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The role of dust in climate changes today, at the last glacial maximum and in the future

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Abstract

Natural mineral aerosol (dust) is an active component of the climate system and plays multiple roles in mediating physical and biogeochemical exchanges between the atmosphere, land surface and ocean. Changes in the amount of dust in the atmosphere are caused both by changes in climate (precipitation, wind strength, regional moisture balance) and changes in the extent of dust sources caused by either anthropogenic or climatically induced changes in vegetation cover. Models of the global dust cycle take into account the physical controls on dust deflation from prescribed source areas (based largely on soil wetness and vegetation cover thresholds), dust transport within the atmospheric column, and dust deposition through sedimentation and scavenging by precipitation. These models successfully reproduce the first-order spatial and temporal patterns in atmospheric dust loading under modern conditions.

Atmospheric dust loading was as much as an order-of-magnitude larger than today during the last glacial maximum (LGM). While the observed increase in emissions from northern Africa can be explained solely in terms of climate changes (colder, drier and windier glacial climates), increased emissions from other regions appear to have been largely a response to climatically induced changes in vegetation cover and hence in the extent of dust source areas. Model experiments suggest that the increased dust loading in tropical regions had an effect on radiative forcing comparable to that of low glacial CO_2 levels.

Changes in land-use are already increasing the dust loading of the atmosphere. However, simulations show that anthropogenically forced climate changes substantially reduce the extent and productivity of natural dust sources. Positive feedbacks initiated by a reduction of dust emissions from natural source areas on both radiative forcing and atmospheric CO_2 could substantially mitigate the impacts of land-use changes, and need to be considered in climate change assessments.

Keywords: Dust cycle; Dust modelling; Radiative forcing; Biogeochemical cycles; Land-surface conditions; Last Glacial Maximum climates; Future climate changes

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1. Introduction

The presence of mineral or soil dust aerosol in the atmosphere has been documented for centuries (e.g. Dobson, 1781; Darwin, 1846) and the effects of dust on local climate has long been the subject of anecdotes and speculation (Zhang, 1982, 1983, 1984). However, the recognition that dust can play multiple roles in mediating physical and biogeochemical exchanges among the atmosphere, land and ocean, and thus is an active component of the global climate system, has come only in the last two decades (see, e.g. Shine and Forster, 1999). This belated discovery has coincided with the remarkable finding that the amount of dust reaching the polar ice sheets has varied by more than an order of magnitude on glacial–interglacial times scales (Thompson and Mosley-Thompson, 1981; Petit et al., 1990, 1999; Taylor et al., 1993a). In addition, there is evidence that changes in land use are causing a substantial increase in the amount of dust in the atmosphere today (UNEP, 1992; Tegen and Fung, 1995; Tegen et al., 1996; Sokolik and Toon, 1996). Thus, dust could have played an important role in past climate changes and may contribute to the course of climate changes in the future.

Our understanding of the role of dust in the climate system is still rudimentary, however, and not as well quantified as we would like. A deeper understanding of the relative magnitude of the various potential effects of dust will be dependent on the development and application of global models of the dust cycle and its interaction with other components of the Earth system. Although there has been much work on modelling dust deflation, transport and deposition at local scales (e.g. Gillette and Passi, 1988; Marticorena and Bergametti, 1995; Marticorena et al., 1997a,b; Shao et al., 1996), backed by extensive observational programmes, modelling of the global dust cycle is still relatively crude (see, e.g. the discussion in Tegen and Fung, 1994). Nevertheless, there has been exciting progress in recent years and it is timely to review the current state of knowledge.

In this review, aimed at a general Earth science and Quaternary science readership, we (a) summarise what is known in a qualitative sense about the role of dust in the climate system; (b) describe the processes controlling the deflation, transport and deposition of dust today, emphasising those aspects that are crucial in terms of modelling the global dust cycle; (c) discuss the possible role of dust in past climates, particularly during the Last Glacial Maximum (LGM) because this has been a focus for both climate and dust cycle modelling; and (d) speculate about the possible roles of dust in future climate changes. We conclude by suggesting some directions for research engaging Quaternary scientists as well as geophysical and biogeochemical modellers.

2. Observations of the modern dust cycle

Mineral dust aerosols are entrained into the atmosphere by aeolian deflation of surface material in areas of sparse vegetation. About 30% of the total continental land area (50×10^6 km^2) is a potential source for dust today (Sokolik and Toon, 1996). The global annual input of mineral aerosols to the atmosphere is estimated to be between 1 and 2 Pg year^{-1}, i.e. about half of the total production of tropospheric aerosols from both natural and artificial sources (d'Almeida, 1989; IPCC, 1994; Andreae, 1995; Duce, 1995). It is difficult to estimate the relative contribution of natural and anthropogenically derived dust to the global dust budget. Estimates based on the area affected by desertification caused by human-induced soil degradation (e.g. UNEP, 1992) suggest that anthropogenic dust sources represent ca. 20–30% of the total (Sokolik and Toon, 1996). However, an estimate of the area of dust production caused by human activities during the past 50 years, using inverse modelling, suggests that the relative size of the source may have been smaller (ca. 14%) although the effectiveness of these source areas as emitters of dust is likely to be greater than natural source areas (Tegen and Fung, 1995; Tegen et al., 1996). Despite the uncertainties, it is clear that human activities are significantly increasing the atmospheric burden of dust through wind erosion of agricultural land, salinization, overgrazing, deforestation and urbanization (UNEP, 1992; Sheehy, 1992).

The natural sources of mineral dust today are mainly in the arid and semi-arid tropics and subtropics (Fig. 1), with the most important source regions being the Sahara–Sahel in northern Africa and the Gobi–Taklamakan in central Asia (Middleton, 1991;

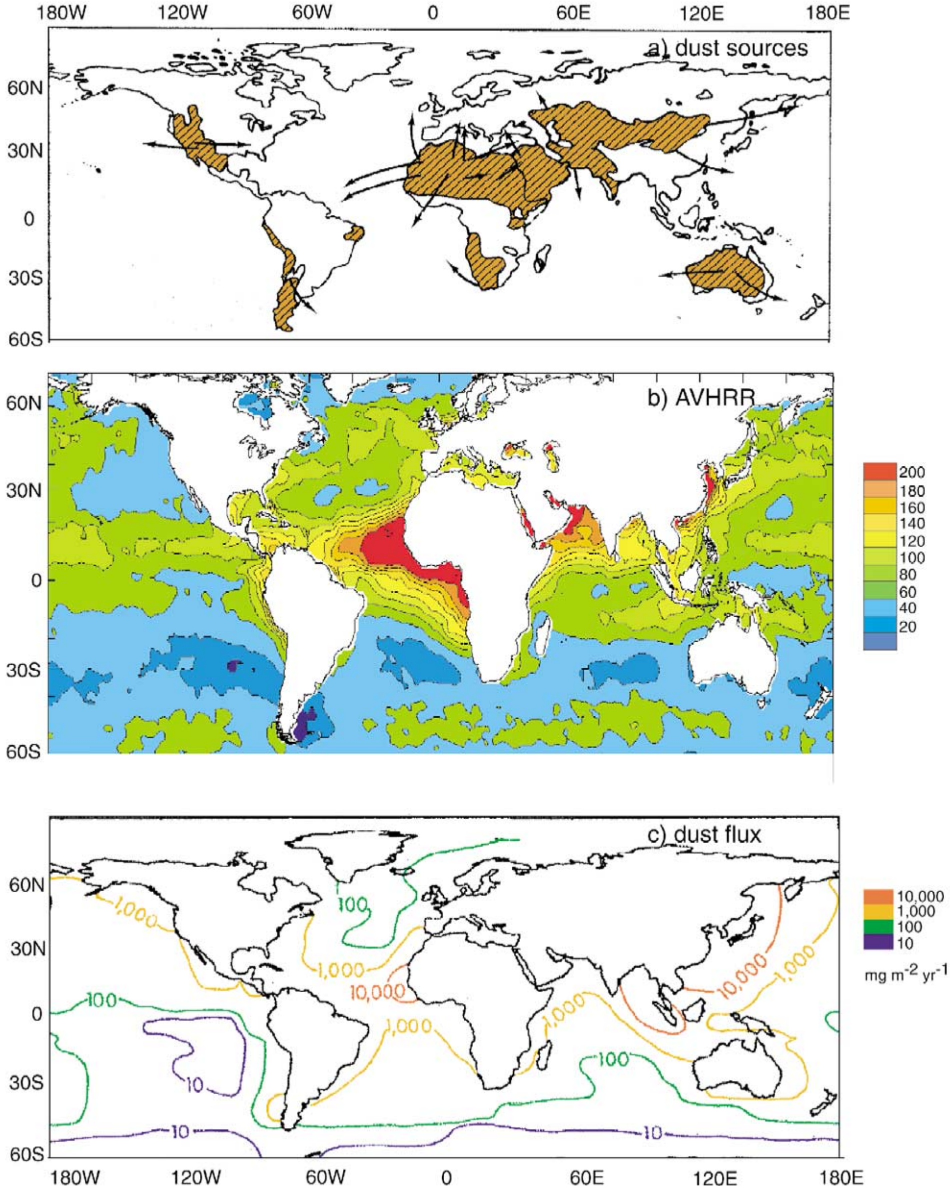

Fig. 1. The modern location of dust sources, transport paths and deposition zones. (a) Modern dust source regions and wind trajectories reconstructed from observations of dust storms (after Livingstone and Warren, 1996); (b) zones of high atmospheric dust concentrations, inferred from mean annual equivalent aerosol optical depth ($\times 1000$) as measured by AVHRR (from Husar et al., 1997); (c) global fluxes (mg m^{-2} $year^{-1}$) of mineral aerosols to the ocean (after Duce et al., 1991).

Middleton et al., 1986; Bergametti, 1992). The amount of dust transported from the Sahara has been estimated at 0.6–0.7 Pg $year^{-1}$ (d'Almeida, 1989), of which ca. 0.22 Pg is deposited in the North Atlantic (Duce et al., 1991), while Asian dust transport into the North Pacific has been conservatively estimated at ca. 0.006–0.012 Pg $year^{-1}$ (Uematsu et al., 1983). Asian dust transport to the China Sea alone has been estimated as 0.067 Pg $year^{-1}$ (Gao et al., 1997); this input represents between 20% and

70% of the total mineral input (depending on the year) to the more limited region of the Yellow Sea (Gao et al., 1992). Prospero et al. (1989) have estimated the dust flux to the western North Pacific at 0.3 Pg year^{-1}. The areas actively contributing dust to the atmosphere within these two regions are limited: possibly as little as 10% and certainly less than 50% of the Sahara contributes to annual emissions under modern climate (d'Almeida, 1986; Westphal et al., 1988; Legrand, 1990; Bergametti, 1992). Other, less important, source areas for dust include the Middle East (Middleton, 1986a,b; Pease et al., 1998), the northern part of the Indian subcontinent (Middleton, 1986b), southern South Africa (Prospero, 1981; Prospero et al., 1981), the interior basins of the southwestern USA and the southern High Plains (Orgill and Sehmel, 1976; Lee and Tchakerian, 1995; Bach et al., 1996), southern South America (Ares, 1994; Buschiazzo et al., 1999) and central Australia (Middleton, 1984; McTainsh and Pitblado, 1987; Shao and Leslie, 1997). Dust production and deflation can also occur under periglacial conditions in the high latitudes (Pye, 1995; Landvik, 1998), but in today's climate the areas affected are of limited extent and do not contribute significantly to the total global dust budget.

