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# Global connections between aeolian dust, climate and ocean biogeochemistry at the present day and at the last glacial maximum

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1 **Global connections between aeolian dust, climate and ocean biogeochemistry at the present**  
2 **day and at the last glacial maximum.**

3

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29

30 **Abstract**

31 Palaeo-dust records in sediments and ice cores show that wind-borne mineral aerosol ('dust') is  
32 strongly linked with climate state. During glacial climate stages, for example, the world was  
33 much dustier, with dust fluxes two to five times greater than in interglacial stages. However, the  
34 influence of dust on climate remains a poorly quantified and actively changing element of the  
35 Earth's climate system. Dust can influence climate directly, by the scattering and absorption of  
36 solar and terrestrial radiation, and indirectly, by modifying cloud properties. Dust transported to  
37 the oceans can also affect climate via ocean fertilization in those regions of the world's oceans  
38 where macronutrients like nitrate are abundant but primary production and nitrogen fixation are  
39 limited by iron scarcity. Dust containing iron, as fine-grained iron oxides/oxyhydroxides and/or  
40 within clay minerals, and other essential micronutrients (e.g. silica) may modulate the uptake of  
41 carbon in marine ecosystems and, in turn, the atmospheric concentration of CO<sub>2</sub>. Here, in order  
42 to critically examine past fluxes and possible climate impacts of dust in general and iron-bearing  
43 dust in particular, we consider present day sources and properties of dust, synthesise available  
44 records of dust deposition at the last glacial maximum (LGM); evaluate the evidence for changes  
45 in ocean palaeo-productivity associated with, and possibly caused by, changes in aeolian flux to  
46 the oceans at the LGM; and consider the radiative forcing effects of increased LGM dust  
47 loadings.

48

49 **Keywords:** aerosols, dust, climate change, palaeoclimatology, radiative forcing, iron  
50 fertilisation.

51 1. *Introduction*

52 Palaeo-dust records show that wind-borne mineral aerosol (here referred to as 'dust') is strongly  
53 linked with climate state. Studies of dust in sediments and ice cores show that dust  
54 concentrations and fluxes have changed greatly in association with changes in climate, for  
55 example, in the transitions from glacial to interglacial regimes. During glacial climate stages, the  
56 world was much dustier, with dust fluxes two to five times greater than in interglacial stages (e.g.  
57 Kohfeld and Harrison, 2001). However, the influence of dust on climate remains a poorly  
58 quantified and actively changing element of the Earth's climate system. We know from present-  
59 day studies that dust can influence climate directly, by changing the radiative properties of the  
60 atmosphere through the scattering and absorption of solar and terrestrial radiation, and indirectly,  
61 by acting as ice nuclei (Sassen et al., 2003) and modifying cloud properties which, in turn, can  
62 impact both the radiative balance of the Earth and the hydrological cycle (Arimoto, 2001). Dust  
63 transported to the oceans can also affect climate indirectly, by supplying elements such as Fe, an  
64 essential micronutrient in enzymes essential to photosynthesis (Martin et al., 1991);  
65 phytoplankton in about 40% of the world ocean are Fe-limited (Moore et al., 2002). Thus, dust  
66 delivery to the ocean can modulate the uptake of carbon in marine ecosystems and, in turn, the  
67 atmospheric concentration of CO<sub>2</sub>.

68

69 Dust not only can affect climate, but the generation and transport of dust is itself extremely  
70 sensitive to climate. At the present day, the most obvious link is with aridity – the globally  
71 dominant sources of dust are all located in arid or semi-arid regions. Global model estimates of  
72 present-day dust mobilization rates are poorly constrained, reflecting the scarcity of temporal and  
73 spatial data coverage, and range between ~ 1 and 3.5 Pg yr<sup>-1</sup> (e.g. Tanaka and Chiba, 2006;  
74 Zender et al., 2003; Engelstaedter et al., 2006); optimized multi-model estimates yield a range of  
75 1.5 to 2.6 Pg yr<sup>-1</sup> (Cakmur et al., 2006). Estimates of deposition to the oceans range from about  
76 0.3 to 2 Pg yr<sup>-1</sup> (Zender et al., 2004; Mahowald et al., 2005). Human activities can have an  
77 impact on dust mobilization but here too estimates are poorly constrained and range from 0 to 50  
78 % of global dust emissions (e.g. Tegen et al., 2004; Mahowald et al., 2004; Yoshioka et al.,  
79 2005).

80

81 As recorded in sediments and ice cores, there have been large and systematic variations in dust  
82 loading in the past. The large changes in dust emissions and transport seen from the palaeo-dust  
83 record may reflect a variety of processes: changes in sources and source conditions; changes in  
84 vegetative cover; sub-aerial erosion of emergent continental shelves; deflation from periglacial  
85 deposits; variations in wind speed and gustiness; changed wind patterns linking sources to  
86 deposition areas; changes in deposition along the dust transport path. In order to interpret the  
87 palaeo-record and to anticipate future changes in climate, we need to have a better understanding  
88 of the factors that affect dust mobilization and its subsequent climatic impacts.

89

90 The complex linkages between climate and the dust generation-transport process on glacial-  
91 interglacial time scales is illustrated in figure 1. There are many processes that can affect the dust  
92 cycle either directly or indirectly under any specific climate scenario. The relative importance of  
93 any specific process can change dramatically when shifting from one climate state to another –  
94 e.g., during glacial stages, low sea levels expose coastal sediments and glacial action can produce  
95 unvegetated, unconsolidated deposits all of which can become major sources of dust. A major  
96 driver in the Earth's climate system is atmospheric CO<sub>2</sub>. Figure 1 shows how dust transport to the  
97 oceans could conceivably modify ocean productivity and atmospheric CO<sub>2</sub> and the consequent  
98 impacts on other climate variables including the dust cycle. One major objective of  
99 palaeoclimatology is to use the dust record in ice cores and sediments, coupled with our  
100 knowledge of Earth surface processes today, to improve our understanding of these linkages in  
101 the current climate state and to use this information to better interpret the Earth's past climate  
102 history.

