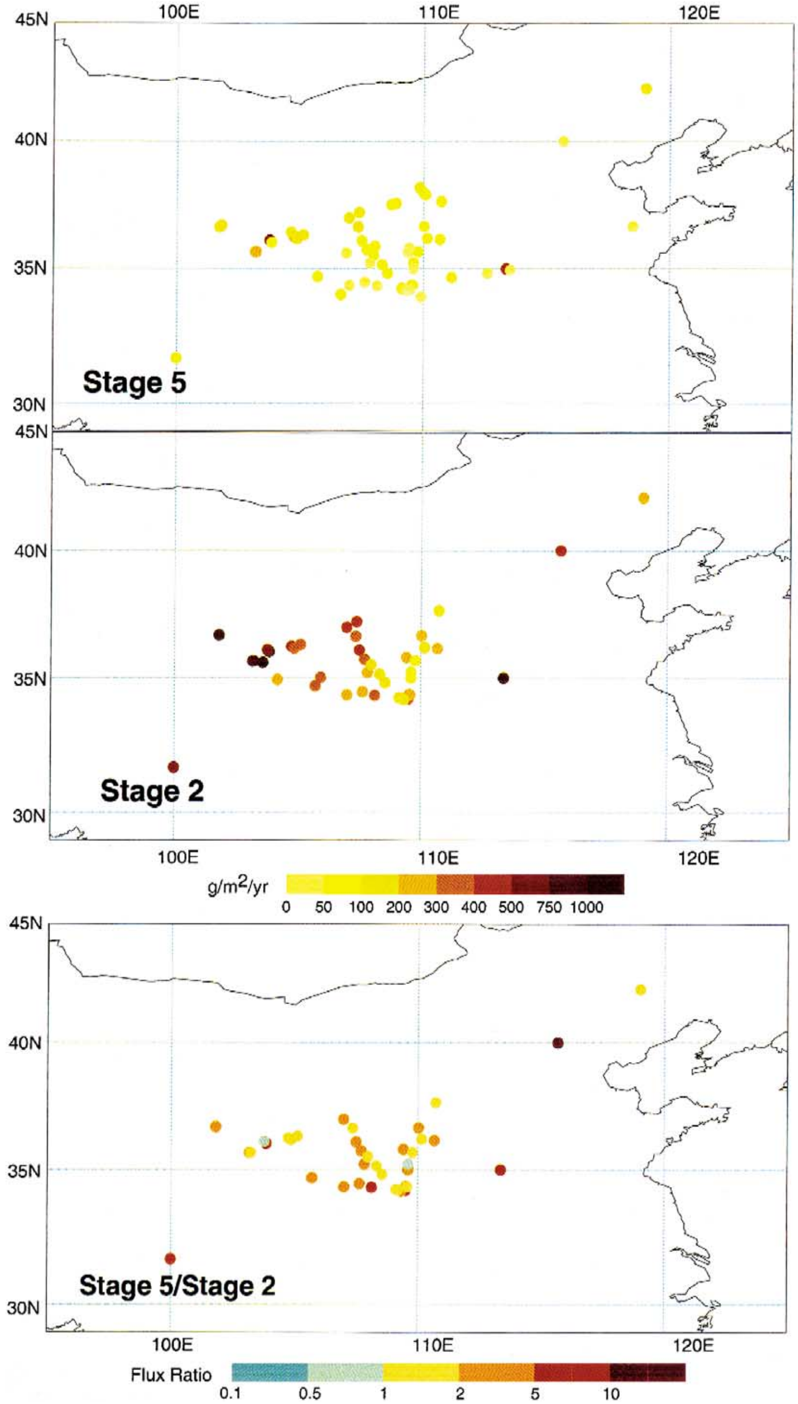


Fig. 9. (a) Stage 5 Interglacial (80–130 ka), (b) Stage 2 (15–24 ka) dust deposition rates and (c) LGM/Stage 5 ratios of dust flux ratios for the Chinese Loess Plateau compiled as part of the DIRTMAP data base (Sun et al., 2000).

1000 $g/m^2/year$, with the highest accumulation rates occurring mostly in the west–northwest parts of the Loess Plateau (Fig. 9b). Like the mid-continental USA, some sites on the Chinese Loess Plateau are affected by local deposition processes, e.g. by their proximity to river basins. Nevertheless, examination of the general glacial–interglacial changes suggests that the accumulation rates during the last glacial period were of the order of one to five times greater than those observed during interglacial periods (Fig. 9c).

7. The future of DIRTMAP

The DIRTMAP database has been used as a means of testing equilibrium simulations of the dust cycle for the LGM and under modern conditions (see e.g. Mahowald et al., 1999), but there are several potential avenues of research in dust modelling that require the use of reliable and flexible validation data sets. Here, we outline some research directions that could exploit the flexibility of the DIRTMAP database, and present ideas about the future expansion of DIRTMAP.

7.1. The future of DIRTMAP as a validation tool

The kinds of models that are currently used to simulate the role of dust in the climate system are relatively simple, operate at a coarse spatial scale (ca. 3–5°), and simulate changes in dust loading in response to the equilibrium climate of specfic intervals (e.g. the LGM). However, the continual development and incorporation of dust processes into earth system models means that processes on a wide range of temporal and spatial scales may soon be included in both regional and global simulations of the dust cycle. In making comparisons with these models, the user must be able to select data from the DIRTMAP data base at the appropriate temporal and spatial scales that reflect the processes incorporated in a given model. DIRTMAP must be flexible enough to remain a useful tool for model validation as earth system models progressively improve.

The coarse spatial resolution of current earth system models means that they cannot be expected to capture local processes of dust generation, e.g. the preferential accumulation of dust downwind of major rivers and the generation of periglacial loess (see e.g. Fig. 5). Furthermore, because these models have been used to simulate the response of dust to equilibrium climate states, they cannot be expected to capture high-resolution dust events as seen, e.g. in the ice cores (Fig. 3) or in well-dated loess deposits. Simulating either short-term variability of this sort or the longer-term (glacial–interglacial) evolution of the dust cycle will require the use of fully coupled earth system models in which ocean processes, vegetation and dust itself are included as dynamic components of the system. Unfortunately, although such models are being developed, they are computationally heavy and the necessity for long (100 to several 1000 years) runs will increase this cost.

Computationally efficient models of intermediate complexity (e.g. Gallée et al., 1991) can be used for longer simulations of the climate system, and thus could provide a useful exploratory tool. Although these models have low-resolution representations of both continents and ocean basins, and so cannot be expected to capture regional details of dust deposition, they could be used to simulate changes in dust deposition across climate transitions, e.g. from glacial to interglacial periods. Evaluation of such simulations, using data from the DIRTMAP data base, is possible, but some thought is required about pretreatment of the data. Zonal averages of the dust deposition to the Atlantic and Pacific Oceans (Fig. 10) capture the general features of the modern pattern of atmospheric dust loadings, and show how this pattern changed under the different climate conditions of the LGM. However, some information is apparently lost in this presentation of the data, e.g. the spatial patterns of decreased dust deposition downwind of source regions. Such complications may be circumvented by careful selection of the regions over which zonal averages are taken. For example, one might choose to divide the Pacific Basin into two halves in order to isolate the influence of continental sources on either sides of the basin. Similar averages of dust deposition might be made over land masses. Special care must be taken in creating regional or zonal averages, as these analyses can be highly sensitive to the selection of regional boundaries (see e.g. Qin et al., 1998; Kohfeld and Harrison, 2000).