Atmospheric dust loadings and transport pathways have been reconstructed using data from various remote-sensing instruments, including the Advanced Very High Resolution Radiometer (AVHRR, e.g. Husar et al., 1997), the thematic ozone mapping spectrometer (TOMS, e.g. Herman et al., 1997), the coastal zone colour scanner (CZCS, e.g. Stegmann and Tindale, 1999) and, at a regional scale, METEOSAT visible channel images (e.g. Dulac et al., 1992; Moulin et al., 1998). There are problems is quantifying dust in the atmosphere (both in terms of optical thickness and dust loading) from satellite retrievals, which reflect uncertainties in the assumptions made about particle size, refractive index and particle shape (see King et al., 1999). Nevertheless, satellite retrievals can be helpful in reconstructing the general magnitude and the spatial/temporal patterns of dust loading. Thus, large-scale dust plumes

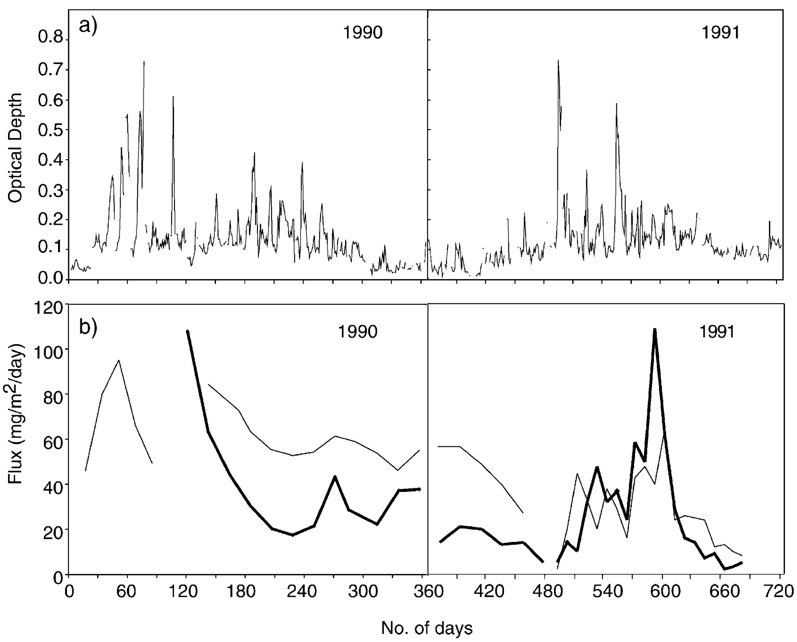

Fig. 2. Temporal characteristics of dust cycle in the subtropical North Atlantic, off of northeast Africa. Time series of (a) optical depth measurements from weekly/monthly AVHRR data (Balkanski, personal communication, 1998), and (b) lithogenic fluxes to marine sediment traps (Site CB: 21°N, 20°W; Wefer and Fischer, 1993; Ratmeyer et al., 1999). Traps were deployed at 730 (fine line) and 3557 m (bold line).

can be observed offshore from the main dust source regions, i.e. off of northwest Africa, over the Arabian Sea, over the northwest Pacific, and off the east coast of North America. The atmospheric dust loading of the northern hemisphere is about double that of the southern hemisphere, because of the greater extent of northern hemisphere sources.

Satellite data and meteorological observations indicate that there is a strong seasonal cycle in dust production and transport, with a spring (Asia) or summer (Africa) maximum in both emissions and atmospheric dust loadings. However, within this framework, dust emissions are characteristically sporadic (Fig. 2). This is because the entrainment of dust into the atmosphere is strongly affected by individual meteorological events. Dust emissions in Asia are often associated with the passage of cold fronts (e.g. Middleton, 1991; Zhou et al., 1994), while dust emissions from the Sahara–Sahel are often associated with strong convective disturbances (Tetzlaff et al., 1989) or large-scale frontal systems (Dayan et al., 1991). As a result of seasonal and interannual changes in atmospheric circulation patterns, which in turn influence the frequency of dust-generating events, dust fluxes to the atmosphere show high temporal variability on timescales ranging from days to decades (e.g. Prospero and Nees, 1986; McTainsh et al., 1989; De Angelis and Gaudichet, 1991; Goudie and Middleton, 1992; Arimoto et al., 1995; Moulin et al., 1997; Derbyshire et al., 1998).

3. The role of dust in the climate system

3.1. The role of dust in energy exchange

Mineral dust modifies the radiation budget through absorption and scattering of both incoming solar radiation (insolation) and outgoing terrestrial radiation (see Lacis and Mishenko, 1995; Tegen and Lacis, 1996; Liao and Seinfeld, 1998). It was originally assumed that the net radiative effect of dust was to produce cooling at the land surface and warming at the altitude of the dust layers (Carlson and Benjamin, 1980; Penner et al., 1994). In reality, dust may cause either a warming or a cooling at the surface depending on (a) the characteristics of the dust, including its concentration, vertical distribution in the atmospheric column, particle size distribution and mineralogy, and (b) external variables including albedo and temperature of the underlying surface.

The characteristics of dust particles affect the radiation budget primarily through affecting the balance between scattering and absorption of energy. The concentration of dust in the atmosphere affects the balance such that an increase in dust concentration increases the amount of energy absorbed. Particle size also affects the balance between scattering and absorption of energy because smaller particles are more effective in scattering energy than larger particles (Waggoner et al., 1981; Tegen and Lacis, 1996; Liao and Seinfeld, 1998). The size range of the dust particles that are transported over long distances in the atmosphere is from 0.1 to 20 μm (Pye, 1987; Duce, 1995) with a mass median diameter of 1.5–3 μm (Bergametti and Dulac, 1999). However, close to source regions, the dust particles may be up to 50 μm in diameter (d'Almeida and Schütz, 1983; Pye, 1987; Duce, 1995). The effectiveness of the scattering of energy also depends on particle shape (particularly the degree of sphericity) and density.

Dust generally consists of a mixture of minerals, including, e.g. quartz, clay minerals, calcite, gypsum and iron oxides, each with its own characteristic optical properties (Sokolik and Toon, 1999), occurring either as individual mineral grains or as pure or mixed-mineral aggregates. The mineralogy (and hence the colour) of a dust parcel is characteristic of its geographical source, although there can be progressive modifications, due to the decrease in grain size accompanied by differential deposition of some mineral species, with increasing distance of transport (Rahn et al., 1979; Prospero et al., 1981; Leinen et al., 1994; Merrill et al., 1994; Pease et al., 1998; Sokolik and Toon, 1999). Saharan dust is richer in iron (and consequently darker in colour) than Asian dust. It is possible to distinguish Saharan dust from Sahelian dust by its mineralogy and colour (Chiapello et al., 1997) even after it has been transported across the Atlantic to Barbados (e.g. Carlson and Prospero, 1972; Caquineau et al., 1998). The mineralogy also has an impact on the degree to which the dust occurs as aggregates, and hence can affect the size distribution of dust parcels. For example, Saharan dust is less aggregated than dust from Australia

(Kiefert et al., 1996). Most modelling studies of the modern dust cycle have ignored the potential effects of differences in dust mineralogy because of difficulties in predicting the occurence and abundance of different mineral species (although see Claquin et al., 1999). Recent studies (e.g. Claquin et al., 1998; Sokolik and Toon, 1999; Sokolik et al., 1998; Miller and Tegen, 1999) show that this is an oversimplification that can have a substantial effect on the estimated radiation budget. The degree to which dust occurs as mixed-mineral aggregates will have an additional impact on the radiation budget, because aggregates tend to be larger than single grain dust.

The radiative effect of the presence of a dust layer in the atmosphere depends on the albedo and temperature of the underlying surface (Sokolik and Toon, 1996; Tegen et al., 1996; Miller and Tegen, 1998), modified by dynamical effects within the atmosphere (Cess et al., 1985; Miller and Tegen, 1998). Thus, in general, dust acts to warm the upper atmosphere and cool the surface over dark surfaces (e.g. ocean, forests). The converse is true over surfaces with high albedo, such as ice sheets, fresh snow and deserts. However, in regions subject to frequent deep convective events, when the atmosphere is well mixed and saturated, the presence of a relatively thick dust layer may have no impact on the surface temperature. Simulations of the modern dust cycle made by Miller and Tegen (1998), for example, show that high dust levels in summer over the Arabian Sea produced no change in surface temperatures, while the much lower dust loadings during the winter (when there is little or no convection) produced a ca. 1°C cooling.

The direct radiative effect of the modern atmospheric dust was estimated to be ca. -0.75 W m^2, close to that of anthropogenic sulphates (-1.1 W m^2) by Andreae (1995, 1996). However, this estimate is probably unrealistic because it was based simply on an estimate of the mean dust content of the atmosphere (derived from estimates of production rates and the use of a mean residence time), and a generic mass extinction coefficient. To estimate the direct radiative effect more realistically, one has to make geographically explicit calculations taking into account the high spatial variability characteristic of atmospheric dust concentrations combined with the large variations in surface albedo. An attempt to estimate the direct radiative forcing of dust, taking into account the spatial heterogeneity by using an atmospheric transport model, indicated that the global forcing at the top of the atmosphere is *positive* but relatively small: 0.14 W m^2, of which 0.09 W m^2 is due to dust from anthropogenically disturbed surfaces (Tegen et al., 1996). The radiative forcing at the surface was -1.1 W m^2, however, and the regional impact of dust on radiative forcing varied from $+5.5$ to -2.1 W m^2. The direct radiation effect is substantially larger close to the source areas in the tropics and subtropics (Tegen et al., 1996). Claquin et al. (submitted for publication) have obtained a similar result for the global radiative forcing at the surface. However, in their simulation, there was a larger range in the magnitude of the regional impacts (from circa $+15$ to circa -17 W m^{-2}).

The presence of dust in the atmosphere could have a further, indirect impact on the radiation budget because dust particles can act as cloud condensation nuclei (Twomey et al., 1984), particularly after within cloud processing (see, e.g. Wurzler et al., 2000). Increases in the number of cloud condensation nuclei (CCN) will, other things being equal, lead to a decrease in the size of the CCN which in turn will increase cloud reflectivity (Shine and Forster, 1999). However, a decrease in the size of individual CCN also reduces the chance of precipitation events and hence increases the lifetime of clouds and decreases wet deposition rates. Thus, changes in CCN abundance and size can have very different impacts on the aerosol loading of the atmosphere, such that the net impact of dust on clouds is unclear (Duce, 1995; Lohman and Feichter, 1997).

3.2. The role of dust in atmospheric chemistry

Modern dust is usually highly alkaline, because of the high carbonate content of most arid and semi-arid soils, and thus has a neutralizing effect on rainwater acidity (Loye-Pilot et al., 1986; Varma, 1989; De Angelis and Gaudichet, 1991; Losno et al., 1991; Avila et al., 1997). Thus, dust can substantially alter the acidity budget of atmospheric deposition. In some parts of the world, e.g. in the western United States, mineral dust has been shown to completely neutralise rainwater acidity (Gillette et al., 1992).

Recent observational (Oltmans and Levy, 1992) and modelling (Dentener et al., 1996; Zhang and

Carmichael, 1999) studies have shown that mineral dust has an important role in the tropospheric photochemical oxidant cycle, and specifically on the abundance of tropospheric ozone (O_3) and the hydroperoxyl radical (HO_2). Dentener et al. (1996) suggest that direct uptake of ozone by mineral dust could decrease ozone concentrations by up to 10% and of the hydroperoxyl radical by a similar amount. According to this modelling study, the effects are largely confined to regions close to the dust sources.

Mineral dust also affects the atmospheric nitrogen and sulphur cycles. Observations and laboratory studies show that considerable amounts of nitrate and sulphate can be adsorbed on the surface of dust particles (e.g. Prospero and Savoie, 1989; Savoie and Prospero, 1989; Savoie et al., 1989; Mamane and Gottlieb, 1992; Wu and Okada, 1994; Carmichael et al., 1996; Li-Jones and Prospero, 1998; Tabazadeh et al., 1998). Savoie et al. (1989) estimated that 60% of the non-seasalt sulphate measured in Barbados was adsorbed on dust particles. Model experiments suggest that between 10–50% of atmospheric sulphate and 40–80% of atmospheric nitrate may be associated with dust particles (Dentener et al., 1996). Nitrogen and sulphur are removed from the gas phase and deposited on the surface of dust particles through heterogeneous processes (including nucleation, condensation and coagulation). These processes, while affected by the pH of the atmospheric constituents (and hence the mineralogy of the dust), are most clearly controlled by the presence of moisture and the size of the dust particles. Dust particles are usually dry when first lofted into the atmosphere, but the moisture content of the dust cloud increases, and so dust particles become more effective at scavenging as transport distance increases. Small dust particles, typical of long-distance transported dust, are most effective at scavenging (Li-Jones and Prospero, 1998). Thus, in contrast to the impact of dust on atmospheric oxidants, the effectiveness of dust in scavenging nitrogen or sulphur from the atmosphere is greater further away from the major dust sources and is both spatially and temporally highly variable.

The removal of nitrogen (NO_2, HNO_2, HNO_3) and sulphur (SO_2) compounds from the atmosphere reduces the overall aerosol loading of the atmosphere, and consequently affects the radiative balance. Dentener et al. (1996) and Zhang and Carmichael (1999) have suggested that SO_2 adsorption by dust particles could reduce the radiative cooling effect of sulphate aerosols.

Both nitrogen and sulphur compounds are involved in tropospheric ozone photochemistry, and decreases in their concentrations through adsorption by dust could therefore indirectly reduce ozone concentrations. Although this effect appears to be smaller than that due to direct uptake of O_3 by dust at a local scale, it may be a more important control on global-scale ozone concentrations because of the presence and effectiveness of long-distance transported dust on nitrate- and sulphate-scavenging (Zhang and Carmichael, 1999). The potential for indirect effects of dust on the radiative balance of the atmosphere is further enhanced because the adsorption of sulphate and nitrate on dust particles enhances the ability of dust to act as CCN (e.g. Dentener et al., 1996; Levin et al., 1996).