103

104

105 For the future, dust cycle models predict large changes in aeolian transport from the continents to  
106 the oceans over coming centuries, in response to anthropogenic climate change (e.g. Mahowald  
107 et al., 2006). In order to anticipate the effects of future changes in dust emissions, it is important  
108 to identify the impacts of past and present-day dust flux changes in terms of feedbacks with  
109 regard to radiative forcing of dust aerosols and any ocean fertilization and resultant productivity-  
110 driven changes in atmospheric CO<sub>2</sub>.

111

112 Here, in order to examine past fluxes and possible climate impacts of dust in general and iron-  
113 bearing dust in particular, we synthesise available records of dust deposition at the last glacial  
114 maximum (LGM); evaluate the evidence for changes in ocean palaeo-productivity associated  
115 with, and possibly caused by, changes in aeolian flux to the oceans at the LGM; and consider the  
116 radiative forcing effects of increased LGM dust loadings. We begin by presenting an overview of  
117 global dust processes and the state of our knowledge about these processes.

118

119 Figure 1

120

## 121 2. Present-Day Global Dust Processes and Distributions

### 122 2.1 Dust source processes.

123 Dust mobilization is a highly complex process that is a function of many atmospheric, soil, and  
124 terrain properties (Goudie, 2008). Consequently the detachment, entrainment and transport of  
125 dust are highly variable both spatially and temporally. Dust generation is mostly associated with  
126 strong and/or gusty winds with the emitted flux varying as the third or fourth power of wind  
127 speed. The necessary wind fields to uplift dust can be generated by a wide range of local  
128 meteorological conditions. In particular, surface heat fluxes that induce convective activity can  
129 lead to dust generation (Engelstaedter and Washington, 2007), e.g., haboobs (Williams, 2008).

Deleted: ranging from synoptic-scale systems

Deleted: to smaller scale convective processes

130 Terrain features can strongly modulate emission rates (Washington et al., 2005; Koren et al.,  
131 2007). Particle detachment is inhibited by any element which increases surface roughness, e.g.  
132 vegetation, snow cover and surface armouring (by pebbles, stones or by salt crusts). Soil  
133 moisture plays a critical role by affecting particle cohesion along with cementation by salt  
134 content and soil structure. Dust emission thus tends to be favoured in arid or semi-arid areas,  
135 with annual rainfall  $< \sim 250$  mm p.a. (Prospero et al., 2002; Washington et al., 2003), little or no  
136 vegetation cover and strong, long-fetch winds.

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137 Many of the most intense present-day dust sources are associated with deep alluvial deposits laid  
138 down in the Pleistocene and Holocene (Prospero et al., 2002). Direct detachment and  
139 entrainment by the wind of particles (especially fine particles,  $< 20 \mu\text{m}$ ,  $\text{PM}_{20}$ ) appears  
140 subordinate to the effects of saltation (fig. 2); the impacts of sand-sized grains and/or aggregates  
141 on their downward trajectory cause 'sandblasting' - the emission of finer dust grains from the  
142 surface and/or from the original saltating aggregate (e.g. Gillette, 1978; Shao et al., 1993; Sow et

146 al., 2009). Thus the dust mobilization process and the resulting properties of the deflated dust can  
147 be extremely sensitive to a wide range of environmental variables (including, for example, land  
148 use change (Tegen et al., 1995)). The complexity of these processes potentially makes dust a  
149 good proxy for climate variability but also makes the dust record difficult to interpret.

150

151 Figure 2.

152

### 153 2.2. Global dust sources and transport.

154 Much of our knowledge of the large-scale transports of dust is based on remotely sensed data,  
155 primarily from satellites. While there is an extensive literature of the directly measured  
156 concentrations and physical properties of dust, these are mostly too isolated in time and space to  
157 provide a coherent picture of dust transports. However, such measurements, when coupled with  
158 remote sensing data, can provide a compelling and informative picture of global dust transport.  
159 Satellites have come to play a dominant role in aerosol research (Kaufman et al., 2002). At  
160 present there is a large variety of sensors and platforms that provide information on aerosol  
161 concentrations and on some aspects of their physical properties - e.g., size, radiative absorption,  
162 polarization (Dubovik et al, 2008) - and also on the altitude distribution of these properties (e.g.,  
163 Liu et al., 2008).

164

165 Satellite measurements of aerosol optical depth (AOD) with NOAA Advanced Very High  
166 Resolution Radiometer (AVHRR) provide a global picture of aerosol transport over the oceans  
167 (Fig. 3a; Husar et al., 1997). Over land, the most active dust source regions are detected with the  
168 Total Ozone Mapping Spectrometer (TOMS) absorbing aerosol product (Fig. 3b; Prospero et al.,  
169 2002). More recently, Shepanski et al. (2007) identified the regions where dust source is  
170 activated using three different infrared channels of the Meteosat Second Generation (MSG)  
171 instrument. This analysis is based on the frequency of dust generation but does not allow to  
172 retrieve the magnitude of the flux resulting from the episodes. Analyzing dust emission flux or  
173 evaluating dust mass requires a careful analysis of dust optical depth retrieved over deserts by  
174 the MISR (Multiangle Imaging SpectroRadiometer) instrument, together with information on the  
175 vertical distribution of dust retrieved from CALIPSO (Cloud-Aerosol Lidar and Infrared  
176 Pathfinder Satellite Observation). Shepanski et al. (2009) were able to retrieve information on

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178 [the diurnal cycle of dust emissions over Western Africa. Sixty five percents of dust source](#)  
179 [activation was found to occur in early morning from 0600–0900 UTC, well before the hours](#)  
180 [when the surface latent heat fluxes become important.](#)