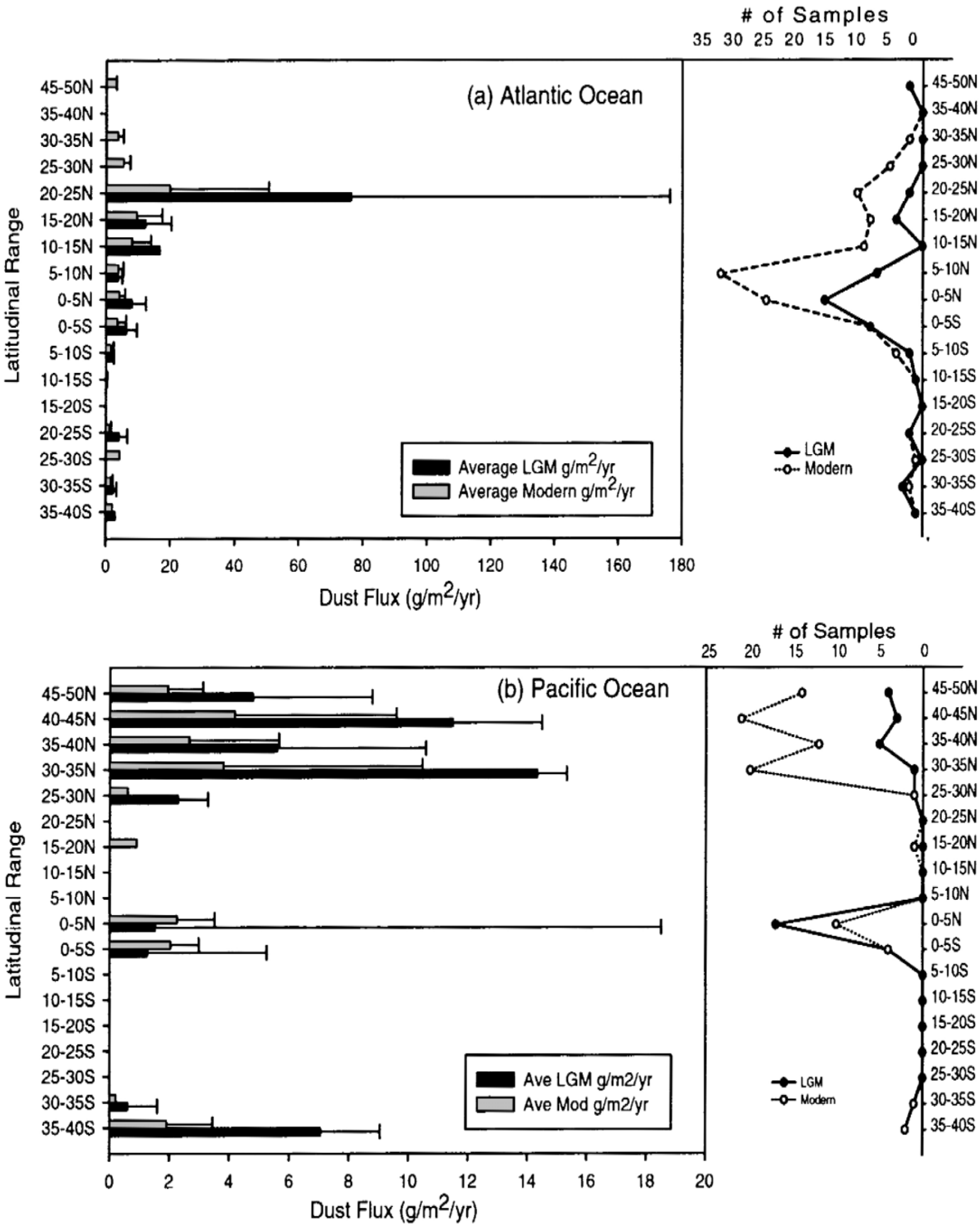


Fig. 10. Dust accumulation rates for the LGM (black bars) and Holocene (grey bars) periods, averaged zonally over 5° latitudinal bands from the (a) Atlantic and (b) Pacific ocean basins. Error bars represent 1 standard deviation. Number of samples contributing to each zonal dust flux estimate is also summarized for the LGM (black dots) and Holocene (open circles) periods.

The DIRTMAP data base could also be used to improve the model-simulated fields of dust delivered to the surface ocean, for the purpose of testing the impact of dust on marine productivity and ultimately on atmospheric CO_2. Marine biological productivity in several regions, notably the eastern equatorial Pacific and the Southern Ocean, has been shown through in situ experiments to be limited by the supply of the iron (Fe). Wind blown dust from the continents is the main source of Fe to the surface ocean. Ocean biogeochemistry models, designed to assess the extent to which this wind-blown iron can affect marine productivity and thereby the glacial–interglacial changes in atmospheric CO_2 concentrations, are still in the early stages of development. Marine records of dust deposition, as well as continental records of the fraction of Fe in potential source regions, could be used to constrain the amount, distribution and quantity of iron that enters the ocean.

Recent studies have demonstrated that correct characterisation of soil mineralogy (in particular, hematite content) and particle size from potential source regions is extremely important to the radiative properties of dust (Claquin et al., 1998, 1999). Thus, the mineralogical data from deposits that are both near potential source regions (loess deposits) as well as those recording long-distance transport (marine sediments and ice cores) might be used as a means of testing simulations of changes in the mineralogical composition of atmospheric dust.

7.2. Improvements in the primary data

The maps presented in Fig. 8 show there are large gaps in the spatial coverage of DIRTMAP. In the case of the ice core and marine records, these gaps represent places where new records are urgently required. In the case of terrestrial records, the gaps largely reflect the need for careful synthesis of existing data. However, it is clear that there are also some regions where additional field studies will be required to complete our knowledge of the spatial patterns of past dust deposition. In addition to syntheses of existing data, or collection of data from new sites, there are a number of other improvements that could be made to maximise the usefulness of existing records. These include the following.

- Improvement of high quality dating of dust records. Dating methods, particularly on loess deposits and marine sediments, are currently largely based on chronostratigraphic correlation and the assumptions inherent therein. Addressing questions of high-frequency changes requires continued development and application of both radiocarbon and luminescence dating techniques.
- Measurement and reporting of the bulk densities of aeolian sediments. Bulk densities of loess sediments are rarely measured and even less frequently reported. Furthermore, error estimates on bulk density measurements of dried marine sediment cores range from 10% to 25%.
- Inclusion of grain size data. In addition to providing an indication of changes in dust deposition processes in all three palaeoenvironments, particle size estimates are one of the key diagnostics for determining potential contamination, particularly in marine sediments.
- Advancement in techniques to isolate the aeolian component of sediments. In loess sediments, this involves using micromorphology and pedological techniques to isolate what portion of the sediment is reworked or formed in situ from actual airborne materials. In marine sediments, this involves the application of multiple techniques (e.g. grain size analysis, magnetic susceptibility, REE, tracer isotopes) to guarantee that materials are atmospherically derived.
- Improvement of modern and interglacial estimates of dust accumulation. Modern estimates of dust accumulation in loess regions are affected by human disturbance. Long-term trap measurements of dust accumulation that encompass periods of both high and low dust deposition could improve estimates of modern dust deposition rates. Measurements of last interglacial dust accumulation rates could provide an alternative baseline for comparison with glacial deposition rates.
- Intercalibration of dust accumulation rate estimates from different sensors and/or measurement techniques. Estimates of recent dust accumulation rates are actually based on information pertinent to very different averaging periods. Surface marine sediments frequently average the last 3–6 ka. Marine and terrestrial sediment trap estimates, on the other hand, integrate dust accumulation rates over time scales of the order of seasons to years. Satellite estimates (usually based on measurements of shortwave reflectance) provide somewhat indirect estimates of atmospheric loadings, integrated on time scales of days to years. Some attempts have been

made to calibrate these different types of modern measurements over selected regions (e.g. Sirocko and Sarnthein, 1989; Ratmeyer et al., 1999), but a global effort to address this issue is still lacking.

· Incorporation of continuous time series records of dust properties. Some time series data are included in the DIRTMAP database, but the focus of the database has been to provide estimates of dust deposition for key time periods. Incorporation of more time-series data will permit future comparisons with transient simulations of the dust cycle, e.g. across a complete glacial–interglacial cycle.

The DIRTMAP database was originally developed as a validation data set for earth system model simulations of the glacial–interglacial dust cycle. However, the DIRTMAP database contains a wide variety of dust data and is a useful repository that could serve the earth science community for many purposes beyond model validation. The full participation of the earth science community is required for DIRTMAP to achieve its fullest potential as a validation data set and a useful data repository.