3.3. The role of dust in biogeochemical cycles

An understanding of the role of dust in the climate system must include consideration of the potential involvement of dust in the major biogeochemical cycles. Specifically, if dust is a source of nutrients for marine and terrestrial ecosystems, changes in dust delivery through time would impact on productivity. Changes in biospheric productivity impact on atmospheric composition (most noticeably on CO_2 but also on other gases such as N_2O) which leads to further impacts on climate.

Mineralogical studies in regions downwind of major deserts have shown that mineral dust is a major component of ocean sediments (e.g. Chester et al., 1979; Duce et al., 1980; Sarnthein and Koopman, 1980; Leinen and Heath, 1981; Leinen et al., 1994; Prospero, 1981; Blank et al., 1985; Rea, 1994). In the central North Pacific, for example, between 75% and 98% of the ocean-floor sediments have been derived from aeolian material (Blank et al., 1985). Mineral dust may therefore provide an important source of nutrients (particularly Si, Fe and P) for marine ecosystems (see Duce et al., 1991; Sunda and Huntsman, 1997; Hutchins and Bruland, 1998). Duce and Tindale (1991) estimated that atmospheric inputs are three times more important than riverine inputs as a source of soluble iron in the North Pacific.

Open-ocean iron-enrichment experiments (Iron-EX-I and Iron-EX-2) have shown that phytoplankton production and biomass increase as a result of iron fertilisation under modern conditions (Kolber et al., 1994; Martin et al., 1994; Coale et al., 1996). The input of iron-rich dust to nutrient-rich, low-chlorophyll regions could be a factor contributing to the regulation of marine primary production and thus to the export of carbon to the deep ocean (Duce, 1986; Martin and Fitzwater, 1988; Martin et al., 1990; Duce and Tindale, 1991; Behrenfeld et al., 1996; Jickells et al., 1996; Watson, 1996). The export of carbon to the deep ocean is an important process for millenial-scale regulation of the CO_2 content of the atmosphere and may have been a contributory cause of low CO_2 at the LGM (Martin, 1990; Martin et al., 1994; Prentice and Sarnthein, 1993; Kumar et al., 1995; Broecker and Henderson, 1998; Falkowski et al., 1998). It has recently been suggested that the input of alkaline dust, by affecting the alkalinity of surface waters and enhancing biological productivity at a regional scale, could also have contributed to CO_2 drawdown at the LGM (Oba and Pedersen, 1999).

There is a considerable body of evidence indicating that aeolian dust is also an important component of soils in those regions downwind of the major modern source areas (Muhs et al., 1987, 1990; Iwasaka et al., 1988; Simonson, 1995). The dust accumulation rate in Mediterranean soils has been estimated to be as much as 10 μm year^{-1} (Loye-Pilot et al., 1986), and rates of 200 μm year^{-1} have been estimated in West Africa (Orange and Gac, 1990). Dust accumulation in soils in Japan has been estimated at between 4 and 7 μm year^{-1} (Inoue and Naruse, 1991), between 2 and 5 μm year $^{-1}$ in southeastern Australia (Tiller et al., 1987) and of a similar order of magnitude in the southwestern USA (Reheis et al., 1995; Mayer et al., 1998). In many regions, dust inputs provide the sole or main source of key nutrients (Schlesinger, 1982; Tiessen et al., 1991; Swap et al., 1992; Chadwick et al., 1999). Regular dust input is particularly important for terrestrial ecosystems in the humid tropics and subtropics where leaching of nutrients takes place rapidly. Tiessen et al. (1991), for example, have suggested that all of the potassium and calcium in the topsoils of the Volta Basin, Ghana, was derived from Harmattan dust storms during the winter. Saharan dust is also thought to be an important source of plant micronutrients in the Amazon basin (Swap et al., 1992). The annual input of Saharan dust to the Amazon basin is estimated to be ca. 0.013 Pg year^{-1}, and result in the addition of between 0.5 and 4 kg ha^{-1} year^{-1} of key trace species such as K, NH_4 and NO_3 (Swap et al., 1992). The implications of changes in the dust cycle on glacial–interglacial timescales for terrestrial ecosystems have barely been considered, and would repay investigation.

4. Controls on the modern dust cycle

4.1. Deflation

Deflation of surface material occurs when the wind speed exceeds a critical threshold (i.e. the critical wind shear velocity; Bagnold, 1941) at which the strength of aerodynamic lift and drag equals the strength of the forces that hold surface particles together. The critical wind shear velocity depends on a number of factors intrinsic to the surface material (including grain size, shape and density) but can also be influenced by whether the surface particles occur as aggregates and the strength of inter-particle forces (Iversen et al., 1987; Zobeck, 1991; Marticorena et al., 1997a). Surface coherence is also increased by the presence of moisture between the particles, so that increases in soil moisture content impede dust deflation (McKenna Neumann and Nickling, 1989; McTainsh et al., 1998). The presence of either inorganic soil crusts (e.g. clay or salt crusts) or cyanobacterial and/or lichen soil crusts also increases surface coherence and impedes deflation (Gillette et al., 1980, 1982; Nickling, 1984; Williams et al., 1995; Marticorena et al., 1997b).

Certain characteristics of the land surface affect dust deflation by absorbing some fraction of the wind energy that would otherwise be available for erosion. In general terms, factors which increase the overall roughness of the surface (including, e.g. the presence of stones and pebbles, and the presence of vegetation such as xerophytic shrubs, even in a leafless state) decrease the energy of the surface winds and hence make it more difficult for the critical wind shear velocity to be reached (Zobeck, 1991; Raupach et al., 1993; Wolfe and Nickling, 1993; McKenna Neuman and Nickling, 1994; Marticorena et al., 1997a,b; Wyatt and Nickling, 1997).

Once the critical threshold for deflation is reached, vertical dust flux increases geometrically with increasing wind shear velocity (u). The relationship between wind shear velocity and dust flux has been proposed on theoretical grounds by Shao et al. (1993) to be proportional to u^3, and by Gillette and Passi (1988) to be proportional to u^4. Although there are field observations of the relationship between wind strength and vertical dust flux for a wide range of climate and surface types, as well as observations from wind tunnel experiments (see, e.g. Gillette et al., 1980; Nickling and Gillies, 1989, 1993; McKenna Neuman and Nickling, 1994; Shao et al., 1996; Stetler and Saxton, 1996; López, 1998), there is sufficient scatter in the data for it to be impossible to decide whether the u^3 or the u^4 relationship is most appropriate to use in a modelling context (Fig. 3). It may be that the assumption of a single proportionality relationship is too simplistic. A recent paper by Lu and Shao (1999) has suggested that dust flux is best described using a U proportionality varying between 2 and 3, dependent on the cohesivity of the surface material. This study emphasises that the energy available for emissions can be increased by bombardment of the surface by saltating particles. In addition to affecting the energetics of the system, saltation bombardment reduces surface coherence by breaking up, e.g. soil crusts and hence lowers the critical wind shear velocity required to initiate deflation.

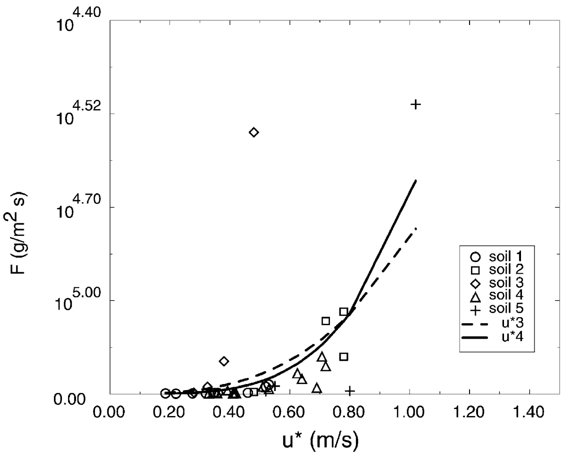

Fig. 3. The relationship between observed dust emissions (fluxes, F) and wind speed (u^*), from observations and wind tunnel experiments. The data are derived from Gillette (1977). The two lines show fitted u^3 and u^4 dependencies.

4.2. Transport

Dust-raising winds vary from small-scale convective vortices (dust devils), of the order of metres, up to gust fronts associated with thunderstorm events (haboobs) which can be up to 1000 km wide (Middleton, 1986b). Dust transport by these systems is a function of wind strength relative to the particle size of the emitted dust. Long-distance transport requires that meteorological conditions are favourable to carry the dust entrained by surface winds into higher levels of the troposphere. Such conditions generally involve *either* the formation of a deep thermally mixed layer by strong daytime heating of the land surface *or* the lifting of parcels of dust-laden air by cold fronts. The formation of a deep thermally mixed layer over the southern Sahara results, for example, in dust-laden air being carried to elevations of 3–5 km and incorporated into the African easterly jet (Karyampudi et al., 1999). Transport within the jet results in dust being transported across the Atlantic, reaching, e.g. the Caribbean and the Amazon Basin within 5–7 days (Prospero et al., 1970). In contrast, the lofting of dust-laden air into the upper troposphere over central Asia is associated with the passage of successive cold fronts. Long-distance transport of this dust by the Westerley jet results in it crossing the Pacific in 12–18 days (Merrill, 1989; Merrill et al., 1989, 1994).

4.3. Deposition

Dust is removed from the atmosphere by two mechanisms (see Prodi and Fea, 1979): dry deposition and scavenging by precipitation (also referred to as wet deposition). Dry deposition occurs through sedimentation, particle agglomeration through impaction and downward turbulent diffusion. Sedimentation and impaction are the most important processes for particles in the size range of most of the dust transported in the atmosphere. Diffusion is only

Table 1
Simulations of the modern dust cycle

Study	Model	Treatment of sources	Deflation	Deposition	Variable particle size	Variable mineralogy	Length of simulation
Joussaume (1983, 1985, 1990, 1993)	LMD AGCM, perpetual February and August simulations	sources defined as areas that are dry in both February and August	proportional to surface wind speed with 1 m s^{-1} minimum cutoff, and the vertical gradient of dust concentration	gravitational settling, turbulent mixing, wet deposition	No (uses 1 μm size fraction)	No	200 days
Genthon (1992)	GISS II tracer transport model ($7.82 * 10°$), with winds derived from $4 * 5°$ version of GISS AGCM	desert, grassland and shrubland from Matthews (1983) prescribed as source areas	u^3 wind speed dependence	gravitational settling, turbulent mixing, wet deposition	Yes, 3 fractions	No	27 months
Tegen and Fung (1994)	GISS tracer transport model ($8 * 10°$), with winds derived from $4 * 5°$ version of GISS AGCM	desert, grassland and shrubland from Matthews (1983) prescribed as source areas	SWC/SSC threshold, 20% for sand, 25% for silt, 50% for clay; u^3 wind speed dependence	gravitational settling, turbulent mixing, wet deposition	Yes, 4 fractions	No	13 months
Tegen and Fung (1995)	GISS tracer transport model ($8 * 10°$), with winds derived from $4 * 5°$ version of GISS AGCM	desert, grassland and shrubland from Matthews (1983) prescribed as source areas, plus disturbed soils	SWC/SSC threshold, 20% for sand, 25% for silt, 50% for clay; u^3 wind speed dependence	gravitational settling, turbulent mixing, wet deposition	Yes, 4 fractions	No	13 months
Tegen et al. (1996), Miller and Tegen (1998)	GISS tracer transport model ($8 * 10°$), with winds derived from $4 * 5°$ version of GISS AGCM and surface winds from ECMWF reanalysis data	desert, grassland and shrubland from Matthews (1983) prescribed as source areas	SWC/SSC threshold, 20% for sand, 25% for silt, 50% for clay; u^3 wind speed dependence	gravitational settling, turbulent mixing, wet deposition	Yes, 8 fractions	No	1 year

Andersen and Genthon (1996)	GISS II tracer transport model ($7.82 * 10°$), with winds derived from $4 * 5°$ version of GISS AGCM	desert, grassland and tundra from Matthews (1983) prescribed as source areas	SWC/SSC threshold, dust production is prevented by snow cover; u^4 wind speed dependence	gravitational settling, turbulent mixing, wet deposition	Yes, 3 fractions	No	2 years
Andersen et al. (1998)	LMDz model	desert from Matthews (1983) prescribed as source areas	SWC/SSC threshold, and dust production is prevented by snow cover; u^3 wind speed dependence	gravitational settling, turbulent mixing, wet deposition	No (uses 1 μm size fraction)	No	2 years
Reader et al. (1999)	CCCma GCMII with passive tracer model	sources defined as areas of bare soil, from CLIMAP (1981)	SWC/SSC threshold, 20% for all soils; u^3 wind speed dependence; empirical adjustment to deal with biases in simulated wind speeds	gravitational settling, turbulent mixing, wet deposition	Yes, 2 fractions	No	1 year
Mahowald et al. (1999)	TM3 transport model ($3.75 * 5°$), with 12-h winds derived from ECHAM3 AGCM	source areas prescribed from BIOME3 simulation ($0.5 * 0.5°$) driven by CLIMATE 2.0 climatology, wherever LAI < 0.35 (tropical or temperate biomes) or < 0.20 (cold biomes)	threshold of SWC/SSC of 20% for sand, 25% for silt, 50% for clay; wind threshold velocity $u > 5$ m s^{-1}; u^3 wind speed dependence	gravitational settling, turbulent mixing, wet deposition	Yes, spectral size fraction scheme	No	13 months

LMD: Laboratoire de Meteorologie Dynamique; GISS: Goddard Institute for Space Studies; CCCma: Canadian Centre for Climate Modelling and Analysis; ECHAM: European Community climate model, Hamburg; SWC/SSC: mean monthly water content/water content at field capacity; LAI: leaf area index.

important for extremely fine particles. Wet deposition occurs either by incorporation of dust into clouds to form cloud-condensation nuclei (in-cloud scavenging) or by incorporation of dust into rain droplets during precipitation events (below-cloud scavenging) (Slinn, 1983).