181

182 It is clear from Figure 3 that at the present day the northern hemisphere is far more ‘dusty’ than  
183 the southern hemisphere. This statement is based largely on *in situ* aerosol measurements made  
184 in different ocean regions (e.g. Duce et al., 1991; Prospero, 1996a; Prospero et al., 1989) and on  
185 satellite products such as shown in Figure 3. Satellite AOD distributions (fig. 3a) show large  
186 areas of high AOD over the oceans; the distribution of these plume-like features is clearly linked  
187 to continental sources. The largest and most persistent plumes (fig. 3a) emanate from arid  
188 regions in North Africa, the Middle East and Asia; the aerosols in these plume regions typically  
189 contain high concentrations of dust (e.g., Husar et al., 1997; Duce et al., 1991; Prospero, 1996;  
190 Prospero et al., 1989). These plumes can be directly linked to dust sources on the continents  
191 using the TOMS satellite (fig. 3b), one of the few sensors that can effectively sense aerosols over  
192 land (Herman et al., 1997).

193

194 It is notable that the major source regions (Fig. 3a and b; Table 1) are concentrated within a  
195 broad ‘dust belt’ that extends from the west coast of North Africa eastward to the Pacific coast of  
196 China. The timing of maximum dust emissions varies across this dust belt. In North Africa,  
197 maximum dust transport occurs in boreal summer, with dust transported across the  
198 Mediterranean to Europe and the Middle East (Moulin et al., 1998) and across the Atlantic to the  
199 Caribbean (Prospero and Lamb, 2003), the southeastern United States (Perry et al., 1997;  
200 Prospero, 1999), and the midlatitude western North Atlantic (Arimoto et al., 1995). There is also  
201 considerable transport in the winter with dust carried into South America (Prospero, 1981). In the  
202 Middle East, activity peaks in the late spring/summer and is at a minimum in the winter. Over  
203 the Indian subcontinent, activity peaks in the spring and decreases in the summer with the onset  
204 of the southwest monsoon. In Asia, dust activity peaks strongly in the boreal spring.

205

206 Of the major dust sources, North Africa is by far the strongest, contributing ~ 40-60 % of global  
207 dust emissions (e.g. Prospero, 1996b; Ginoux, 2001). For example, the so-called Sahara–Sahel  
208 ‘dust corridor’ extends over ~ 4000 km from Chad to Mauritania and emits very large volumes



209 of dust over the Atlantic Ocean (Moreno et al., 2006). Within this broad ‘corridor’, specific  
210 sources stand out as ‘hot spots’ of activity. Most notable is the Bodele Depression, Chad, the  
211 world’s most persistently intense dust source at present day, responsible for an estimated ~ 15-20  
212 % of North African emissions (Washington et al., 2006). While the Bodele stands out as a unique  
213 source on a global scale, much of North Africa serves as an effective dust source. Indeed, much  
214 of North Africa is frequently covered by a dust pall during much of the year. A large region of  
215 intense and frequent dust activity lies in Mali, Mauritania and southern Algeria (Prospero et al.,  
216 2002; Goudie and Middleton, 2006). Many of the most active sources are associated with  
217 drainage systems from the Ahaggar Mountains. This region is probably more representative  
218 generally of arid region dust sources and source processes, with many types of sources  
219 contributing to the dust emissions: i.e., old alluvial deposits, playas, dried lakes and wadis.  
220 Another area of dust activity lies in Tunisia and Northeast Algeria, greatest activity occurring in  
221 association with an extensive system of salt lakes and dry lakes in the lowlands south of the Tell  
222 Atlas Mountains. Major dust activity also occurs in the eastern Libyan desert in a region marked  
223 with an interconnected a chain of wadis (Wadi al-Farigh, Wadi al Hamim). Other active sites are  
224 found in Egypt, Sudan and the flanks of the Ethiopian Highlands.

225

226 The other major source regions of the Northern Hemisphere are the Arabian peninsula (~10-20  
227 % of global emissions, Table 1) and central and eastern Asia (~15-20 %). For the Arabian  
228 peninsula, two major active areas extend along the eastern side of the peninsula along the Persian  
229 Gulf, and along the Oman coastal margins (fig. 3), comprising coastal sabkhas and dissected  
230 wadi systems, respectively (Goudie and Middleton, 2006). The lower Tigris-Euphrates basin,  
231 from north of Baghdad to the Persian Gulf, also provides a summer peak in dust activity, sourced  
232 from an immense alluvial plain. Very high concentrations of dust and pollution-derived species  
233 are found over the Arabian Sea early in the year (fig. 3a) carried by the winter monsoon and the  
234 action of systems moving from the North (Verma et al., 2007). Dust is carried from sources in  
235 the eastern Sahara and in northwest India (including the Thar Desert) and Pakistan.

236 Concentrations of dust and pollutant species drop sharply across the inter-tropical convergence  
237 zone and with the onset of the southwest (summer) monsoon (Savoie et al., 1987; Krishnamurti  
238 et al., 1998) and demonstrate the effectiveness of the inter-tropical convergence zone in blocking  
239 the transport of aerosols to the southern hemisphere.