Acknowledgements

This paper is a contribution from the MAGIC (Mineral Aerosol and Glacial–Interglacial Cycles) project, supported by grants from the Swedish Natural Science Research Council (NFR) and the US National Science Foundation (NSF). The members of MAGIC include: Y. Balkanski, G. Bergametti, T. Claquin, M. Hansson, S.P. Harrison, K.E. Kohfeld, N. Mahowald, I.C. Prentice, H. Rodhe, C. Roelandt, M. Schulz, and M. Werner. The DIRTMAP data base (http://www.bgc-jena.mpg.de/bgc_prentice) is endorsed and supported by the INQUA Loess Commission, the IGBP/GAIM Paleo Trace Gas and Mineral Aerosol Challenge (TRACES), the IGBP/PAGES Paleo Mapping Project (PMAP), and the International Geological Correlation Programme (IGCP) #413. The MPI-BGC provides funding support for the DIRTMAP database. The members of the DIRTMAP Steering Committee are E. Derbyshire, S.P. Harrison, D. Muhs, A. Wintle, and L. Zhou. J.M. Sun assisted in the compilation of data from the Chinese Loess Plateau. We thank Ed Derbyshire for his invitation to present DIRTMAP to the international loess community at the Lössfest Meeting, K.K. Andersen for helpful discussions, and I. Tegen, D. Muhs and P. Biscaye for helpful reviews of an earlier version of this paper.

References

Aleinikoff, J.N., Muhs, D., Sauer, R.R., Fanning, C.M., 1999. Late Quaternary loess in northeastern Colorado: II. Pb isotopic evidence for the variability of loess sources. Geol. Soc. Am. Bull. 111 (12), 1876–1883.

Alley, R.B., 2000. The Younger Dryas cold interval as viewed from central Greenland. Quat. Sci. Rev. 19 (1–5), 213–226.

Alley, R.B., Finkel, R.C., Nishizumi, K., Anankakrishnan, S., Shuman, C.A., Mershon, G., Zielinski, G.A., Mayewski, P.A., 1995. Changes in continental and sea-salt atmospheric loadings in central Greenland during the most recent deglaciation: model-based estimates. J. Glaciol. 41 (139), 503–514.

Alley, R.B., Shuman, C.A., Meese, D.A., Gow, A.J., Taylor, K.C., Cuffey, K.M., Fitzpatrick, J.J., Grootes, P.M., Zielinski, G.A., Ram, M., Spinelli, G., Elder, B., 1997. Visual-stratigraphic dating of the GISP2 ice core: basis, reproducibility, and application. J. Geophys. Res. 102, 26367–26381.

Andersen, K.K., Armengaud, A., Genthon, C., 1998. Atmospheric dust under glacial and interglacial conditions. Geophys. Res. Lett. 25 (13), 2281–2284.

Andreae, M.O., 1995. Climatic effects of changing atmospheric aerosol levels. In: Henderson-Sellers, A. (Ed.), World Survey of Climatology: Future Climates of the World.

Arnold, E., Leinen, M., King, J., 1995. Paleoenvironmental variation based on the mineralogy and rock-magnetic properties of sediment from Sites 885 and 886. In: Rea, D.K., Basov, I.A., Scholl, D.W., Allan, J.F. (Eds.), Proceedings of the Ocean Drilling Program, Scientific Results. Ocean Drilling Program, College Station, TX, pp. 231–245.

Arnold, E., Merrill, J., Leinen, M., King, J., 1998. The effect of source area and atmospheric transport on mineral aerosol collected over the North Pacific Ocean. Global Planet. Change 18, 137–159.

Basile, I., Grousset, F.E., Revel, M., Petit, J.R., Biscaye, P.E., Barkov, N.I., 1997. Patagonian origin of glacial dust deposited in East Antarctica (Vostok and Dome C) during glacial stages 2, 4 and 6. Earth Planet. Sci. Lett. 146, 573–589.

Beget, J., 1990. Middle Wisconsin climate fluctuations recorded in central Alaskan loess. Geogr. Phys. Quat. 44 (1), 3–13.

Biscaye, P.E., 1965. Mineralogy and sedimentation of recent deep-sea clay in the Atlantic Ocean and adjacent seas and oceans. Geol. Soc. Am. Bull. 76, 732–803.

Biscaye, P., Eittreim, S., 1977. Suspended particulate loads and transports in the nepheloid layer of the abyssal Atlantic Ocean. Mar. Geol. 23, 155–172.

Biscaye, P.E., Chesselet, R., Prospero, J.M., 1974. Rb–r, 87Sr/86Sr Isotope system as an index of provenance of continental dusts in the open Atlantic Ocean. J. Rech. Atmos. 8 (3–4), 819–829.

Biscaye, P.E., Grousset, F.E., Revel, M., Gaast, S.V.D., Zielinski, G.A., Vaars, A., Kukla, G., 1997. Asian provenance of glacial dust (Stage 2) in the GISP2 ice core, Summit, Greenland. J. Geophys. Res. 102 (C12), 26315–26886.

Blunier, T., Schwander, J., Stauffer, B., Stocker, T., Dallenbach, A., Indermuhle, A., Tschumi, J., Chappellaz, J., Raynaud, D., Barnola, J.M., 1997. Timing of the Antarctic cold reversal and the atmospheric CO_2 increase with respect to the Younger Dryas event. Geophys. Res. Lett. 24 (21), 2683–2686.

Broecker, W.S., Peng, T.-H., 1982. Tracers in the Sea. Lamont-Doherty Geological Observatory, Palisades, NY, 690 pp.

Busacca, A.J., McDonald, E.V., 1994. Regional sedimentation of Late Quaternary loess on the Columbia Plateau: sediment source areas and loess distribution patterns. Wash. Div. Geol. Earth Resour. Bull. 80, 181–190.

Catubig, N.R., Archer, D.E., Francois, R., deMenocal, P., Howard, W., Yu, E.-F., 1998. Global deep-sea burial rate of calcium carbonate during the last glacial maximum. Paleoceanography 13 (3), 298–310.

Chen, F.H., Bloemendal, J., Wang, J.M., Li, J.J., Oldfield, F., 1997. High-resolution multi-proxy climate records from Chinese loess: evidence for rapid climatic changes over the last 75 kyr. Palaeogeogr., Palaeoclimatol., Palaeoecol. 130, 323–335.

Chlachula, J., Evans, M.E., Rutter, N.W., 1998. A magnetic investigation of a Late Quaternary loess/palaeosol record in Siberia. Geophys. J. Int. 132, 128–132.

Claquin, T., Schulz, M., Balkanski, Y., Boucher, O., 1998. Uncertainties in assessing radiative forcing by mineral dust. Tellus Ser. B, Chem. Phys. Meteorol. 50 (5), 491–505.

Claquin, T., Schulz, M., Balkanski, Y., 1999. Modeling the mineralogy of atmospheric dust sources. J. Geophys. Res. 104, 22243–22256.

Clemens, S.C., 1998. Dust response to seasonal atmospheric forcing: proxy evaluation and calibration. Paleoceanography 13 (5), 471–490.

Clemens, S.C., Prell, W.L., 1990. Late Pleistocene variability of Arabian Sea summer monsoon winds and continental aridity: eolian records from the lithogenic component of deep-sea sediments. Paleoceanography 5 (2), 109–145.

CLIMAP, P.M., 1981. Seasonal reconstructions of the Earth's surface at the last glacial maximum. Geological Society of America Map and Chart Series, MC-36.

Cragin, J.H., Herron Jr., M.M., Klouda, C.C.L., Klouda, G., 1977. Interhemispheric comparison of changes in the composition of atmospheric precipitation during the late Cenozoic era. In: Dunbar, M.J. (Ed.), Polar Oceans, Proceedings of the Polar Oceans Conference. Arctic Institute of North America, Calgary, Alberta, pp. 617–641.

Damuth, J.E., Jacobi, R.D., Hayes, D.E., 1983. Sedimentation processes in the Northwest Pacific Basin revealed by echo character mapping studies. Geol. Soc. Am. Bull. 94, 381–395.

Dansgaard, W., White, J.W.C., Johnsen, S.J., 1989. The abrupt termination of the Younger Dryas climate event. Nature 339, 532–533.

Dansgaard, W., Johnsen, S.J., Clausen, H.B., Dahl-Jensen, D., Gundestrup, N.S., Hammer, C.U., Hvidberg, C.S., Steffensen, J.P., Sveinbjörnsdóttir, A.E., Jouzel, J., Bond, G.C., 1993. Evidence for general instability of past climate from a 250 kyr ice-core record. Nature 264, 218–220.