Dry deposition is less effective at removing dust from the atmosphere than wet deposition (Duce et al., 1980; Slinn, 1983; Arimoto et al., 1985; Uematsu et al., 1985). On average, only about 10–20% of the atmospheric dust load is removed by dry deposition. However, the relative importance of dry deposition increases as the mean size of the particles increases. Thus, in regions close to dust sources, dry deposition may remove 50% or more of the total dust from the atmosphere (e.g. Gao et al., 1997). The partitioning between dry and wet deposition, and the effectiveness of each, is highly dependent on local meteorological conditions. This results in both wet and dry deposition rates being highly variable temporally and spatially. For example, about 50% of the annual flux of mineral dust at Midway, in the northern Pacific, measured during the SEAREX (Sea-Air Exchange) campaign, occurred during a 2-week period (Prospero et al., 1989). Similarly, ca. 0.001 Pg dust was deposited at a sampling site in Corsica as the result of a single 3-day dust storm which originated in the northern Sahara (Dulac et al., 1992).

5. Simulation of the modern global dust cycle

Most of what is known about the controls on dust emission, transport and deposition is a result of observational programmes at a local or regional scale. Until recently, the modelling of dust has also focussed on the local to regional scale, and particularly on the controls of dust emission at these scales (e.g. Westphal et al., 1988; Marticorena and Bergametti, 1995; Marticorena et al., 1997a; Schulz et al., 1998). The high spatial and temporal variability characteristic of all aspects of the dust cycle poses serious problems for modelling at the global scale, as does the need to consider the heterogeneous nature of the dust itself. Global modelling of the dust cycle only began in the mid-1980s (Table 1).

Dust is treated as a passive tracer in global models. In the first global simulations (Joussaume, 1983, 1985, 1990, 1993), which were run for perpetual February and perpetual August conditions, the dust sources were defined as those areas in which the simulated soil moisture, averaged over the both months was less than 2 mm/day, and deflation was a function of wind speed. The dust was treated as though it were uniform in size (1 μm) and mineralogy. Dust deposition was allowed to take place through both wet and dry processes. Despite its relative simplicity, the model produced a reasonable simulation of the extent of modern source areas (except that the Australian source was too small) and of the spatial patterns of dust transport. Atmospheric dust loadings were twice as high in August than February, a seasonal difference supported by the observations of dust emissions.

Subsequent modelling work has focused on improved prescription of source areas in terms of surface wetness and vegetative cover (Table 1). In work with the Goddard Institute for Space Studies (GISS) tracer transport model (Tegen and Fung, 1994, 1995), deflation was allowed to occur when the soil moisture content was less than a threshold value which varied according to soil texture; specifically deflation over sandy soils occurred when the soil moisture content was less than 20% of field capacity on sandy soils, 25% of field capacity on silty soils, and 50% of field capacity on clay soils. This scheme is widely used in other simulations of the dust cycle (e.g. Tegen et al., 1996; Mahowald et al., 1999). Tegen and Fung (1994) also used vegetation type to prescribe the extent of potential source areas for emissions. Specifically, deflation was only allowed to occur in regions characterised by desert, grassland and shrubland vegetation, prescribed according to the global map of Matthews (1983).

However, approaches that define potential dust source areas in terms of the simple presence/absence of specific vegetation types (biomes) neglect the fact that there can be considerable variation in the relative proportions of vegetated and bare ground surface within biomes, as well as changes in the seasonal coverage of vegetation. These issues could be taken into account by using, e.g. satellite NDVI (normalised difference vegetation index, a measure of greenness) measurements or vegetation models, which explicitly simulate the density of the vegetative cover (e.g. as leaf area index (LAI)) depending on climate and soil properties. Mahowald et al. (1999) have used this vegetation-modelling approach to define dust source areas. In their experiments, Mahowald et al. (1999) used an off-line vegetation model (BIOME3: Haxeltine and Prentice, 1996) to simulate LAI as a function of modern climate, and then defined potential dust source areas to occur wherever LAI was below defined thresholds. The recent development of coupled atmosphere-vegetation models (e.g. Foley et al., 1998), which predict both the spatial patterns and the seasonal cycle of LAI, opens up further possibilities for improving the definition of vegetation-determined dust source areas. Despite the evidence that other characteristics of the land surface (e.g. roughness, surface armouring, soil crusting) can have a significant impact on whether a region is a source of dust, and modelling studies showing that prescribing such characteristics considerably improves the simulation of dust emissions from northwestern Africa (Marticorena et al., 1997a), no model of the global dust cycle has yet attempted to take local variations in land-surface conditions into account. This largely reflects the lack of global data sets of surface characteristics pertinent to dust erosion at a local scale.

Further improvements to dust models have focussed on the characterisation of the dust itself. Genthon (1992) introduced the idea of modelling

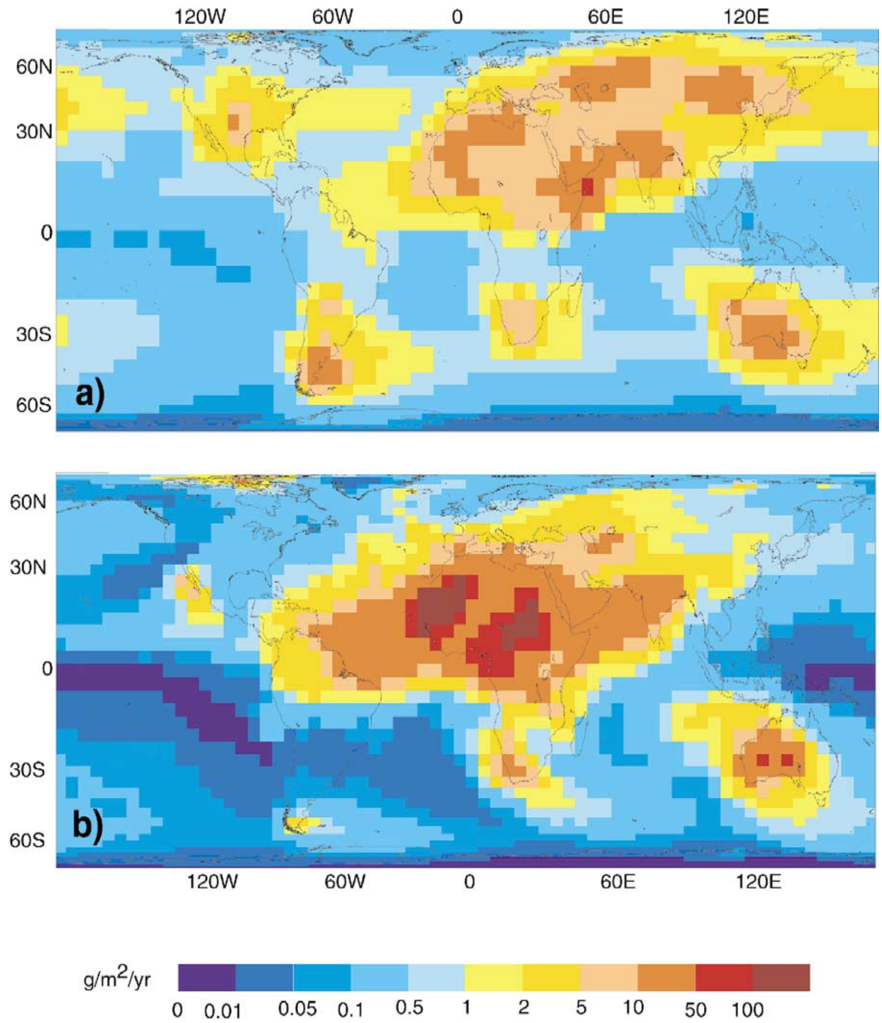

Fig. 4. Modern dust deposition fields as simulated by (a) Tegen and Fung (1995), and (b) Mahowald et al. (1999).

different size distributions, treating the dust load as comprising three fractions (clay, silt, sand). Tegen and Fung (1994) split the silt-sized fraction so that the dust load was treated as comprising four fractions (clay, fine silt, coarse silt, sand). Several other studies have used a similar approach, with dust of different sizes being treated as comprising discrete tracers, although the number of size fractions considered has been increased (Table 1). More recent tracer transport models have used a computationally efficient spectral scheme to describe the dust size distribution (e.g. Schulz et al., 1998; Mahowald et al., 1999). The recognition that dust mineralogy is important for the correct simulation of the radiative impacts of dust (Claquin et al., 1998; Sokolik et al., 1998), has led to recent attempts to take differences in dust mineralogy into account in model simulations (e.g. Claquin et al., submitted).

Global modelling of the dust cycle is still in its infancy, and much needs to be done to improve the existing models. Nevertheless, the most recent simulations of the modern dust cycle (e.g. Tegen and Fung, 1995; Andersen et al., 1998; Mahowald et al., 1999) reproduce the broad-scale spatial and temporal features of the observed dust cycle (Fig. 4). In particular, these simulations capture the major dust plumes originating from northern Africa and transported over the subtropical Atlantic and over the Arabian Sea, and from central Asia and transported over the northern Pacific. The simulations also show the seasonal migration of these dust plumes, in response to the migration of the sub-tropical anticyclones and the inter-tropical convergence zones and capture the seasonal cycle in atmospheric dust loading shown by observations.

Some key features of the modern simulations made with the current generation of dust cycle models are clearly unrealistic. The simulated Asian dust plume tends to be less important than shown by satellite observations, largely because wind strengths appear to be underestimated in the simulations. This appears to be a function of model resolution: simulations of the dust cycle are commonly made with atmospheric models with a resolution of ca. 4° latitude by 5° longitude. Tegen and Miller (1998) have shown that using higher-resolution winds, specifically ECMWF winds at ca. $1.25° \times 1.25°$ resolution (Gibson et al., 1997; Kallberg, 1997) produced a better simulation of the magnitude of Asian dust emissions. The current generation of dust models tends to overestimate dust emissions from Australia. This probably reflects problems in the specification of surface conditions, and specifically soil texture, for this region from the currently available global data sets (e.g. Zobler, 1986; Webb et al., 1991). Experiments with a more realistic soils data set for Australia yield much lower rates of emissions (Roelandt, unpublished results).

6. Dust at the last glacial maximum

6.1. Ice-core records of changes in dust during the quaternary

Stratigraphic records from polar ice cores show that the atmospheric dust content has varied substantially both on glacial–interglacial, and at sub-Milankovitch time scales. Aeolian deposition rates were ca. 2–20 times greater during glacials than interglacials (e.g. Petit et al., 1990; Hammer et al., 1985; Steffensen, 1997). The factor for concentrations of dust in the ice cores is even larger because of low precipitation rates during the glacial periods. The dust record from the polar ice cores shows millenial-scale variations similar to other palaeoenvironmental indicators in the ice cores (e.g. CO_2, CH_4, $\delta^{18}O$ atm, δD) (e.g. Zielinski and Mershon, 1997). High-resolution records of changes in dust content during the last glaciation show even larger inter-annual to decadal variability (Taylor et al., 1993b; Steffensen, 1997).