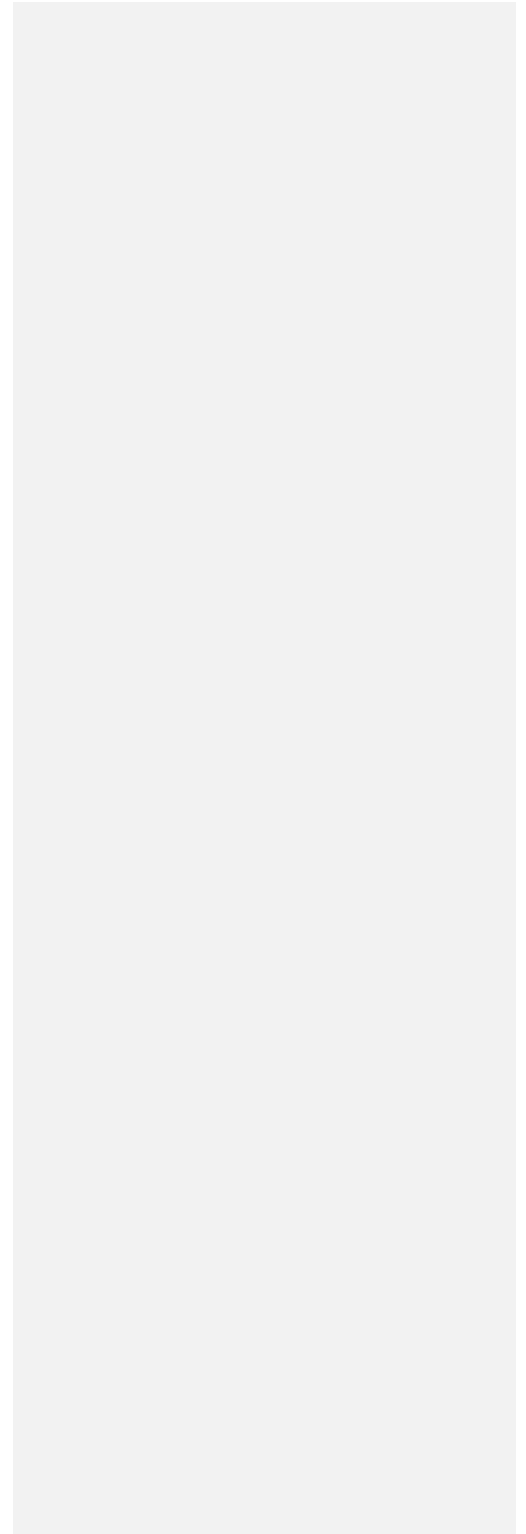
240

241 In eastern Asia, dust sources - mostly in China - are extremely active in the boreal spring. Very  
242 large quantities of dust (often mixed with high concentrations of pollutants) are carried out of  
243 China eastward, across the coastal waters, and over Korea and Japan where the dust is such a  
244 persistent phenomenon, noted in many centuries of historical records, that it is given a name,  
245 'kosa' – yellow dust (Kar and Takeuchi, 2004). Asian dust is routinely observed in relatively  
246 high concentrations at stations in the central North Pacific (Prospero et al., 1989). The strong  
247 seasonal dust cycle and the trans-Pacific extent of the transport can be seen in figure 3a.  
248 Although there are many active dust sources in China (Prospero et al., 2002; Goudie and  
249 Middleton, 2006; Wang et al., 2004), several stand out in satellite images: the Tarim Pendi basin  
250 (containing the Taklimakan Desert; central Inner Mongolia Plateau and the Gobi Desert); the  
251 Hexi Corridor and the Loess Plateau; and western Inner Mongolian Plateau. The Tarim Pendi  
252 basin, which receives very little rainfall (50-100 mm a year along its periphery and ~ 10 mm in  
253 the central regions), is most active from February through May. The Gobi desert, located east of  
254 the Tarim Basin on the Mongolian Plateau, is a major dust source at present. Every spring,  
255 spectacular dust storms can be seen generated over the Gobi; similar to many dust events in  
256 North Africa, the most active sources cover a relatively small area of the surface in effect acting  
257 as 'point' sources. Gobi dust outbreaks can be followed n satellite imagery as they move across  
258 the Pacific to North America (Husar et al., 2001). The Loess Plateau extends over ~440,000 km<sup>2</sup>  
259 and is covered with loess with depths ranging from typically ~100 m to as much as 350 m  
260 (Maher et al., 2008). The Loess Plateau is unique in the global dust picture in that it is the only  
261 loess region that serves as a highly active dust source at the present day. These loessic soils are  
262 now susceptible to enhanced deflation because of the intense agricultural activity in this region  
263 (Wang et al., 2004). While various estimates exist as to the quantitative effects of human  
264 activities in dust mobilization (*see Section 1*), it is clear that of all the major dust sources, the  
265 Asian sources are those where humans could be having a major effect on emissions, although  
266 this is still a matter of considerable debate (Wang et al., 2006).