Davidson, C.I., Bergin, M.H., Kuhns, H.D., 1996. The deposition of particles and gases to ice sheets. In: Wolff, E.W., Bales, R.C. (Eds.), Chemical Exchange Between the Atmosphere and Polar Snow. NATO ASI Series, Springer-Verlag, New York, pp. 275–306.

De Angelis, M., Legrand, M., Petit, J.R., Barkov, N.I., Korotkevitch, Y.S., Kotlyakov, V.M., 1984. Soluble and insoluble impurities along the 950 m deep Vostok ice core (Antarctica)-climate implications. J. Atmos. Chem. 1, 215–239.

Delmas, R.J., Legrand, M., 1989. Long term changes in the concentrations of major chemical compounds (soluble and insoluble) along deep ice cores. In: Oeschger, H., Oeschger Jr., C.C.L. (Eds.), Dahlem Konferenzen: The Environmental Record in Glaciers and Ice Sheets. Wiley, New York, pp. 319–341.

Denne, J.E., Luza, K.V., Richmond, G.M., Jensen, K.M., Fishman, W.D., Wermund, G., 1993. Quaternary Geologic Map of the Wichita 4°×6° Quadrangle, United States. In: Richmond, G.M., Christiansen, A.C. (Eds.), Quaternary Geologic Atlas of the United States, Miscellaneous Investigations Series. US Geological Survey, Denver, CO.

Dentener, F.J., Carmichael, G.R., Zhang, Y., Lelieveld, J., Crutzen, P.J., 1996. Role of mineral aerosol as a reactive surface in the global troposphere. J. Geophys. Res. 101 (D17), 22869–22889.

DePaolo, D.J., Manton, W.I., Grew, E.S., Halpern, M., 1982. Sm–Nd, Rb–Sr, U–Th–Pb systematics of granulit facies rocks from Fyfe Hills, Enderby Land, Antarctica. Nature 298, 614–618.

Derbyshire, E., Kemp, R., Meng, X., 1995. Variations in loess and palaeosol properties as indicators of palaeoclimatic gradients across the loess plateau of North China. Quat. Sci. Rev. 14, 681–697.

Ding, Z., Yu, Z., Rutter, N.W., Liu, T., 1994. Towards an orbital time scale for Chinese loess deposits. Quat. Sci. Rev. 13, 39–70.

Ding, Z., Sun, J., Liu, T., Zhu, R.X., Yang, S.L., Guo, B., 1998. Wind-blown origin of the Pliocene red clay formation in the Chinese Loess Plateau. Earth Planet. Sci. Lett. 161, 135–143.

Ding, Z., Xiong, S., Sun, J., Yang, S.L., Gu, Z.Y., Liu, T.S., 1999. Pedostratigraphy and paleomagnetism of a ∼7.0 Ma eolian loess-red clay sequence at Lingtai, Loess Plateau, north-central China, and the implications for paleomonsoon evolution. Palaeogeogr., Palaeoclimatol., Palaeoecol. 152, 49–66.

Duce, R.A., 1995. Source, distributions, and fluxes of mineral aerosols and their relationship to climates. In: Charlson, R.J., Heintzenberg, J. (Eds.), Dahlem Workshop on Aerosol Forcing of Climate. Wiley, New York, pp. 43–72.

Eden, D.N., Qizhong, W., Hunt, J.L., Whitton, J.S., 1994. Mineralogical and geochemical trends across the Loess Plateau, North China. Catena 21, 73–90.

Fisher, D.A., 1979. Comparison of 105 years of oxygen isotope

and insoluble impurity profiles from the Devon Island and Camp Century ice cores. Quat. Res. 11, 299–305.
Frazee, C.J., Fehrenbacher, J.B., Krumbein, W.C., 1970. Loess distribution from a source. Soil Sci. Soc. Am. Proc. 34, 296–301.
Fuhrer, K., Neftel, A., Anklin, M., Maggi, V., 1993. Continuous measurements of hydrogen peroxide, formaldehyde, calcium and ammonium concentrations along the new GRIP ice core from Summit, central Greenland. Atmos. Environ. 27A (12), 1873–1880.
Fuhrer, K., Neftel, A., Anklin, M., Staffelbach, T., Legrand, M., 1996. High-resolution ammonium ice core record covering a complete glacial–interglacial cycle. J. Geophys. Res. Atmos. 101 (D2), 4147–4164.
Fuhrer, K., Wolff, E.W., Johnsen, S.J., 1999. Timescales for dust variability in the Greenland Ice Core Project (GRIP) ice core in the last 100,000 years. J. Geophys. Res. 104 (D24), 31043–31052.
Gallée, H., Ypersele, J.P.v., Fichefet, T., Tricot, C., Berger, A., 1991. Simulation of the last glacial by a coupled, sectorially averaged climate-ice sheet model: 1. The climate model. J. Geophys. Res. 96 (D7), 13139–13161.
Gaudichet, A., Petit, J.R., Lefevre, R., Lorius, C., 1986. An investigation by analytical transmission electron microscopy of individual insoluble microparticles from Antarctic (Dome C) ice core samples. Tellus 38B, 250–261.
Gaudichet, A., Angelis, M.D., Lefevre, R., Petit, J.R., Korotkevitch, Y.S., Petrov, V.N., 1988. Mineralogy of insoluble particles in the Vostok Antarctic ice core over the last climatic cycle (150-ka). Geophys. Res. Lett. 15 (13), 1471–1474.
Gaudichet, A., Angelis, M.D., Joussaume, S., Petit, J.R., Korotkevitch, Y.S., Petrov, V.N., 1992. Comments on the origin of dust in East Antarctica for present and ice age conditions. J. Atmos. Chem. 14, 129–142.
Genthon, C., 1992. Simulations of desert dust and sea-salt aerosols in Antarctica with a general circulation model of the atmosphere. Tellus Ser. B 44, 371–389.
Gillette Jr., D.A., Fryrear, I.H.B., Fryrear, D.W., 1974. The influence of wind velocity on the size distribution of aerosols generated by the wind erosion of soils. J. Geophys. Res. 79, 4068–4075.
Giorgi, F., 1988. Dry deposition velocities of atmospheric aerosols as inferred by applying a particle dry deposition paramterization to a general circulation model. Tellus 40B, 23–41.
Goldberg, E.D., Griffin, J.J., 1970. The sediments of the northern Indian Ocean. Deep-Sea Res. 17, 513–537.
Goldstein, S., O'Nions, R.K., Hamilton, P.J., 1984. A Sm–Nd isotopic study of atmospheric dusts and particulates from major river sytems. Earth Planet. Sci. Lett. 70, 221–236.
Gray, H.H., Bleuer, N.K., Lineback, J.A., Swadley, W.C., Richmond, G.M., Miller, R.A., Goldthwait, R.P., Ward, R.A., 1991. Quaternary Geologic Map of the Louisville $4° \times 6°$ Quadrangle, United States. In: Richmond, G.M., Fullerton, D.S. (Eds.), Quaternary Geologic Atlas of the United States, Miscellaneous Investigations Series. US Geological Survey, Denver, CO.
Grousset, F.E., Biscaye, P.E., Zindler, A., Prospero, J., Chester, R., 1988. Neodymium isotopes as tracers in marine sediments and aerosols: North Atlantic. Earth Planet. Sci. Lett. 87, 367–378.
Grousset, F.E., Biscaye, P.E., Ravel, M., Petit, J.-R., Pye, K., Joussaume, S., Jouzel, J., 1992a. Antarctic (Dome C) ice-core dust at 18 k.y. B. P.: isotopic contraints on origins. Earth Planet. Sci. Lett. 111, 175–182.
Grousset, F.E., Rognon, P., Coudegaussen, G., Pedemay, P., 1992b. Origins of Peri-Saharan dust deposits traced by their Nd and Sr isotopic composition. Palaeogeogr., Palaeoclimatol., Palaeoecol. 93 (3–4), 203–212.
Grousset, F.E., Parra, M., Bory, A., Martinez, P., Bertrand, P., Shimmield, G., Ellam, R.M., 1998. Saharan wind regimes traced by Sr–Nd isotopic composition of subtropical Atlantic sediments: Last Glacial Maximum vs. today. Quat. Sci. Rev. 17 (4–5), 395–409.
Hall, I.R., McCave, I.N., 1998. Late Glacial to Recent accumulation fluxes of sediments at the shelf edge and slope of NW Europe, 48–50 * N. In: Stoker, M.S., Evans, D., Cramp, A. (Eds.), Geological Processes on Continental Margins: Sedimentation, Mass-Wasting and Stability. Geological Society of London Special Publications, London, pp. 339–350.
Hallberg, G.R., Lineback, J.A., Mickelson, D.M., Knox, J.C., Goebel, J.E., Hobbs, H.C., Whitfield, J.W., Ward, R.A., Boellstorf, J.D., Swinehart, J.B., Dreeszen, V.H., 1991. Quaternary Geologic Map of the Des Moines $4° \times 6°$ Quadrangle, United States. In: Richmond, G.M., Fullerton, D.S., Christiansen, A.C. (Eds.), Quaternary Geologic Atlas of the United States, Miscellaneous Investigations Series. US Geological Survey, Denver, CO.
Hammer, C.U., 1977. Dust studies on Greenland ice cores, isotopes and impurities in snow and ice-symposium. Proceedings of the Grenoble Symposium, pp. 365–370.
Hammer, C.U., Clausen, H.B., Dansgaard, W., Neftel, A., Kristinsdottor, P., Johnson, E., 1985. Continuous impurity analysis along the Dye 3 deep core. In: Langway, J.C.C., Oeschger, H., Dansgaard, W. (Eds.), Greenland Ice Core: Geophysics, Geochemistry, and the Environment. Geophys. Monogr., AGU, Washington, DC, pp. 90–94.
Handy, R.L., 1976. Loess distribution by variable winds. Geol. Soc. Am. Bull. 87, 915–927.
Hansson, M., 1993. Detecting Changes in Climate and Atmospheric Composition with Tracers in Arctic Ice Caps. Stockholm University, Stockholm, Sweden.
Hansson, M.E., 1994. The Renland ice core. A northern hemisphere record of aerosol composition over 120,000 years. Tellus 46B, 390–418.
Harrison, S.P., Jolly, D., Laarif, F., Abe-Ouchi, A., Dong, B., Herterich, K., Hewitt, C., Joussaume, S., Kutzbach, J.E., Mitchell, J., Noblet, N.D., Valdes, P., 1998. Intercomparison of simulated global vegetation distributions in response to 6 kyr B.P. orbital forcing. J. Clim. 11, 2721–2742.
Harrison, S.P., Kohfeld, K.E., Roelandt, C., Claquin, T., in press. The role of dust in climate changes today, at the last glacial maximum and in the future. Earth Sci. Rev.