The degree to which the atmospheric dust loading is a response to or a contributory cause of climate changes on glacial–interglacial time scales is still uncertain. The relationship between changes in atmospheric dust loading and other palaeoenvironmental indicators in the ice cores is not straightforward. Thus, the CO_2 concentration had already reached near-glacial levels (ca. 205 ppmv) by the time dust concentrations in the Antarctic ice cores began to increase around 65 kyr (Fig. 5). In contrast, the decrease in dust loading appears to be synchronous with, or even to precede the increases in atmospheric CO_2 concentrations during deglaciations. The differ-

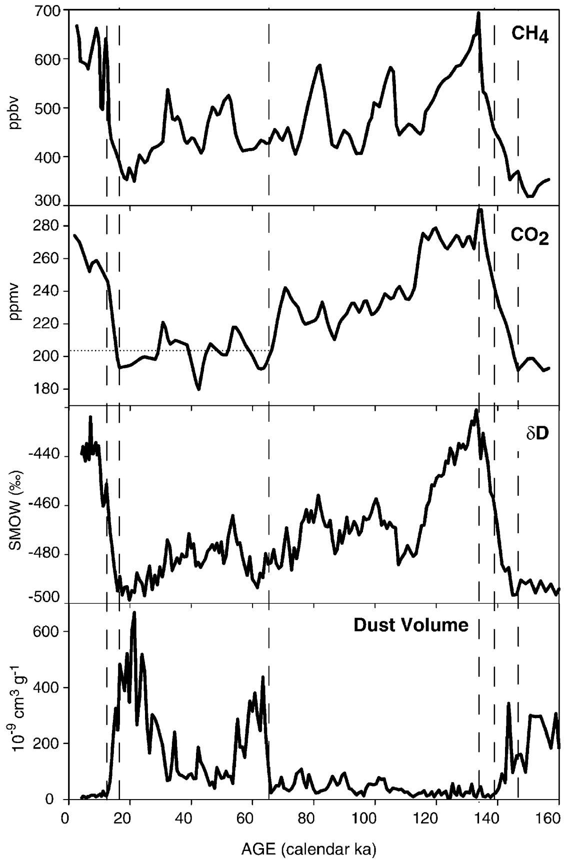

Fig. 5. Changes in concentrations of CH_4 (Jouzel et al., 1993) and CO_2 (Barnola et al., 1987), δD (Jouzel et al., 1987), and dust (Petit et al., 1999) during the last glacial–interglacial cycle as measured in the Vostok ice core.

ent relationships between dust, other atmospheric constituents and climate through time suggests that there may be complex feedback effects in operation. So far, much of this complexity has been disregarded and investigations of the palaeoclimatic role of dust have focussed on the last glacial maximum (LGM). The LGM provides a logical first target for modelling of the palaeodust cycle because (a) the glacial forcing (changes in ice sheet height and extent, in atmospheric composition and in sea-surface conditions) is relatively well known (e.g. Barnola et al., 1987; CLIMAP, 1981; Peltier, 1994; Weinelt et al., 1996; de Vernal et al., 1997; Rochon et al., 1998; Mix et al., 1999), (b) the observed changes in dust loading were large between the LGM and the Holocene, and (c) there is a wealth of radiometrically dated palaeoenvironmental data that can be used to evaluate the simulations.

6.2. Observed patterns of dust deposition at the LGM

Stratigraphic records of dust deposition at the LGM and for the late Holocene (broadly representing modern conditions) are available from polar (Thompson and Mosley-Thompson, 1981; Petit et al., 1990; Taylor et al., 1993a; Hansson, 1994; Hammer et al., 1985; Mayewski et al., 1994; Steffensen, 1997; Petit et al., 1999) and tropical (Thompson et al., 1990, 1995, 1997, 1998) ice caps, marine sediments (Sirocko et al., 1991; Rea, 1994 and references therein; Hesse, 1994; Ruddiman, 1997), and loess deposits (Ding et al., 1994; Shackleton et al., 1995). These records confirm that the LGM was an interval of much higher atmospheric dust loadings than today, but they also show the existence of strong spatial patterns of dust enhancement (Fig. 7f; Kohfeld and Harrison, 2000; this volume). Although much of the world experienced increased dust deposition, there were regions where dust fluxes at the LGM were less than today. For example, dust fluxes from Africa to the tropical and subtropical Atlantic were three to five times higher than modern, but were less than modern in the Gulf of Guinea. Marine cores from the North Pacific showed increased dust deposition (one to two times modern values) but dust deposition in the tropical Pacific was reduced compared to modern. The reduction in deposition rates in the tropical Pacific is consistent with spectral analysis of marine records covering several glacial–interglacial cycles off northwestern South America which show that dust loadings were less during glacials than interglacials (Rea, 1994). Records from other regions show increased dust deposition at the LGM, but these increases were slight compared to the major changes observed, e.g. over the Atlantic or in the polar regions. Non-coastal marine cores in the Arabian Sea show increases in dust deposition of only 60–80% compared to modern. In contrast, LGM dust deposition in polar regions, though small in absolute magnitude (0.001–0.1 g m^2 year^{-1}) represent a factor 2–20 enhancement.

6.3. Dust provenance at the LGM

The pathways of dust transport at the LGM can to some extent be inferred from the observed spatial patterns in dust deposition rates. It is more difficult to specify the source of dust reaching the polar ice caps, and yet this information is critical to our understanding of the causes of the increased dust loading during the LGM. One approach that has been successfully applied to reconstruct dust transport paths is the use of mineralogical and isotopic tracers (see, e.g. Grousset and Chesselet, 1986; Grousset et al., 1988, 1992; Nakai et al., 1993; Zhang et al., 1996; Caquineau et al., 1998). The most commonly used tracers are the clay minerals (e.g. kaolinite, smectite, illite) and the long-lived radiogenic isotopes (e.g. $^{87}Sr/^{86}Sr$, $^{206,207,208}Pb$, ^{204}Pb, $^{143}Nd/^{144}Nd$). Dust samples from the geological archive (e.g. an ice core or a marine core) are characterised in terms of the presence/absence and relative abundance of these tracers, and then compared with samples taken from potential sources (Grousset et al., 1992; Basile et al., 1997; Biscaye et al., 1997). The method assumes that the mineralogy and the radiogenic isotope composition of dust reflects the geographical location of its source area, and is not substantially modified during transport. The assumption that the original composition of the dust is not changed during transport is best satisfied by considering only the fine fraction (< 5 μm).

This approach has been used to establish the provenance of glacial dust in the polar ice cores. Glacial dust from the GISP 2 ice core (Greenland) has abundant illite, no smectite, and a kaolinite/chlorite ratio of 0.3–1.2 (Biscaye et al., 1997). Comparisons with samples from glacial-age loess, sands, tills and lacustrine sediments from northern North America (Alaska, Illinois, Toronto), the Sahara, the Ukraine, the Gobi Desert and the Chinese Loess Plateau (Biscaye et al., 1997) suggest that the dust in the Greenland core could not have come from Africa (high smectite, high kaolinite/chlorite ratio) or North America (high smectite, low $^{143}Nd/^{144}Nd$ ratios). Biscaye et al. (1997) suggest eastern Asia is the most likely source for the dust deposited in Greenland during the LGM. Similar studies of the LGM record from the Antarctic (Vostok, Dome C) suggest that the dust deposited there was derived from Patagonia (Grousset et al., 1992; Basile et al., 1997).

Mineralogical and isotopic tracers can potentially be used to identify the transport pathways of dust contributing to terrestrial loess. Aleinikoff et al. (1998), for example, have used Pb isotopes to source LGM loess deposits in North America. Such analyses are potentially more difficult than provenance studies on dust from ice cores because of the possibility that the aeolian material has been contaminated by admixture of fine-grained material of non-aeolian origin or modified by reworking. Either of these processes could potentially invalidate the assumption that the dust composition is not modified during transport and so reflects the mineralogy and isotopic composition of its source area. Nevertheless, the application of mineralogical and isotopic tracers to elucidate the origins and transport history of individual terrestrial loess deposits is promising (Aleinikoff et al., 1998, 1999; Muhs et al., 1999).

Minerological and isotopic tracers provide a powerful tool to identify dust transport pathways at key times in the past. However, the validity of the conclusions drawn from such studies depends on whether the potential dust source areas have been adequately sampled (Kohfeld and Harrison, this volume). For example, the northern hemisphere high latitudes were eliminated as a potential source for dust reaching Greenland based on only six samples from north of 50°N (Biscaye et al., 1997). Similarly, Australia was eliminated as a potential source for dust reaching Vostok and Dome C based on five samples from two sites (Grousset et al., 1992). The sampling problem is compounded by the difficulties of adequately dating, in particular, loess and sand deposits to ensure that the source samples are temporally equivalent to the target samples (see, e.g. the debate about the existence of glacial age loesses in Alaska; Beget, 1996).

6.4. Causes of changes in atmospheric dust concentrations at the LGM

Three factors have been invoked to explain why atmospheric dust loadings and dust deposition rates were higher during glacials compared to interglacials: increased winds speeds, the reduced intensity of the hydrological cycle, and expansion of dust source areas. Increased wind speeds could allow more dust to be entrained and carried to remote areas. The reduced intensity of the hydrological cycle could allow dust to remain longer in the atmosphere (Hansson, 1994; Yung et al., 1996). An increase in dust supply could result from the expansion of source areas due to (a) extensive fine-grained outwash deposition along the margins of the northern hemisphere ice sheets (Junge, 1979); (b) the increased land area particularly in the tropics and subtropics, with the exposure of the continental shelves consequent on lowered sea level (Ono and Naruse, 1997); or (c) reduced vegetation cover and/or soil moisture (Petit et al., 1981; Joussaume, 1990; Genthon, 1992).

Simulations of glacial climates show an increase in glacial wind intensities both at the surface and aloft (e.g. COHMAP Members, 1988; Harrison et al., 1992; Kutzbach et al., 1998; Pinot et al., 1999). There is, however, little direct evidence for increased wind strengths at the LGM. Extensive dune fields across mid-continental North America have been used to infer high-velocity surface winds south of the Laurentide and Cordilleran ice sheets at the LGM (Kutzbach and Wright, 1985), but more recent studies suggest that these deposits were not formed exclusively during the glacial period. Some of these dunes were reactivated in the mid-Holocene (Muhs et al., 1996). In marine sediments, increased wind strengths have been inferred from increases in the median grain size of aeolian material (Sarnthein and Koopman, 1980; Sarnthein et al., 1981; Rea and Leinen, 1988; Clemens and Prell, 1990) and from the shallowing of thermocline depths off northern Africa (Molfino and McIntyre, 1990; Ravelo et al., 1990; Ruddiman, 1997) and in the equatorial Pacific (Andreasen and Ravelo, 1997; Trend, 1999).

It follows from the Clausius–Clapeyron relationship that globally colder temperatures lead to in a reduction in the intensity of the hydrological cycle (e.g. see Rind, 1986; Hartmann, 1994). Although the magnitude of global cooling at the LGM is still uncertain, climatic reconstructions based on both marine (Rostek et al., 1993; Schneider et al., 1995; Bard et al., 1997; Sonzogni et al., 1998) and terrestrial (Farrera et al., 1999) evidence indicate that the tropics were cooled by ≈ 3°C. Far more extreme temperature changes occurred in the northern extratropical regions (Webb et al., 1993; Peyron et al., 1998; Tarasov et al., 1999). Reconstructions based on pollen evidence for changes in vegetation indicate that southern Europe was ca. 10°C colder in annual mean (Peyron et al., 1998). Pollen-reconstructed temperatures across European Russia were between 5°C and 10°C lower than present in summer and > 20°C lower than present in winter (Tarasov et al., 1999). Temperature reconstructions from eastern North America show conditions were > 6°C colder than today everywhere except the southern Florida peninsula (Prentice et al., 1991; Webb et al., 1993), with extremes of up to 12–16°C (depending on the season) colder than today. Geomorphic and biostratigraphic evidence of changes in lake levels indicate that runoff was less than today over most of the globe (Farrera et al., 1999; Kohfeld and Harrison, 2000), although some regions (e.g. southwestern USA; high elevations in the Andes) are notable exceptions. Pollen and plant macrofossil records of the expansion of desert, grassland and shrubland vegetation in areas occupied today by forests (Edwards et al., 2000; Elenga et al., 2000; Prentice et al., 2000; Tarasov et al., 2000; Yu et al., 2000), as a consequence of the reduction in plant available moisture (see Farrera et al., 1999), also support the idea that the intensity of the hydrological cycle was reduced compared to today, though again certain regions show a different response, notably the southwestern USA (Thompson et al., 1993; Thompson and Anderson, 2000).