267

268 Figure 3 a and b

269



270 For the southern hemisphere, contemporary dust sources have been mapped by Prospero et al.  
271 (2002) using TOMS data. Australia, South America and southern Africa each have localized  
272 areas of dust emissions. However, there is little evidence of major and sustained dust activity in  
273 these regions. Although there are few measurements of aerosols over the southern oceans, these  
274 show much lower concentrations than over the northern oceans (e.g., Prospero et al., 1989; Baker  
275 et al., 2006; Wagener et al., 2008). Measurements of AOD (e.g. Fig. 3a) made by a wide variety  
276 of satellite systems (Zhu et al., 2007) fail to detect substantial aerosol plume effects that can be  
277 attributed to dust sources. This may partly reflect some under-estimation by TOMS due to  
278 difficulties of dust detection in cloudy conditions (e.g. Gassó and Stein, 2007). However,  
279 compilations of satellite aerosol event imagery contain many hundreds of examples of dust  
280 plumes, the overwhelming majority of which are linked to Northern Hemisphere sources (see,  
281 for example: <http://visibleearth.nasa.gov/>; <http://rapidfire.sci.gsfc.nasa.gov/gallery/>). In stark  
282 contrast, model results (IPCC, 2001a; Li et al., 2008) often identify emission of large dust  
283 plumes from Australia (comparable to those sourced from NW Africa). Ground-truthing of  
284 either the remotely sensed or modelled data is severely constrained by the lack of modern dust  
285 data. For example, while the ModelE dust component of the GISS Global Circulation Model  
286 (Cakmur et al., 2006) is constrained by data from 96 individual sites, only 18 of those are in the  
287 southern hemisphere (and of these, 13 were measured during a historic low for dust emissions  
288 from Australia (Mackie et al., 2008). The main Australian source is the Lake Eyre basin (Bullard  
289 et al., 2008), followed by the Murray-Darling basin, both comprising extensive river systems,  
290 floodplains and dunefields (Bullard and McTainsh, 2003). There are two main dust transport  
291 paths, one to the northwest and one to the southeast (Bowler, 1976). Although the southeast path  
292 is often described as crossing the Tasman Sea (and Australian dust is often reported in New  
293 Zealand, e.g. Marx et al., 2005), individual events are variable and can contribute sediment  
294 further south, to the Southern Ocean and southwest Pacific (McGowan et al., 2000; McTainsh et  
295 al., 2005; Shao et al., 2007). Iron is a conspicuous component of Australian soils and dust storms  
296 (Taylor et al., 1983; McTainsh, 1989; Mackie et al., 2008) but the concentrations and mineralogy  
297 of Antipodean long range transport (LRT) dusts are poorly known. From magnetic  
298 measurements, modern dusts from eastern Australia and the palaeo-dust contained within  
299 Tasman Sea sediments contain fine-grained haematite/goethite, which varies directly with the  
300 concentration of the mineral dust fraction (Hesse, 1997). New Zealand appears to be an

301 insignificant source of dust at the present day (Prospero et al, 2002), unsurprising as prevailing  
302 trans-Tasman winds are to the east (McGowan et al., 2000).

303

304 Although southern Africa has some significant dust sources (notably the Etosha, Namibia and  
305 Magkadikgadi, NE Botswana, pans; Prospero et al., 1992; Washington et al. 2003), it is not a  
306 major exporter of dust. Iron-rich South African dust that enters the ocean is largely (> 90 %)   
307 deposited in the S. Indian Ocean (e.g. Piketh et al., 2000) along latitude ~ 35 °E with only ~ 4 %  
308 entering the S. Atlantic (Garstang et al., 1996) Some work (Stuut et al. 2005) points to deflation  
309 of silt-rich (> 20 µm) dust to the southwest into the South Atlantic, but it is unknown if this is  
310 then transported into the higher latitudes.

311 For South America, the southern region is spanned by a continuous band of arid or semi-arid  
312 terrains extending from the Peruvian to the Patagonian coast (the 'Diagonal àrida'/*arid*  
313 *diagonal*). Based on TOMS data, Prospero et al. (2002) identified three main dust sources in  
314 southern S. America which overlap most of this *arid diagonal*: the Puna (22-26 °S)/Altiplano  
315 (15-22 °S) Plateau, central-west Argentina (27-35 °S) and Patagonia (39-52 °S). Preferential  
316 emission sources appear linked to large dry lakes (Puna/Altiplano), thousands of small enclosed  
317 basins (Patagonia) and extensive alluvial fans (central-west Argentina). Patagonia is strongly  
318 influenced by the southern hemisphere westerlies. Patagonian-sourced dust is estimated to supply  
319 ~90 % (~ 30 g m<sup>-2</sup> a<sup>-1</sup>) of the present day sediment flux to the SW Atlantic from S. America  
320 (Gaiero et al. 2003). In the high arid plateau (Puna/Altiplano) of the central Andes, the climate is  
321 mostly governed by the subtropical jet stream and dust emitted from this area can be directly  
322 injected into the tropospheric winds. Both wind systems can produce eastward dust transport and  
323 deposition to proximal areas like the Pampas and the nearby Atlantic Ocean or, distally, to the  
324 Southern Ocean or the East Antarctic Plateau. The characteristics of present day Patagonian dust  
325 are relatively well known (Gaiero et al., 2003; Gaiero et al., 2004; Gaiero et al., 2007). Clay-size  
326 fractions are dominated by the clay mineral, smectite. The chemical/isotopic composition of dust  
327 reflects the contribution of Jurassic and Quaternary explosive volcanism (Gaiero et al., 2007).  
328 Dust deposition rates vary from ~5 g m<sup>-2</sup> a<sup>-1</sup> at the Patagonian coast, ~28 g m<sup>-2</sup> a<sup>-1</sup> (Gaiero et al.,  
329 in prep.) at the Central Pampa (33 °S); and ~26 g m<sup>-2</sup> a<sup>-1</sup> for the southern Pampa (39 °S) regions  
330 (Ramsperberger et al., 1998; Gaiero et al., 2003).

331

332 Major dust sources require not only the operation of efficient dust deflation and transport  
333 processes but also a replenishment of dust supply, through active weathering and comminution  
334 of parent materials. TOMS (fig. 3b) shows a clear association of dust sources with specific types  
335 of terrains and environments. All major sources in figure 2b are located in arid regions (annual  
336 rainfall  $< \sim 200\text{--}250$  mm) and centered over topographical lows or on lands adjacent to strong  
337 topographical highs. Although the source regions themselves are arid or hyper-arid, fluvial action  
338 is evident everywhere by the presence of ephemeral rivers and streams, alluvial fans, playas, and  
339 saline lakes. Furthermore, even presently arid source areas have had a relatively recent pluvial  
340 history. The association of dust sources with water features is consistent with our understanding  
341 of the processes involved in the production of fine particles through weathering (Pye, 1989;  
342 Smalley et al., 2005). Fluvial and chemical weathering processes appear much more efficient in  
343 the production of small particles (i.e., particles  $< \sim 10$   $\mu\text{m}$  diameter) than are aeolian processes  
344 (i.e. grinding and impaction). Subsequently, through fluvial action, small particles are carried to  
345 depositional basins or alluvial plains where, after drying, they are subject to deflation by wind.  
346 Thus, precipitation totals and timing are important both in the (low-lying) area of emission and in  
347 any adjacent, dust-supplying highlands. TOMS shows a remarkably consistent association of  
348 dust sources with playas and associated terrain environments [Prospero et al., 2002], such as the  
349 alluvial fans that ring the basins in which the playas are found (Reheis et al., 1995). The playa is  
350 the ultimate receptacle of almost all the fine-grained sediments eroded within a basin over a  
351 geological time period, thus they are storehouses of fine, dry, unconsolidated (and thus wind-  
352 erodible) sediments often in amounts far in excess of production rates in the present climate. An  
353 important characteristic of other major dust sources is the presence of deep and extensive alluvial  
354 deposits. During pluvial phases, these basins were flooded and thick layers of sediment were  
355 deposited which are now exposed. Many of the most active TOMS sources were flooded during  
356 the Pleistocene - Holocene. The prime example is the Bodele Depression in Chad, the largest  
357 source of long-range dust in the world (Warren et al., 2007).