Hesse, P.P., 1994. The record of continental dust from Australia in Tasman Sea sediments. Quat. Sci. Rev. 13, 257–272.

Hesse, P.P., McTainsh, G.H., 1999. Last Glacial Maximum to early Holocene wind strength in the mid-latitudes of the southern hemisphere from aeolian dust in the Tasman Sea. Quat. Res. 52, 343–349.

Holbrook, D.F., Gilliland, W.A., Luza, K.V., Pope, D.E., Wermund, E.G., Miller, R.A., Bush, W.V., Jensen, K.N., Fishman, W.D., 1990. Quaternary Geologic Map of the Vicksburg 4°×6° Quadrangle, United States. In: Richmond, G.M., Fullerton, D.S., Weide, D.L. (Eds.), Quaternary Geologic Atlas of the United States, Miscellaneous Investigations Series. US Geological Survey, Denver, CO.

Hovan, S.A., Rea, D.K., Pisias, N.G., 1991. Late Pleistocene continental climate and oceanic variability recorded in Northwest Pacific sediments. Paleoceanography 6 (3), 349–370.

Imbrie, J., Hays, J.D., Martinson, D.G., McIntyre, A., Mix, A.C., Morley, J.J., Pisias, N.G., Prell, W.L., Shackleton, N.J., 1984. The orbital theory of Pleistocene climate: support from a revised chronology of the marine d18O record. In: Berger, A., Imbrie, J., Hays, J., Kukla, G., Saltzman, B. (Eds.), Milankovitch and Climate Part 1. NATO ASI Series C: Mathematical and Physical Sciences. Reidel, Dordrecht, pp. 269–305.

Janacek, T.R., 1984. Eolian sedimentation in the northwest Pacific Ocean: a preliminary examination of the data from Deep Sea Drilling Sites 576 and 578. In: L.H.B.e.a. G.R. Heath (Ed.), Init. Reports, DSDP. US Govt. Printing Office, Washington, DC, pp. 589–603.

Johnson, L.R., 1979. Particle size fractionation of eolian dusts during transport and sampling. Mar. Geol. 21, M17–M21.

Joseph, L.H., Rea, D.K., Pluijm, B.A.v.d., 1998. Use of grain size and magnetic fabric analyses to distinguish among depositional environments. Paleoceanography 13 (5), 491–501.

Joussaume, S., Taylor, K.E., 1995. Status of the Paleoclimate Modeling Intercomparison Project (PMIP). In: Gates, W.L. (Ed.), Proceedings of the First International AMIP Scientific Conference, 15–19 May 1995, Monterey, CA. 532 pp.

Joussaume, S., Taylor, K.E., Braconnot, P., Mitchell, J.F.B., Kutzbach, J., Harrison, S.P., Prentice, I.C., Broccoli, A.J., Abe-Ouchi, A., Bartlein, P.J., Bonfils, C., Dong, B., Guiot, J., Herterich, K., Hewitt, C.D., Jolly, D., Kim, J.W., Kislov, A., Kitoh, A., Loutre, M.F., Masson, V., McAvaney, B., McFarlane, N., deNoblet, N., Peltier, W.R., Peterschmitt, J.Y., Pollard, D., Rind, D., Royer, J.F., Schlesinger, M.E., Syktus, J., Thompson, S., Valdes, P., Vettoretti, G., Webb, R.S., Wyputta, U., 1999. Monsoon changes for 6000 years ago: results of 18 simulations from the Paleoclimate Modeling Intercomparison Project (PMIP). Geophys. Res. Lett. 26, 859–862.

Jouzel, J., Barkov, N.I., Barnola, J.M., Bender, M., Chappellaz, J., Benthon, C., Kotlyakov, V.M., Lorius, C., Petit, J.R., Raynaud, D., 1993. Extending the Vostok ice-core record of paleoclimate to the penultimate glacial period. Nature 364 (6436), 407–412.

Kageyama, M., D'Andrea, F., Ramstein, G., Valdes, P.J., Vautard, R., 1999. Weather regimes in past climate atmospheric general circulation model simulations. Clim. Dyn. 15 (10), 773–793.

Kemp, R.A., Derbyshire, E., Xingmin, M., Fahu, C., Baotian, P., 1995. Pedosedimentary reconstruction of a thick loess–paleosol sequence near Lanzhou in North-Central China. Quat. Res. 43, 30–45.

King, M.D., Kaufman, Y.J., Tanré, D., Nakajima, T., 1999. Remote sensing of tropospheric aerosols from space: past, present, and future. Bull. Am. Meteorol. Soc. 80 (11), 2229–2258.

Kohfeld, K.E., Harrison, S.P., 2000. How well can we simulate past climates? Evaluating the models using global palaeoenvironmental datasets. Quat. Sci. Rev. 19 (1–5), 321–347.

Kolla, V., Biscaye, P.E., 1977. Distribution and origin of quartz in the sediments of the Indian Ocean. J. Sediment. Petrol. 47 (2), 642–649.

Kolla, V., Henderson, L., Biscaye, P.E., 1976. Clay mineralogy and sedimentation in the western Indian Ocean. Deep-Sea Res. 23, 949–961.

Kolla, V., Kostecki, J.A., Robinson, F., Biscaye, P.E., 1981. Distributions and origins of clay minerals and quartz in surface sediments of the Arabian Sea. J. Sediment. Petrol. 51 (2), 563–569.

Kukla, G., 1987. Loess stratigraphy in Central China. Quat. Sci. Rev. 6, 191–219.

Kutzbach, J.E., Webb, T., III, 1993. Conceptual basis for understanding Late-Quaternary climates. In: H.E.W. Jr., J.E. Kutzbach, T.W. III, Ruddiman, W.F., Street-Perrott, F.A., Bartlein, P.J. (Eds.), Global Climates since the Last Glacial Maximum. University of Minnesota Press, Minneapolis, pp. 5–11.