Geomorphic and biostratigraphic data also provide abundant evidence for an increase in potential dust source areas at the LGM (Fig. 6). Vegetation data compiled as part of the Palaeovegetation Mapping Project (BIOME 6000; Prentice and Webb, 1998) shows that both temperate and tropical forests were substantially fragmented and reduced in extent at the LGM (Edwards et al., 2000; Elenga et al., 2000; Prentice et al., 2000; Tarasov et al., 2000; Yu et al., 2000). The expansion of desert, grassland and shrubland vegetation would have resulted in a substantial increase in the area subject to deflation either seasonally (in the grassland and shrubland regions) or throughout the year (in the extended desert regions). Dry lake basins (e.g. McTainsh, 1985, 1989; Chadwick and Davis, 1990; White and Drake, 1993)

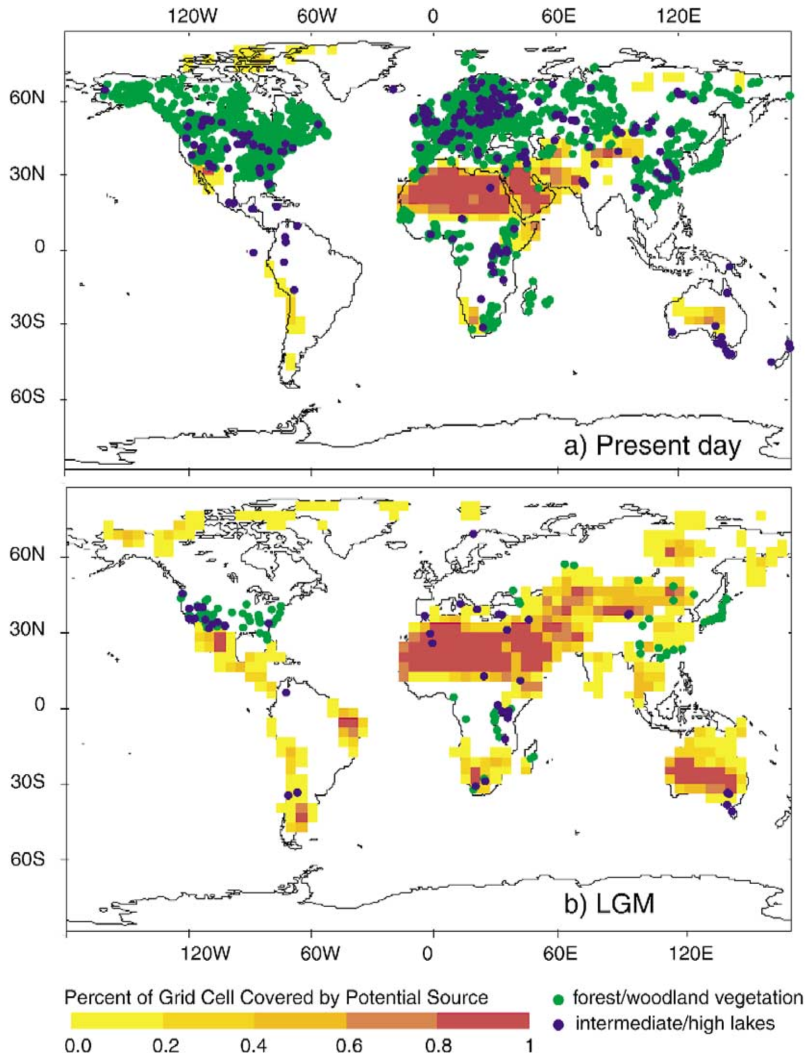

Fig. 6. Simulated potential source areas from the Mahowald et al. (1999) experiments, (a) at the present day, and (b) at the LGM as a result of simulated changes in climate and vegetation changes (redrawn from Kohfeld and Harrison, 2000). Superimposed on these maps are the observed modern and LGM distribution of forests (green circles; from the BIOME 6000 data set: Edwards et al., 2000; Elenga et al., 2000; Tarasov et al., 2000; Yu et al., 2000; Takahara et al., 2000; Williams et al., 2000; Thompson and Anderson, 2000; Prentice et al., 2000) and intermediate or high lakes (blue circles; from the Global Lake Status Data Base: Kohfeld and Harrison, 2000).

could also have acted as local sources for dust deflation. Given the predominantly fine-grained nature of lacustrine sediments, dry lake basins would have been particularly effective sources for dust. The extensive loess cover in mid-continental North America (Ruhe, 1983), in South America (Sayago, 1995), in continental Europe (Kukla, 1975, 1977), in Australia (Nanson et al., 1992; McTainsh and Lynch, 1996) and in central Asia (Velichko, 1984; Dodonov and Baiguzina, 1995) provides evidence that the LGM was considerably more arid than today. These glacial-age loesses (Pye, 1987) indicate the location of major centres of dust deposition (i.e. sinks of the atmospheric dust load) but they also, through re-entrainment, would themselves have provided a significant source of dust to the glacial atmosphere.

Table 2
Characteristics of simulations of the dust cycle at the Last Glacial Maximum

Study	Effects examined	Models	Resolution	Length of simulation	Estimate of change in dust concentration over Greenland and Antarctica
Joussaume (1990, 1993)	LGM winds and hydrology	LMD AGCM, perpetual February and August	3.6° (lat) × 5.625° (long), 11 vertical layers	100 days	× 4 for Greenland × 1.2 for Antarctica
Genthon (1992)	LGM winds and hydrology	GISS Model II	7.82° (lat) × 10° (long), 9 vertical layers	27 months	× 1.1 for West Antartica × 2.2 for East Antartica
Andersen and Genthon (1996)	LGM winds and hydrology	GISS Model II	7.82° (lat) × 10° (long), 9 vertical layers	1 year	less than modern in Greenland
Yung et al. (1996)	LGM hydrology	2-D model	10° in latitude 40 vertical layers	1 year	× 3 for Antarctica
Andersen et al. (1998)	LGM winds and hydrology, change in source areas due to changing soil moisture	LMDz	3.75° (lat) × 5.625° (long), 15 vertical layers	1 year	× 2.7 for Greenland × 1.4 for Antartica
Reader et al. (1999)	LGM winds and hydrology; source areas inferred from inverse exercise	CCCma GCMII; passive tracer transport model	T32 spectral resolution, 10 vertical layers	1 year	× 6 for Greenland × 3 for Antartica
Mahowald et al. (1999)	LGM winds and hydrology, change in source area due to soil moisture and vegetation changes	ECHAM3, BIOME3, TM3	ECHAM3: 4° (lat) × 5° (long), 19 vertical layers BIOME3: 0.5° × 0.5° TM3: 3.75° (lat) × 5.° (long), 9 vertical layers	13 months	× 2.5 for Greenland × 12.6 for Antartica

LMD: Laboratoire de Meteorologie Dynamique; GISS: Goddard Institute for Space Studies; CCCma: Canadian Centre for Climate Modelling and Analysis; ECHAM: European Community climate model, Hamburg.

6.5. Simulation of the dust cycle at the LGM

Several modelling studies (Table 2) have tried to simulate the dust cycle at the LGM in response to changes in wind intensity and the hydrological cycle only (Joussaume, 1990, 1993; Genthon, 1992; Andersen and Genthon, 1996; Yung et al., 1996). These simulations did not consider the expansion of source areas during the LGM, and were unable to match the observed increases in deposition at the polar ice caps.

Two recent studies (Andersen et al., 1998; Reader et al., 1999) have examined the impact of climatically induced changes in soil moisture on dust deflation. Andersen et al. (1998) suggest that the simulated increase in source areas at the LGM caused by the reduction in soil moisture relative to today was modest. Thus, although incorporating the effect of reduced soil moisture on source areas produces an improvement in the simulated dust fields (Andersen et al., 1998; Reader et al., 1999), the simulations still fail to capture all of the observed increases in dust deposition. In the Reader et al. (1999) experiment, the areas of land produced as a result of sea-level lowering at the LGM were prescribed to be bare soil. Thus, these simulations also provide an inadvertent test of the hypothesis put forward by Ono and Naruse (1997) that the exposed areas of the continental shelf acted as extended source areas. The simulations show that expanding the source areas of dust on the continental shelf does not produce a sufficient overall increase in atmospheric dust loadings. Inverse analyses, in which the simulated LGM dust fields were compared to a limited set of ice core and marine records of LGM aeolian deposition, confirm that more significant changes in source areas are required than can be explained by purely physical changes in the land-surface characteristics (Reader et al., 1999).

The most recently published simulation shows that incorporating changes in source areas due to vegetation changes produces a semi-quantitatively realistic simulation of the LGM dust cycle (Mahowald et al., 1999). Mahowald et al. (1999) used BIOME3 (Haxeltine and Prentice, 1996) to estimate natural dust source regions, a simple source function related to soil texture and wind velocity (Marticorena and Bergametti, 1996; Schulz et al., 1998) to estimate dust entrainment, and an atmospheric transport model (TM3, developed from TM2; Heimann, 1995) to calculate three-dimensional dust transport. The modern and LGM wind fields and surface climates required to drive these models were derived from an atmospheric general circulation model (ECHAM 3.2; Lorenz et al., 1996) forced with CLIMAP sea-surface temperatures for the LGM and appropriately reduced atmospheric CO_2. BIOME3 was run in two modes: with and without allowing for the physiological effects of lowered CO_2 at the LGM (≈ 200 ppmv; Barnola et al., 1987; Raynaud et al., 1993; Smith et al., 1999) on plant productivity and water-use efficiency. Thus, it was possible to compare the dust fields obtained as a result of changing climate alone (i.e. winds and hydrological cycle), changing source areas due to climate-induced vegetation changes, and changing source areas due to vegetation changes brought about by the physiological effects of lowered CO_2 as well as climate (Fig. 7). This study confirmed that simulated changes in precipitation and wind intensity alone could not account for the full observed increase of atmospheric dust content at the LGM, although climatic changes alone could explain the increased dust deposition shown by marine cores off western Africa. However, a simulated expansion of source areas in high latitudes (Siberia, Patagonia) and in central Asia produced a 2.5-fold increase in the global dust content of the atmosphere and a 20-fold increase in high latitudes, consistent with the high dust fluxes observed in polar ice cores. About half of the simulated expansion in source area was due to climate change, and about half to the physiological effect of lowered glacial CO_2 in reducing the vegetation cover of C_3 plants (see, e.g. Ehleringer et al., 1997; Jolly and Haxeltine, 1997; Levis et al., 1999). The simulated expansion of source regions is consistent with comparable estimates (Harrison et al., submitted) based on several other AGCMs that performed identical LGM simulations within the Palaeoclimate Modelling Intercomparison Project (PMIP; Joussaume and Taylor, 1995; Pinot et al., 1999), and with the expanded LGM desert areas shown by pollen data compiled in the BIOME 6000 project (Prentice et al., 2000; Yu et al., 2000). The simulated expansion of source areas in the high northern latitudes is consistent with pollen evidence for greater aridity (Edwards et al., 2000; Tarasov et al., 2000) and the postulated occurrence of glacial-age loess deposits (e.g. Beget, 1996). The model results are consistent with isotopic measurements on ice cores, which indicate a predominantly central Asian source for the LGM dust recorded in Greenland (Biscaye et al., 1997) and a Patagonian source for Antarctic dust (Grousset et al., 1992; Basile et al., 1997), except that the model also simulates a significant Australian source which is not confirmed by isotopic sourcing of Antarctic dust.

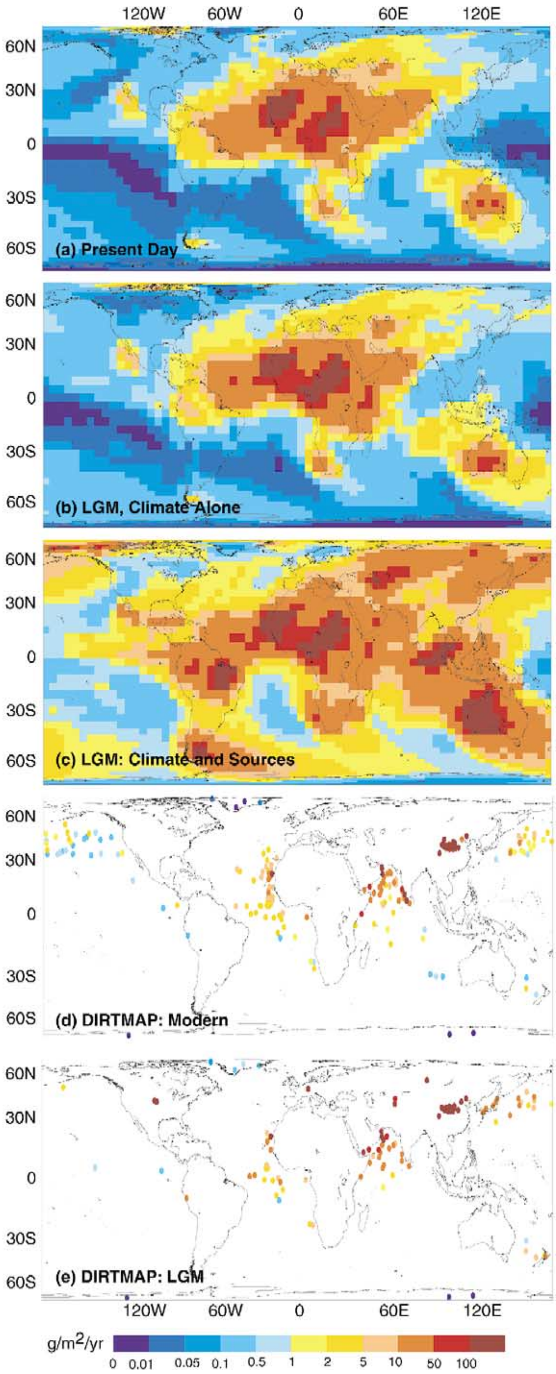

Fig. 7. Simulated dust fields at (a) the present day, (b) the LGM due to climate changes only, and (c) the LGM due to climate-induced changes in biome distribution and the physiological impacts of lowered CO_2, from the Mahowald et al. (1999) experiments, compared to (d) the modern observed dust deposition, (e) the LGM observed dust distribution, from the DIRTMAP data base (Kohfeld and Harrison, this volume).