358 Thus, the most productive dust-source environments at the present day are the result of the  
359 complex interplay between climate, terrain and geological history. In order to understand the  
360 linkages between climate, dust production, and transport, we must understand in detail how these  
361 features interact.

362

363 Over the past decade, global circulation models and chemical transport models have come to  
364 play an important role in understanding dust-climate relationships. At present, there are over a  
365 dozen models that simulate the dust cycle, represent its optical properties and predict dry and wet  
366 deposition over the global domain (Textor et al., 2006). However, the accuracy of global aerosol  
367 models is limited by uncertainties in aerosol emission source characteristics, in our knowledge of  
368 atmospheric transport and removal processes and also errors in the meteorological data  
369 assimilated into the model computations. Thus there can be large differences among the outputs  
370 of the various models and, in the absence of aerosol measurements over large areas of the Earth,  
371 it is difficult to select the "best" model. Regardless of these shortcomings, most of the currently  
372 reported dust budget values are based on transport models (e.g. Werner et al., 2002; Luo et al.,  
373 2003; Zender et al., 2003 and 2004; Ginoux et al., 2004; Mahowald et al., 2005; Cakmur et al.,  
374 2006; Miller et al., 2006; Tanaka and Chiba, 2006; Li et al., 2008).

375

376 *3. Dust impacts on climate.*

377 *3.1 Dust as a source of bioavailable iron*

378 Dust may play a critical, indirect role in climate change, through biogeochemical interactions  
379 (e.g. via iron or other essential element fertilization of ocean and/or terrestrial ecosystems) and  
380 chemical interactions with ozone and sulphur cycles (Cwiertny et al., 2008; Chin et al., 2000). In  
381 the case of ocean fertilization, dust containing iron, as fine-grained iron oxides/oxyhydroxides  
382 and/or within clay minerals, and other essential micronutrients (e.g. silica) may play a significant  
383 role for those regions of the world's oceans where macronutrients like nitrate are abundant but  
384 primary production and nitrogen fixation is limited by iron scarcity. Hence, as formalised by  
385 Martin et al. (1990, 1991), the efficiency of the 'biological pump' is low and there is significant  
386 potential for enhanced export production and drawdown of atmospheric CO<sub>2</sub>. The Southern  
387 Ocean and the equatorial and northwestern subarctic Pacific are the major high nitrate, low  
388 chlorophyll (HNLC) areas, totalling some 30-50 % of the world ocean (Moore et al., 2004;  
389 Aumont et al., 2003). Diatoms are the key phytoplankton group for formation of algal blooms in  
390 such areas and are strongly iron-limited. Additionally, in tropical and subtropical regions, iron  
391 deficiency may also limit phytoplankton growth in subtropical gyres (Johnson et al., 1997;  
392 Sedwick et al., 2005) and nitrogen fixation by the diazotrophs (Falkowski, 1997; Moore et al.,  
393 2006). Supply of iron-bearing dust may thus directly and indirectly limit primary productivity for

394 large portions of the world ocean (Moore and Doney, 2007), affecting biological carbon export at  
395 the global scale (Moore and Braucher, 2008).

396

397 Consistently higher dissolved iron concentrations ( $>1$  nM) are observed for ocean areas below  
398 major dust plumes (Sarthou et al., 2003). The 'iron hypothesis' (Martin, 1990) suggests that past  
399 increases in aeolian, iron-bearing dust supply to the surface of HNLC ocean waters during  
400 glacial stages relieved iron limitation on primary productivity. The resultant increase in the  
401 export of biogenic carbon from surface to deep ocean waters (the 'biological pump') caused a  
402 decrease in surface ocean  $p\text{CO}_2$  that was balanced by an increased drawdown of  $\text{CO}_2$  from the  
403 atmosphere. The details of this process, and the possible magnitude of its effects on the global  
404 carbon cycle, are the subject of much international debate (e.g. Honda et al., 2006; Maher and  
405 Dennis, 2001; Moore and Doney, 2007).

406

407 The sensitivity of HNLC regions to increased iron supply has been demonstrated by a series of  
408 artificial mesoscale iron fertilization experiments in the HNLC regions (e.g. Boyd et al., 2007;  
409 Coale et al., 2004; Coale et al., 1996). The common outcomes to these experiments were the  
410 expansion of phytoplankton standing stocks, with distinct population shifts to large diatom  
411 species, and resultant drawdown in atmospheric  $\text{CO}_2$  (deBaar et al., 2008; Tsuda et al., 2003). It  
412 has been difficult to quantify export production via subsurface storage of carbon (rather than  
413 subsequent re-release from the ocean mixed layer). For the NW Pacific SEEDS experiment,  
414 drawdown of atmospheric carbon was estimated at  $\sim 150$ - $1500$  C atoms per iron atom added  
415 (Tsuda et al., 2003). The most recent experiment was also the largest (Lohafex, 2009), with 6  
416 tonnes of dissolved iron added to a  $300$  km<sup>2</sup> ocean patch in the southwest Atlantic sector of the  
417 Southern Ocean. However, whilst there was a doubling of (non-diatom) phytoplankton biomass  
418 within the first two weeks, increased grazing pressure of zooplankton prevented further  
419 development of the algal bloom, with only minor amounts of carbon being removed from the  
420 ocean surface layer as a result.