Kutzbach, J.E., Guetter, P.J., Ruddiman, W.F., Prell, W.L., 1989. Sensitivity of climate to late Cenozoic uplift in Southern Asia and the American West - numerical experiments. Journal of Geophysical Research-Atmospheres 94, 18393–18407.

Lacis, A.A., Mishenko, M.I., 1995. Climate forcing, climate sensitivity, and climate response: a radiative modeling perspective on atmospheric aerosols. In: Charlson, R.J., Heintzenberg, J. (Eds.), Aerosol Forcing of Climate. Wiley, London, pp. 11–42.

Legrand, M., Mayewski, P.A., 1997. Glaciochemistry of polar ice cores: a review. Rev. Geophys. 35 (3), 219–243.

Legrand, M., DeAngelis, M., Maupetit, F., 1993. Field investigation of major and minor ions along Summit (Central Greenland) ice cores by ion chromatography. J. Chromatogr. 460, 251–258.

Leinen, M., 1989. The late Quaternary record of atmospheric transport to the northwest Pacific from Asia. In: Leinen, M., Sarnthein, M. (Eds.), Paleoclimatology and Paleometeorology: Modern and Past Patterns of Global Atmospheric Transport. NATO ASI Series C: Mathematical and Physical Sciences, Kluwer Academic Publishing, Dordrecht, pp. 693–732.

Leinen, M., Prospero, J.M., Arnold, E., Blank, M., 1994. Mineralogy of aeolian dust reaching the North Pacific Ocean 1. Sampling and analysis. J. Geophys. Res. 99 (D10), 21017–21023.

Li-Jones, X., Prospero, J.M., 1998. Variations in the size distribution of non-sea-salt sulfate aerosol in the marine boundary layer at Barbados: impact of African dust. J. Geophys. Res. 103, 16073–16084.

Lineback, J.A., Bleuer, N.K., Mickelson, D.M., Farrand, W.R., Goldthwait, R.P., 1983. Quaternary Geologic Map of the Chicago 4°×6° Quadrangle, United States. In: Richmond, G.M., Fullerton, D.S. (Eds.), Quaternary Geologic Atlas of the United States, Miscellaneous Investigations Series. US Geological Survey, Denver, CO.

Liu, T.S., 1966. Composition and Texture of Loess (in Chinese). Science Press, Beijing, 132 pp.

Liu, T.S. et al., 1985. Loess and the Environment. China Ocean Press, Beijing, 251 pp.

Lorius, C., Jouzel, J., Ritz, C., Merlivat, L., Barkov, N.I., Korotkevich, Y.S., Kotlyakov, V.M., 1985. A 150,000-year climatic record from Antarctic ice. Nature 316 (6029), 591–596.

Maggi, V., 1997. Mineralogy of atmospheric microparticles deposited along the Greenland Ice Core Project ice core. J. Geophys. Res. 102 (C12), 26725–26734.

Maher, B.A., Thompson, R., 1991. Mineral magnetic record of the Chinese loess and paleosols. Geology 19, 3–6.

Mahowald, N., Kohfeld, K.E., Hansson, M., Balkanski, Y., Harrison, S.P., Prentice, I.C., Schulz, M., Rodhe, H., 1999. Dust sources and deposition during the Last Glacial maximum and current climate: a comparison of model results with palaeodata from ice cores and marine sediments. J. Geophys. Res. 104 (D13), 15895–15916.

Manighetti, B., McCave, I.N., 1995. Late glacial and Holocene palaeocurrents around Rockall Bank, NE Atlantic Ocean. Paleoceanography 10 (3), 611–626.

Martinson, D.G., Pisias, N.G., Hays, J.D., Imbrie, J., Moore, T.C., Shackleton, N.J., 1987. Age dating and the orbital theory of the ice ages: development of a high-resolution 0 to 300,000-year chronostratigraphy. Quat. Res. 27, 1–29.

Mayewski, P.A., Meeker, L.D., Whitlow, S., Twickler, M.S., Morrison, M.C., Bloomfield, P., Bond, G.C., Alley, R.B., Gow, A.J., Grootes, P.M., Meese, D.A., Ram, M., Taylor, K.C., Wumkes, W., 1994. Changes in atmospheric circulation and ocean ice cover over the North Atlantic during the last 41,000 years. Science 263, 1747–1751.

Mayewski, P.A., Meeker, L.D., Twickler, M.S., Whitlow, S., Yang, Q., Lyons, W.B., Prentice, M., 1997. Major features and forcing of high-latitude northern hemisphere atmospheric circulation using a 110,000-year-long glaciochemical series. J. Geophys. Res. 102 (C12), 26345–36366.

McCave, I.N., 1986. Local and global aspects of the bottom nepheloid layers in the world ocean. Neth. J. Sea Res. 20 (2/3), 167–181.

McCave, I.N., Manighetti, B., Robinson, S.G., 1995. Sortable silt and fine sediment size/composition slicing: parameters for palaeocurrent speed and palaeoceanography. Paleoceanography 10 (3), 593–610.

Miller, R.L., Tegen, I., 1998. Climate response to soil dust aerosols. J. Clim. 11, 3247–3267.

Miller, R.L., Tegen, I., 1999. Radiative forcing of a tropical direct circulation by soil dust aerosols. J. Atmos. Sci. 56, 2403–2433.

Miller, R.A., Maher, S.W., Copeland, C.W., Rheams, J., LeNeathery, K.F., Gilliland, T., Friddell, W.A., Nostrand, M.S., Wheeler, A.K.V., Holbrook, W.H., Bush, D.F., 1988. Quaternary Geologic Map of the Lookout Mountain 4°×6° Quadrangle, United States. In: Richmond, G.M., Fullerton, D.S. (Eds.), Quaternary Geologic Atlas of the United States, Miscellaneous Investigations Series. US Geological Survey, Denver, CO.

Mosley-Thompson, E., Thompson, L.G., Grootes, P., Gundestrup, N., 1990. Little Ice Age (Neoglacial) paleoenvironmental conditions at Siple Station, Antarctica. Ann. Glaciol. 14, 199–204.

Muhs, D.R., Aleinikoff, J.N., Stafford Jr., T.W., Kihl, R., Been, J., Mahan, S.A., 1999. Late Quaternary loess in northeastern Colorado: I. Age and paleoclimatic significance. Geol. Soc. Am. Bull. 111 (12), 1861–1875.

Nakai, S.i., Halliday, A.N., Rea, D.K., 1993. Provenance of dust in the Pacific Ocean. Earth Planet. Sci. Lett. 119, 143–157.

Petit, J.R., Briat, M., Royer, A., 1981. Ice age aerosol content from East Antarctic ice core samples and past wind strength. Nature 293, 391–394.

Petit, J.R., Mounier, L., Jouzel, J., Korotkevich, Y.S., Kotlyakov, V.I., Lorius, C., 1990. Paleoclimatological and chronological implications of the Vostok core dust record. Nature 343, 56–58.

Petit, J.R., Jouzel, J., Raynaud, D., Barkov, N.I., Barnola, J.M., Basile, I., Bender, M., Chappellaz, J., Davis, M., Delaygue, G., Delmotte, M., Kotlyakov, V.M., Legrand, M., Lipenkov, V.Y., Lorius, C., Pepin, L., Ritz, C., Saltzman, E., Stievenard, M., 1999. Climate and atmospheric history of the past 420,000 years from the Vostok ice core, Antarctica. Nature 399 (6735), 429–436.

Pinot, S., Ramstein, G., Harrison, S.P., Prentice, I.C., Guiot, J., Stute, M., Joussaume, S., groups, P.p., 1999. Tropical paleoclimates at the Last Glacial Maximum: comparison of Paleoclimate Modeling Intercomparison Project (PMIP) simulations and paleodata. Clim. Dyn. 15 (11), 857–874.

Porter, S.C., Zhisheng, A., 1995. Correlation between climate events in the North Atlantic and China during the last glaciation. Nature 375, 305–308.

Prentice, I.C., Webb III, W., 1998. BIOME 6000: reconstructing global mid-Holocene vegetation patterns from palaeoecological records. J. Biogeogr. 25, 997–1005.

Pye, K., 1987. Aeolian Dust and Dust Deposits. Academic, San Diego, 334 pp.

Qin, B., Harrison, S.P., Kutzbach, J.E., 1998. Evaluation of modelled regional water balance using lake status data: a comparison of 6 ka simulations with the NCAR CCM. Quat. Sci. Rev. 17, 535–548.