6.6. The effects of dust on radiative forcing at the LGM

Atmospheric general circulation model simulations of the LGM have been unable to produce the observed magnitude of cooling in the high-to mid-latitudes as a response to ice-sheet albedo and CO_2 forcing alone (e.g. Masson et al., 1998). Simple energy-balance calculations have suggested that the dustiness of the glacial atmosphere could have produced an additional cooling of 1–3°C (Harvey, 1988; Anderson and Charlson, 1990; Crowley and North, 1991; Hughes, 1992; Table 3). However, the only study to address this issue with a three-dimensional model (Overpeck et al., 1996) showed a large positive forcing over the ice-sheet surfaces and thus an overall global warming due to glacial dust (Fig. 8). Peltier and Marshall (1995) argued that this type of positive forcing, in particular due to the accumulation of dust in the ablation zones of the great northern hemisphere ice sheets, may have hastened their ultimate collapse. However, the dust field prescribed by Overpeck et al. (1996) was based on modern maximum dust concentrations, and not explicitly related to either wind patterns or sources at the LGM. Thus, the latitudinal distribution of dust was unrealistic and, in particular, under-represented the relative amounts of dust in the tropics relative to northern high latitudes. In addition, the radiative effects of dust over the ocean were largely neglected, as were changes in land-surface albedo outside the glaciated

Table 3
Simulations of the radiative effects of LGM dust loadings

Study	Model	Derivation of dust field	Estimated change in TOA forcing ($W\ m^{-2}$)	Estimated change in global surface temperature (K)
Harvey (1988)	2-D energy balance climate model (EBCM)	increase in continental aerosol optical depth greater by 2 in the tropics, with a linearly increasing multiplication factor up to 6 at 90°N and 8 at 90°S in the tropics	−2.04	−1.72
Overpeck et al. (1996)	GISS AGCM simulation with a) prescribed CLIMAP SSTs	prescribed dust field with same magnitude and distribution in each month	−0.40	0.19
	b) mixed-layer ocean		0.0	0.13
Claquin et al. (submitted for publication)	offline radiative calculation, using Morcrette scheme; cloud, humidity and temperatures fields from ECHAM3 AGCM simulation with prescribed CLIMAP SST	realistic global dust fields from Mahowald et al. (1999)	N of 45°N: −0.3 to −0.9 tropics: −2.2 to −3.2 S of 45°S: −0.3 to −0.2	not estimated

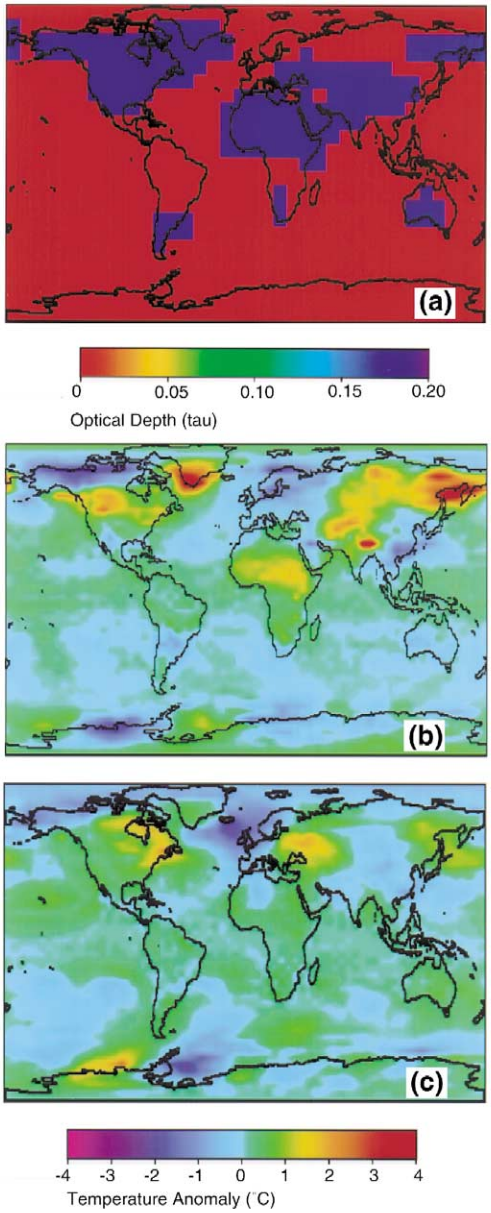

Fig. 8. (a) Prescribed tropospheric dust loading (expressed as optical depth) at the LGM used in the Overpeck et al. (1996) simulations. The simulated changes in surface temperatures (LGM-control dust simulations) are derived from (b) a simulation with prescribed CLIMAP sea-surface temperatures, and (c) a simulation in which the sea-surface temperatures at the LGM were simulated using a mixed-layer ocean model. Both LGM experiments show a significant warming in the high northern latitudes.

areas. Although the study of Overpeck et al. (1996) shows that LGM dust could have an important radiative effect (Table 3), and emphasises that dust does not necessarily produce cooling at the Earth's surface, it does not provide a realistic estimate of the magnitude, spatial distribution or even necessarily the global sign of the effect.

Claquin et al. (submitted for publication) have assessed the radiative impact of dust at the LGM (Table 3), using the more realistic dust fields simulated by Mahowald et al. (1999) and taking into account the effects of mineralogical provenance on radiative properties. Seasonally varying surface albedo fields for the present-day and at the LGM were estimated using a simple scheme from BIOME3-simulated vegetation type, leaf area index and snowpack. The present-day surface albedo distribution simulated by this procedure compares reasonably well with the broad scale patterns shown by ERBE satellite data (Fig. 9), particularly in the summer. The surface albedo was significantly higher in the high- to mid-northern latitudes at the LGM (Fig. 9g,h), in response to the presence of the Laurentide and European ice sheets, and to the expansion of seasonally snow-covered tundra. There was a year-round increase in surface albedo in the subtropics, due to the expansion of deserts. Claquin et al. (submitted for publication) showed that the high-latitude (poleward of 45°N) change in forcing (Fig. 10) was indeed positive but small (-0.3 to -0.9 W m^{-2}) compared to ≈ -20 W m^{-2} caused by the extension of snow and ice (Hewitt and Mitchel, 1997). Much of the high-latitude dust was not located over the ice sheets, but over unglaciated regions close to the expanded Central Asian source. The tropical change in forcing was negative (-2.2 to -3.2 W m^{-2}). This negative forcing represents an effect of comparable magnitude to the cooling effect of low atmospheric CO_2 concentrations (Hewitt and Mitchel, 1997). These results suggest that high-latitude warming by ice-age dust was not important for the energy balance of the glaciated regions or the dynamics of the ice sheets, while it may have contributed significantly to the tropical sea-surface cooling during glacial times. This implies a predominantly positive feedback role for dust because low sea-surface temperatures, CO_2 concentration and dust loadings in low latitudes are mutually reinforcing.

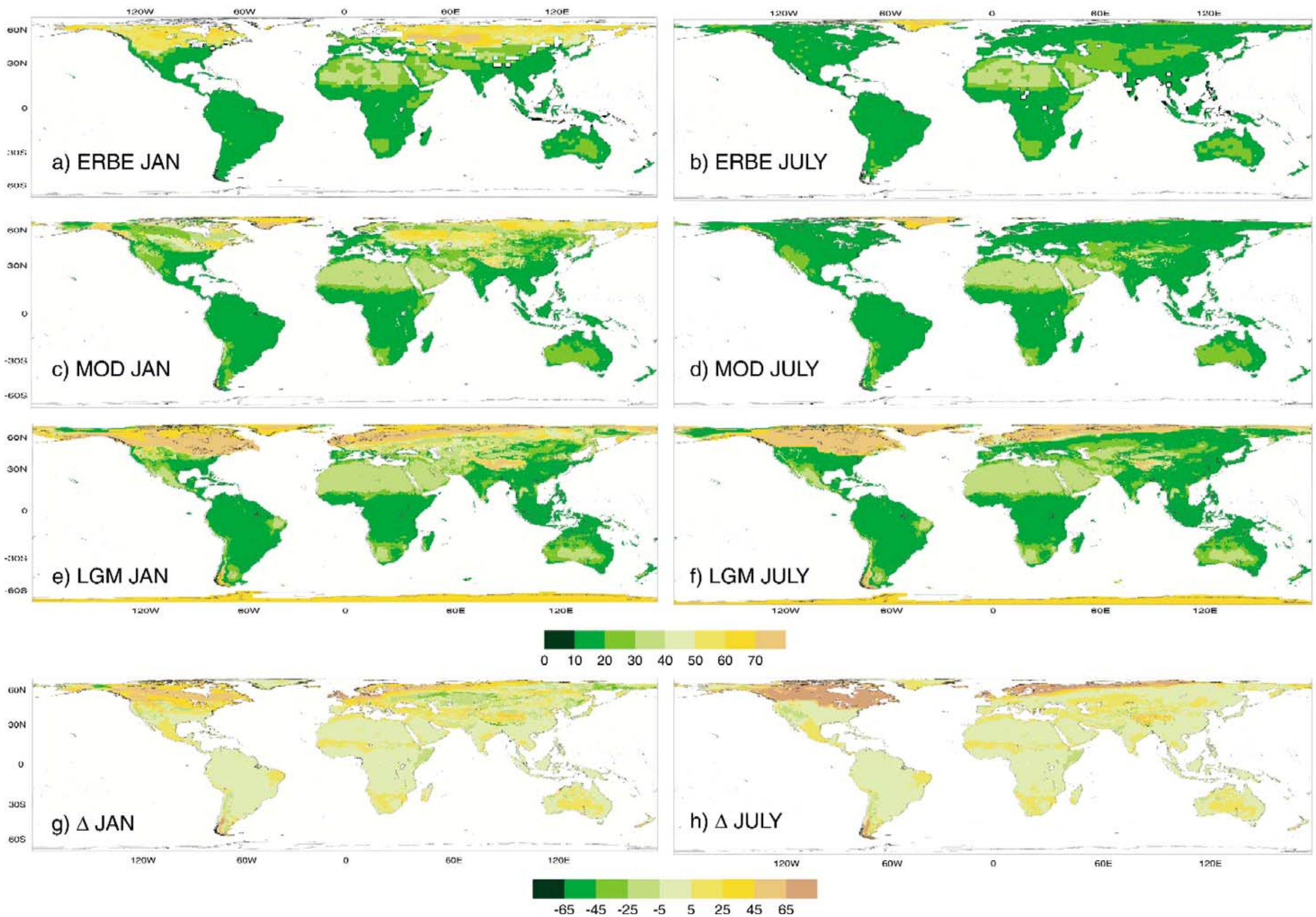

Fig. 9. Observed clear-sky albedo (expressed as a range between 0 and 100) in (a) January and (b) July from ERBE data (Barkstrom et al., 1990) compared to modern albedo fields for (c) January and (d) July simulated from a multi-year climatology from Claquin et al. (submitted for publication). Note that the ERBE data could be influenced by atmospheric aerosols and sub-pixel scale clouds, and this may also explain some discrepancies with the simulated surface albedo at a local-to regional scale. The simulated modern albedo fields can be compared with the simulated LGM albedo fields in (e) January and (f) July, and the simulated change in (g) January and (h) July albedo between the modern-day and LGM (expressed as normalized anomaly).