421

422 Based on these magnitudes of export production, and taking into account the rather low fractions  
423 of iron estimated to be bioavailable from mineral dust (e.g. Baker et al., 2006a), modelled  
424 glacial-stage changes in atmospheric  $\text{CO}_2$  resulting from dust-driven iron fertilisation range from

425 ~ 20 % (Bopp et al., 2003) to > 50 % (Watson et al., 2000) of the total interglacial-glacial change  
426 of ~ 100 ppmv. However, extrapolation of the artificial iron enrichment experiments to estimate  
427 past carbon sequestration may be unrealistic given their short duration and use of unfeasibly  
428 large and rapid iron additions, which lead to loss of most (80 – 95 %) of the iron added during  
429 the enrichment experiments (Bowie et al., 2001). More realistic estimates of carbon sequestration  
430 may be derived from analysis of a recent (Nov. 2004 – Feb. 2005), natural phytoplankton bloom  
431 in the Southern Ocean (Blain et al., 2007). This natural event was driven by indirect iron supply  
432 from upwelling of bottom waters (Lefevre and Watson, 1999; Maher and Dennis, 2001), rather  
433 than direct aeolian iron supply.  $p\text{CO}_2$  measurements of the surface waters showed that the bloom  
434 was an important  $\text{CO}_2$  sink, with mean  $p\text{CO}_2$  drawdown 2–3 x higher than that observed for the  
435 artificial Southern Ocean iron experiments, probably reflecting the longer duration of the natural  
436 bloom. Carbon sequestration efficiency was estimated to be at least 10 times higher than that  
437 estimated from the artificial experiments. The higher sequestration efficiency of the natural  
438 bloom has been attributed to its more complete and longer-lasting development, aided by slow  
439 and continuous natural additions of iron (and eventually other required macro-nutrients), from  
440 adjacent deep water supply (Blain et al., 2007). These results indicate first that aeolian supply of  
441 iron to the oceans, both directly and indirectly (i.e. deposited elsewhere and subsequently  
442 dissolved and transported), may have been significant in palaeo-ocean and atmospheric  $\text{CO}_2$   
443 changes. Second, the HNLC oceans may presently be, and previously have been, more sensitive  
444 to aeolian iron supply than has been indicated to date by the artificial experiments. However,  
445 the SOIREE experiment (Boyd and Abraham, 2001) showed that the photosynthetic competency  
446 of iron-limited Southern Ocean phytoplankton increased only when the concentration of  
447 dissolved iron exceeded a threshold value of ~0.2 nM, i.e. ~ 2 x twice the ambient  $[\text{Fe}]_{\text{dissolved}}$   
448 value. Deposition of dust can thus increase the iron inventory of surface waters without  
449 stimulating a phytoplankton bloom (Boyd et al., 2009). Previous reports of correlation between  
450 dust events and remotely observed increases in surface water chlorophyll (e.g. Gabric et al.,  
451 2002; Shaw et al., 2008) need to be evaluated with respect to a physiologically realistic lag time  
452 between the increase in dust/iron supply and phytoplankton response (Boyd et al., 2004). For  
453 example, whilst Shaw et al. (2008) attributed a bloom in waters of the Great Barrier Reef Lagoon  
454 (<50 km from mainland) coastal to a large dust storm (the largest in over 40 years), Mackie  
455 (2009) pointed out that the time ( $\ll 1$  day) between dust deposition and observations of



456 chlorophyll increase precluded dust as a causal agent. In response, Shaw et al. (2009) suggested  
457 that smaller dust events before the large dust storm were responsible. However, in waters so  
458 close to the coast, subject to numerous such small dust events, it is unlikely that iron limitation  
459 would develop in any case.

460

461 An assumption of many of the dust-cycle and iron supply models is that atmospheric input of  
462 bioavailable iron to the surface ocean varies linearly with dust loading. Present day field  
463 collection and dust leaching studies (Sedwick et al., 2007; Baker & Jickells, 2006; Jickells et al.,  
464 2005) indicate that it is critical to identify the nature (mineralogy), source (e.g. size distribution)  
465 and transport history of the dust, in order to evaluate the possible proportion of bioavailable iron  
466 carried within the terrigenous dust loading (Baker and Jickells, 2006; Baker and Croot, 2009). A  
467 diverse range of dissolution experiments has been reported for a range of iron-bearing  
468 aluminosilicates and iron oxides (e.g. Journet et al., 2008), potential source soils (e.g. Visser et  
469 al., 2003) and increasing numbers of natural dust samples (e.g. Spokes et al., 1994; Baker et al.,  
470 2006b; Desboeufs et al., 2005; Mackie et al., 2005). Generally low but spatially variable iron  
471 solubilities appear the norm, i.e. < 0.6 % (e.g. Baker et al., 2006b; Journet et al., 2008). At or  
472 close to source areas, soil mineralogy appears to be the key influence on iron solubility. For  
473 example, for some African sources, higher solubilities (0.3 %) were displayed by iron-bearing,  
474 smectite-rich dust from the central Sahara and lower values (0.1 %) by illite/haematite/goethite-  
475 rich dust from the western Sahara and the Sahel (Journet et al., 2008). Conversely, at areas  
476 remote from major dust sources, and associated with decreased dust concentrations, increased  
477 levels of iron solubility have been reported, and ascribed to progressive particle fining with long  
478 range transport (e.g. Baker and Croot, 2009). However, as discussed below, the particle size of  
479 LRT dust appears to stabilise rather than become progressively finer. Other factors of likely  
480 significance for iron solubility include enhancement of fine particle reactivity during the  
481 transport process, especially related to ligand formation and surface adsorption and/or desorption  
482 processes. Depending on ambient pH, iron-hydroxyl groups at iron oxide surfaces can achieve  
483 negative surface charge by dissociation ( $\text{Fe-OH} \rightarrow \text{FeO}^- + \text{H}^+$ ) or association ( $\text{FeOH} + \text{H}^+ \rightarrow$   
484  $\text{FeOH}^{2+}$ ) of protons (e.g. Cornell and Schwertmann, 2003; Taylor et al., 1983). For the iron  
485 oxides/ hydroxides, the zero point of charge (zpc, i.e. where there are equal concentrations of  
486 surface  $\text{FeO}^-$  and  $\text{FeOH}^{2+}$ ) varies between ~ pH 5 - 8. Similar pH-dependence of surface and