Rahn, K.A., Borys, R.D., Shaw, G.E., Schütz, L., Jaenicke, R., 1979. Long-range impact of desert aerosol on atmospheric chemistry: two examples. In: Morales, C. (Ed.), Saharan Dust: Mobilization, Transport, and Deposition. Wiley, Chichester, pp. 243–266.

Ratmeyer, V., Fischer, G., Wefer, G., 1999. Lithogenic particle fluxes and grain size distributions in the deep ocean off NW Africa: implications for seasonal changes of aeolian dust input and downward transport. Deep-Sea Res. I 46, 1289–1337.

Ravelo, A.C., Fairbanks, R.G., Philander, S.G.H., 1990. Reconstructing tropical Atlantic hydrography using planktonic foraminifera and an ocean model. Paleoceanography 5 (3), 409–431.

Rea, D.K., 1994. The paleoclimatic record provided by eolian deposition in the deep sea: the geologic history of wind. Rev. Geophys. 32 (2), 159–195.

Rea, D.K., Hovan, S.A., 1995. Grain size distribution and depositional processes of the mineral component of abyssal sediments: Lessons from the North Pacific. Paleoceanography 10 (2), 251–258.

Rea, D.K., Leinen, M., 1988. Asian aridity and the zonal westerlies: Late Pleistocene and Holocene record of eolian deposition in the Northwest Pacific Ocean. Palaeogeogr., Palaeoclim., Palaeoecol. 66, 1–8.

Reader, M.C., Fung, I., McFarlane, N., 1999. The mineral dust aerosol cycle during the Last Glacial Maximum. J. Geophys. Res.-Atmos. 104 (D8), 9381–9398.

Royer, S., Angelis, M.D., Petit, J.R., 1983. A 30,000 year record of physical and optical properties of microparticles from an East Antarctic ice core and implications for paleoclimate reconstruction models. Clim. Change 5, 381–412.

Ruddiman, W.F., 1997. Tropical Atlantic terrigenous fluxes since 25,000 yrs B.P. Mar. Geol. 136, 189–207.

Ruddiman, W.F., McIntyre, A., Ruddiman, W.F.M.A., 1981. The North Atlantic Ocean during the last deglaciation. Palaeogeogr., Palaeoclim., Palaeoecol. 35 (2–4), 145–214.

Ruddiman, W.F., Sarnthein, M., Backman, J., Balduf, J.G., Curry, W., Dupont, L.M., Janacek, T., Pokras, E.M., Raymo, M.E., Stabell, B., Stein, R., Teidemann, R., 1989. Late Miocene to Pleistocene evolution of climate in Africa and low-latitude Atlantic: Overview of Leg 108 results. Proceeding of the Ocean Drilling Program, Scientific Results 108, 463–484.

Ruhe, R.V., 1983. Depositional environment of Late Wisconsin loess in the midcontinental United States. In: Wright, J.H.E. (Ed.), Late-Quaternary Environments of the United States. The Late Pleistocene, University of Minnesota Press, Minneapolis, pp. 130–137.

Salamatin, A.N., Lipinkov, V.Y., Barkov, N.I., Jouzel, J., Petit, J.R., Raynaud, D., 1998. Ice core age dating and paleotemperature calibration based on isotope and temperature profiles from deep boreholes at Vostok Station (East Antarctica). J. Geophys. Res.-Atmos. 103 (D8), 8963–8977.

Sarnthein, M., 1978. Sand desers during glacial maximum and climatic optimum. Nature 271 (5648), 43–46.

Sarnthein, M., Koopman, B., 1980. Late Quaternary deep-sea record on northwest African dust supply and wind circulation. Palaeoecol. Africa 12, 239–253.

Sarnthein, M., Thiede, J., Pflaumann, U., Erlenkeuser, H., Fuetterer, D., Koopmann, B., Lange, H., Siebold, E., 1982. Atmospheric and oceanic circulation patterns off Northwest Africa during the past 25 million years. In: U.V. Rad, K., Hinz, M., Sarnthein, E. (Eds.), Geology of the Northwest African Continental Margin. Springer-Verlag, Berlin, pp. 547–604.

Schultz, L., 1979. Sahara dust transport over the North Atlantic Ocean-model circulations and measurements. In: Morales, C. (Ed.), Saharan Dust: Mobilization, Transport, Deposition. Wiley, New York, pp. 267–277.

Shaw, G.E., 1989. Aerosol transport from sources to ice sheets. In: Oeschger, H., Oeschger Jr., C.C.L. (Eds.), Dahlem Konferenzen: The Environmental Record in Glaciers and Ice Sheets. Wiley, New York, pp. 13–27.

Shine, K.P., Foster, P.M., 1999. The effect of human activity on radiative forcing of climate change: a review of recent developments. Global Planet. Change 20, 205–225.

Sirocko, F., Lange, H., 1991. Clay-mineral accumulation rates in the Arabian Sea during the late Quaternary. Mar. Geol. 97, 105–119.

Sirocko, F., Sarnthein, M., 1989. Wind-borne deposits in the Northwestern Indian Ocean: Record of Holocene sediments versus modern satellite data. In: Leinen, M., Sarnthein, M. (Eds.), Paleoclimatology and Paleometeorology: Modern and Past Patterns of Global Atmospheric Transport. Kluwer Academic Publishing, Dordrecht, pp. 401–433.

Sirocko, F., Sarnthein, M., Lange, H., Erlenkeuser, H., 1991. Atmospheric summer circulation and coastal upwelling in the Arabian Sea during the Holocene and the Last Glaciation. Quat. Res. 36, 72–93.

Sokolik, I.N., Toon, O.B., 1996. Direct radiative forcing by anthropogenic airborne mineral aerosols. Nature 381, 681–683.

Sokolik, I.N., Toon, O.B., 1999. Incorporation of mineralogical composition into models of the radiative properties of mineral aerosol from UV to IR wavelengths. J. Geophys. Res. 104 (D8), 9423–9444.

Sowers, T., Bender, M., Labeyrie, L., Martinson, D., Jouzel, J., Raynaud, D., Pichon, J.J., Korotkevich, Y.S., 1993. A 135,000-year Vostok-SPECMAP common temporal framework. Paleoceanography 8 (6), 737–766.

Steffensen, J.P., 1988. Analysis of the seasonal variation in dust, Cl−, NO3−, and SO42− in two central Greenland firn cores. Ann. Glaciol. 10, 171–177.

Steffensen, J.P., 1995. Microparticles and chemical impurities in ice cores from Dye 3, South Greenland and their interpretation in palaeoclimatic reconstructions. PhD Thesis, University of Copenhagen, Copenhagen, Denmark, 106 pp.

Steffensen, J.P., 1997. The size distribution of microparticles from selected segments of the Greenland Ice Core Project ice core representing different climatic periods. J. Geophys. Res. 102, 26755–26763.

Steffensen, J.P., Clausen, H.B., Christensen, J.M., 1996. On the spatial variability of impurity content and stable isotopic composition in recent Summit snow. In: Wolff, E.W., Bales, R.C. (Eds.), Chemical Exchange between the Atmosphere and Polar Snow. Springer-Verlag, Berlin, pp. 607–615.

Steffensen, J.P., Clausen, H.B., Hammer, C.U., Legrand, M., Angelis, M.D., 1997. The chemical composition of cold events within the Eemian section of the Greenland Ice Core Project ice core from Summit, Greenland. J. Geophys. Res. 102 (C12), 26747–26754.

Street-Perrott, F.A., Marchand, D.S., Roberts, N., Harrison, S.P., 1989. Global lake-level variations from 18,000 to 0 years ago: a palaeoclimatic analysis. US DOE/ER/60304-H1 TR046. US/Department of Energy, Washington, DC.

Sun, D., An, Z., Shaw, J., Bloemendal, J., Sun, Y.B., 1998a. Magnetostratigraphy and paleoclimatic significance of late Tertiary aeolian sequences in the Chinese Loess Plateau. Geophys. J. Int. 134, 207–212.