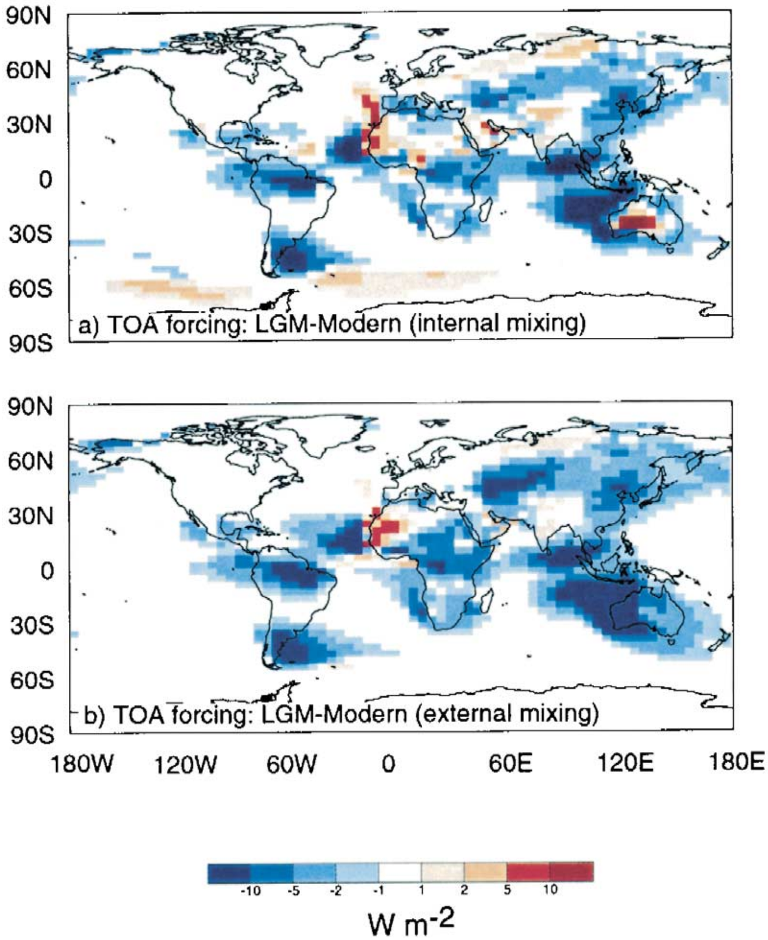

Fig. 10. Simulated changes in radiative forcing (LGM-control simulation) due to changes in dust loading and surface albedo. Results are from the experiments by Claquin et al. (submitted for publication). (a) with internal mixing and (b) with external mixing of minerals in the aerosol.

6.7. The effects of dust on biogeochemical cycles at the LGM

Recognition of the role of dust in providing micronutrients to the open ocean today (e.g. Martin and Fitzwater, 1988; Martin et al., 1994; Coale et al., 1996) has led to speculations that this mechanism may have played a role in CO_2 changes on glacial–interglacial timescales (Martin, 1990; Prentice and Sarnthein, 1993; Martin et al., 1994; Kumar et al., 1995; Broecker and Henderson, 1998; Falkowski et al., 1998). On the basis of the time lag between the reduction in the atmospheric dust loading as recorded by Antarctic ice cores and the subsequent rise in atmospheric CO_2 levels associated with Terminations I and II, Broecker and Henderson (1998) have suggested that dust could have been implicated in CO_2 drawdown during the glacial period. They propose an indirect mechanism whereby the increased delivery of iron (via dust) to the surface waters of the Southern Ocean led to increased N_2 fixation, gradually increasing the global ocean N inventory and thereby increasing phytoplankton productivity. This increased productivity increases the strength of the “biological pump” by which carbon is exported from the surface waters to the deep ocean. When dust fluxes to the Southern Ocean decreased during the deglaciation, the increased N_2-fixation could no longer be supported, resulting in a decrease in marine productivity and an increase in surface water and atmospheric CO_2 levels. However, this mechanism does not explain why the atmospheric CO_2 concentration reached near-glacial levels (205 ppmv) before the dust fluxes increased in the Antarctic ice cores around 65 ka.

The potential effect of increased dust loadings at the LGM on iron fluxes to the ocean and hence to ocean productivity has been examined in a model

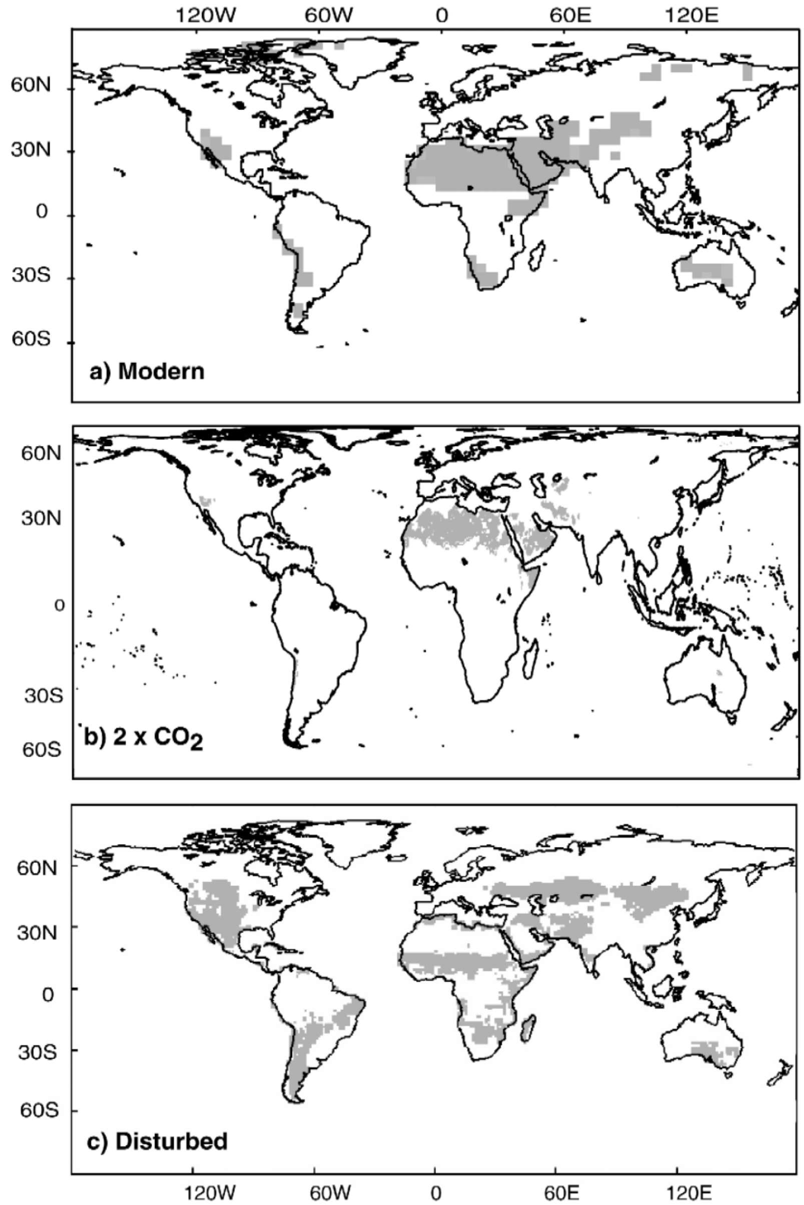

Fig. 11. Dust source areas under modern climate (a) as simulated by Mahowald et al. (1999) compared to dust source areas (b) in response to climate changes due to a doubling of CO_2, and (c) dust source areas due to anthropogenic disturbance of the land surface (Tegen and Fung, 1995). The climate scenario used in (b) was derived from a simulation made with the ECHAM3 atmospheric model (T106) based on the sea-surface temperatures from a couple atmosphere-ocean simulation of the transient response to increasing greenhouse gas concentrations (Bengtsson et al., 1995, 1996; Ohmura et al., 1996; Neilson and Drapek, 1998; Neilson et al., 1998).

experiment (Archer et al., 1998) based on the simulated dust fields from Mahowald et al. (1999). In this experiment, which is made with a 3-D ocean biogeochemistry model (Archer and Johnson, 2000), the LGM dust field had only a slight effect on atmospheric CO_2. Although the simulated dust field more than doubled the iron flux to the ocean, most of this input occurs in regions, which are not iron-limited. The overall effect of iron-fertilization (and the colder sea surface temperatures) resulted in a reduction of CO_2 of only 8 ppmv. So this experiment suggests that biogeochemical effects of dust would be only a minor contributor to the atmospheric CO_2 lowering. On the other hand, ocean biogeochemical models are at an early stage of development and there are numerous open questions, e.g. concerning the pathways of uptake of iron by marine ecosystems (Butler, 1998; Hutchins et al., 1999), the controls on N_2 fixation and denitrification (Falkowski, 1997), and the simulation of tracer transport by the present-generation of ocean circulation models in which 3-D ocean biogeochemistry models are embedded (see, e.g. Orr, 1996, 1999).

7. The role of dust in future climate changes

Changes in the atmospheric burden of dust could have a significant effect on future climates. Simulations of the radiative impacts of dust under present climate conditions (e.g. Tegen et al., 1996) already incorporate the effects of anthropogenically derived dust. Human activities are clearly increasing the atmospheric burden of dust, and the impact of future changes in land-use could substantially increase the importance of the anthropogenic component of the atmospheric dust loading. However, the impact of anthropogenically forced climate changes on the extent and productivity of natural dust sources has not been considered in these calculations.

To investigate the possible impact of anthropogenically forced climate changes on dust source areas, independent of land-use changes, we have made BIOME3 simulations driven by a “snapshot” climatic scenario derived from the ECHAM3 atmospheric model (T106; Wild et al., 1997) forced by sea-surface temperatures from a coupled ocean-atmosphere simulation of the transient response to increasing greenhouse gas concentrations (Bengtsson et al., 1995, 1996; Ohmura et al., 1996; Neilson and Drapek, 1998; Neilson et al., 1998). The snapshot corresponds to the decade of CO_2 doubling in the applied CO_2 scenario. These simulations (Fig. 11b), which include physiological CO_2 effects as well as climatic effects, show substantial reductions in the area of desert and semi-desert vegetation and a reduction in potential dust source regions of about 20% compared to modern (Fig. 11a). This simulation does not take into account anthropogenic land-use changes, which may have a large impact of opposite sign to the climate-induced changes, e.g. in the overexploited environments of the Sahel region (Fig. 11c). However, the reduction in the potential natural source areas for dust deflation, which will be amplified when expressed in terms of the reduction in emissions, is sufficiently large to have a significant radiative impact on climate and possibly a further, biogeochemically mediated effect on atmospheric CO_2—both representing positive feedbacks in the climate system that have not been considered in climate change assessments up to now.

8. Conclusions and future directions for research

The incorporation of dust deflation, transport and deposition processes in Earth-system models is just beginning. In the context of past climates, full coupling between dust cycle models and atmospheric models and ocean models is required in order to assess the relative importance or unimportance of the many interactions and feedbacks that have been postulated to involve dust. Improving the physical and biological realism of ocean biogeochemistry models will be a key element of this task. The use of dynamic vegetation schemes (e.g. Foley et al., 1986, 1998; Sitch et al., submitted for publication), either offline or fully integrated within the architecture of the atmospheric model, to predict the intra-annual variations in vegetation cover that controls short-term (daily) variations in dust emissions will also be important.

The LGM provides an excellent first test for coupled Earth-system models incorporating the dust cycle. However, a correct simulation of the LGM dust cycle is not sufficient to guarantee that such models are capable of reproducing the rapid changes in atmospheric dust loading shown by ice-core and other palaeoenvironmental records of dust, or the time sequence of changes in different records. Equilibrium simulations of other key times will certainly be required as the boundary conditions varied substantially during the course of the glaciation and deglaciation. In the end, the richness of the Greenland and Antarctic ice-core records will provide powerful logical constraints on the causal mechanisms involving dust and other measured quantities including atmospheric CO_2.

Spatially extensive data sets documenting the observed changes in dust accumulation before, at and since the LGM will be required in order to test the model simulations of the dust cycle. The DIRTMAP (Dust Indicators and Records of Terrestrial and Marine Palaeoenvironments) data base (Kohfeld and Harrison, this volume), which was an essential component of the model testing strategy used in Mahowald et al. (1999) and Claquin et al. (submitted for publication), should be further enhanced by the incorporation of information on dust accumulation rates in terrestrial environments. The challenge to the loess community is to generate a quantitative, global data set of aeolian dust accumulation rates and mineralogy, in the first instance for the LGM but subsequently for other key times during the glaciation and deglaciation.

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