Sun, D., Shaw, J., An, Z., Chen, M.Y., Yue, L.P., 1998b. Magnetostratigraphy and paleoclimatic interpretation of a continuous

7.2 Ma late Cenozoic eolian sediment from the Chinese Loess Plateau. Geophys. Res. Lett. 25, 85–88.

Sun, J.M., Kohfeld, K.E., Harrison, S.P., 2000. Records of aeolian dust deposition on the Chinese Loess Plateau during the Late Quaternary. Technical Reports - Max-Planck-Institute für Biogeochemie 1, pp. 318.

Svensson, A., Biscaye, P.E., Grousset, F.E., 2000. Characterization of late glacial continental dust in the Greenland Ice Core Project ice core. J. Geophys. Res. 105 (D4), 4637–4656.

Swap, R., Garstang, M., Greco, S., Talbot, R., Kallberg, P., 1992. Saharan dust in the Amazon Basin. Tellus Ser.-B l44B (2), 133–149.

Swinehart, J.B., Dreeszen, V.H., Richmond, G.M., Tipton, M.J., Bretz, R., Steece, F.V., Hallberg, G.R., Geobel, J.E., 1994. Quaternary Geologic Map of the Platte River $4° \times 6°$ Quadrangle, United States. In: Richmond, G.M. (Ed.), Quaternary Geologic Atlas of the United States, Miscellaneous Investigations Series. US Geological Survey, Denver, CO.

Taylor, S.R., McClennan, S.M., 1985. The Continental Crust: Its Composition and Evolution. Blackwell, Cambridge, MA, 312 pp.

Taylor, K., Alley, R., Fiacco, J., Grootes, P., Lamorey, G., Mayewski, P., Spencer, M.J., 1992. Ice-core dating and chemistry by direct-current electrical-conductivity. J. Glaciol. 38 (130), 325–332.

Taylor, K.C., Hammer, C.U., Alley, R.B., Clausen, H.B., Dahl-Jensen, D., Gow, A.J., Gundestrup, N.S., Kipfstuhl, J., Moore, J.C., Waddington, E.D., 1993. Electrical conductivity measurements from the GISP2 and GRIP Greenland ice cores. Nature 366, 549–552.

Tegen, I., Fung, I., 1995. Contribution to the atmospheric mineral aerosol load from land surface modification. J. Geophys. Res. 100 (D9), 18707–18726.

Tegen, I., Lacis, A.A., 1996. Modeling of particle size distribution and its influence on the radiative properties of mineral dust aerosol. J. Geophys. Res. 101 (D14), 19237–19244.

Tegen, I., Miller, R., 1998. A general circulation model study on the interannual variability of soil dust aerosol. J. Geophys. Res. 103 (D20), 25975–25995.

Tegen, I., Lacis, A., Fung, I., 1996. The influence on climate forcing of mineral aerosols from disturbed soils. Nature 380, 419–422.

Thompson, L.G., 1977. Variations in microparticle concentration, size distribution and elemental composition fround in Camp Century, Greenland and Byrd station, Antarctica, deep ice cores. Proceedings of the Grenoble Symposium on Isotopes and Impurities in Snow and Ice. IAHS-AISH, pp. 351–364.

Thompson, L.G., 2000. Ice core evidence for climate change in the Tropics: implications for our future. Quat. Sci. Rev. 19 (1–5), 19–35.

Thompson, L.G., Mosley-Thompson, E., 1981. Microparticle concentration variations linked with climatic change: evidence from polar ice. Science 212, 812–815.

Thompson, L.G., Mosley-Thompson, E., 1989. One-half millennia of tropical climate variability as recorded in the stratigraphy of the Quelccaya Ice Cap, Peru. Geophysical Monograph, vol. 55, American Geophysical Union, Washington, DC.

Thompson, L.G., Mosley-Thompson, E., Davis, M.E., Bolzan, J.F., Dai, J., Yao, T., Gundestrup, N., Wu, X., Klein, L., Xie, Z., 1989. Holocene–Late Pleistocene climatic ice core records from Qinghai–Tibetan plateau. Science 246, 474–477.

Thompson, L.G., Mosley-Thompson, E., Davis, M.E., Lin, P.-N., Henderson, K.A., Cole-Dai, J., Bolzan, J.F., Liu, K.-b., 1995. Late Glacial Stage and Holocene tropical ice core records from Huascaràn, Peru. Science 269, 46–50.

Thompson, L.G., Yao, T., Davis, M.E., Henderson, K.A., Mosley-Thompson, E., Lin, P.N., Beer, J., Synal, H.A., Cole-Dai, J., Bolzan, J.F., 1997. Tropical climate instability: the last glacial cycle from a Qinghai–Tibetan ice core. Science 276 (5320), 1821–1825.

Thompson, L.G., Davis, M.E., Mosley-Thompson, E., Sowers, T.A., Henderson, K.A., Zagorodnov, V.S., Lin, P.-N., Mikhalenko, V.N., Campen, R.K., Bolzan, J.F., Cole-Dai, J., Francou, B., 1998. A 25,000-year tropical climate history from Bolivian ice cores. Science 282, 1858–1864.

Thorp, J., Smith, H.T.U., 1952. Pleistocene eolian deposits of the United States, Alaska, and Parts of Canada: National Research Council Committee for the Study of Eolian Deposts. Geological Society of America, New York.

Unruh, A., O'Nions, R.K., Zhu, X., Zhou, L., Dodonov, A., 1999. A provenance study of Eurasian loess using lead isotopes. In: Derbyshire, E. (Ed.), Loessfest '99. Loess: Characterization, Stratigraphy, Climate and Societal Significance, Royal Holloway University of London Center for Quaternary Research, Bonn, Germany, pp. 238–240.

Vlag, P.A., Oches, E.A., Banerjee, S.K., Solheid, P.A., 1999. The paleoenvironmental-magnetic record of the Gold Hill Steps loess section in central Alaska. Phys. Chem. Earth, Part A-Solid Earth Geodesy 24 (9), 779–783.

Wasson, R.J., 1987. Desert Dune Building, Dust Raising and Palaeoclimate in the Southern Hemisphere During the Last 280,000 Years, CLIMANZ 3. CSIRO, Division of Water Resources, Melbourne, pp. 123–137.

Wefers, M., Jaenicke, R., 1990. Global 3D distribution of desert aerosols from a numerical simulation. In: Masuda, S., Takahasi, K. (Eds.), 3rd Internationl Aerosol Conference, Kyoto, Japan, pp. 1086–1089.

Whitfield, J.W., Ward, R.A., Denne, J.E., Holbrook, D.F., Bush, W.V., Lineback, J.A., Luza, K.V., Jensen, K.M., Fishman, W.D., 1993. Quaternary Geologic Map of the Ozark Plateau $4° \times 6°$ Quadrangle, United States. In: Richmond, G.M., Weide, D.L. (Eds.), Quaternary Geologic Atlas of the United States, Miscellaneous Investigations Series. US Geological Survey, Denver, CO.

Whitlow, S., Mayewski, P.A., Dibb, J.E., 1992. A comparison of major chemical species input timing and accumulation at South Pole and Summit Greenland. Atmos. Environ., Part A 26 (11), 2045–2054.

Zdanowicz, C.M., Zielinski, G.A., Wake, C.P., Fisher, D.A., Koerner, R.M., 2000. A Holocene record of atmospheric dust deposition on the Penny Ice Cap, Baffin Island, Canada. Quat. Res. 53, 62–69.

Zhang, Y., Carmichael, G.R., 1999. The role of mineral aerosol in tropospheric chemistry in East Asia—a model study. J. Appl. Meteorol. 38, 353–366.

Zhang, X.Y., Arimoto, R., An, Z.S., 1997. Dust emission from

Chinese desert sources linked to variations in atmospheric circulation. J. Geophys. Res. 102 (D23), 28041–28047.
Zhang, X.Y., Arimoto, R., An, Z.S., 1999. Glacial and interglacial patterns for Asian dust transport. Quat. Sci. Rev. 18 (6), 811–819.
Zhou, L.P., Oldfield, F., Wintle, A.G., Robinson, S.G., Wang, J.T., 1990. Partly pedogenic origin of magnetic variations in Chinese loess. Nature 346, 737–739.
Zielinski, G.A., Mershon, G.R., 1997. Paleoenvironmental implications of the insoluble microparticle record in the GISP2 (Greenland ice core during the rapidly changing climate of the Pleistocene–Holocene transition). Geol. Soc. Am. Bull. 109, 547–559.