487 inter-layer charge is displayed by the clay minerals, but with much lower zpc values, from < pH  
488 2-3 for smectite to pH 2 – 4.5 for kaolinite . Below the zpc, negatively-charged adsorbates, such  
489 as natural organic matter, can subsequently be retained at or near the mineral surface. Such  
490 adsorption may play a key role both in reducing particle scavenging and increasing iron  
491 solubility in the marine environment. Iron reduction can result from photolysis of  $\text{Fe}^{3+}$  -  
492 complexed with bacterially-produced siderophores (Barbeau et al., 2001) complexes. The  
493 solubilised  $\text{Fe}^{2+}$  could then undergo direct biological uptake, be re-oxidised to  $\text{Fe}^{3+}$  and/or  
494 subjected to subsequent complexation. Recent reports suggest that Fe-reducing compounds, such  
495 as dimethyl sulphide, might interactively be released from iron-deficient phytoplankton  
496 (Johansen and Key, 2006), although the iron compound modelled in this study was ferrihydrite,  
497 the most microcrystalline and reactive of the soil oxyhydroxides. It is clear that large  
498 uncertainties presently remain regarding the biogeochemistry of iron complexation, release,  
499 uptake and scavenging in the marine environment (Baker and Croot, 2009).

500

501 The interpretation of Fe solubility in aerosol samples is complicated at the present day by the  
502 possible presence of Fe derived from anthropogenic sources, in particular, combustion processes,  
503 both industrial and biomass burning. Measurements on Bermuda (Sedwick et al., 2007) show  
504 that air masses from North American air typically contain relatively low concentration of aerosol  
505 iron ( $0.5 \text{ nmol Fe m}^{-3}$ ) but the fractional solubility is very high, 19 %. In contrast dust-laden  
506 African air masses contained relatively high concentrations of aerosol iron ( $27.8 \text{ nmol Fe m}^{-3}$ )  
507 but the fractional solubility was low, only 0.44 %. The net result is that the quantity of soluble Fe  
508 derived from the aerosols was the same in both cases. The finest particle size fraction yielded the  
509 highest solubility for both aerosol types. Model results (Luo et al., 2008) suggest that  
510 combustion-derived iron can account for as much as 50 % of the total iron deposited in some  
511 regions. However, over the open ocean it is usually less than 5 % of the total iron although close  
512 to the East Asian continent in the North Pacific values can be as high as 30 %.

513

514 The observation of increased iron solubility associated with low dust concentrations would imply  
515 proportionally enhanced supply of bioavailable iron to ocean areas distal from dust sources  
516 (notably, the Southern Ocean) and less to high dust flux areas (Baker and Croot, 2009). The  
517 nature of the causal link between dust concentration and solubility requires further evaluation, in

518 order for robust incorporation of varying values of iron solubility of dust into global models of  
519 iron flux (e.g. Fung et al., 2000; Bopp et al., 2003). Most models for dust-iron processes use a  
520 single uniform value for iron solubility from dust but a recent report of 2-3 % solubility of iron  
521 from glacial flour compared to <1 % for arid soils (Schroth et al., 2009) provides a new variable  
522 for models to incorporate. The impacts of ocean acidification, likely to modify aerosol and iron  
523 solubilisation processes, also require evaluation (Breitbarth et al., 2009).

524

525 *3.2 Dust impacts on climate: radiative effects.*

526 *3.2.1 Dust particle size*

527 Dust size distributions in ocean sediments and ice cores are often used as a measure of dust  
528 generation and transport dynamics. The size distributions of dust particles also have a critical  
529 influence on the direct and indirect climate impact of dust (see below) and they are also a major  
530 uncertainty in dust modelling. Most of the long-traveled dust mass is located in the size range  
531 under about 20  $\mu\text{m}$  diameter (e.g. Arimoto et al., 1997), with a mass median diameter of  $\sim 1.5 -$   
532  $3 \mu\text{m}$  (Balkanski et al., 1995; Reid et al., 2003). Particles in this size range can remain suspended  
533 in the troposphere for days to weeks, and be carried over great distances ( $\gg 1000 \text{ km}$ ) before  
534 being removed from the atmosphere by deposition processes. Asian dust, for example, has been  
535 shown to contribute to dust samples spanning western to eastern China (Gao et al., 1992), the  
536 Yellow Sea (Gao et al., 1997), Japan and Korea (e.g. Park et al., 2005), the Pacific Ocean (e.g.  
537 Duce et al., 1980; Rea et al., 1985; Uematsu et al., 1985). More distally, it has a significant  
538 impact on air quality in the western United States (VanCurren and Cahill, 2002); and is a major  
539 source of dust in Greenland ice cores (Bory et al., 2003). North African dust is carried in great  
540 quantities to the Amazon basin (Swap et al., 1992), into the Caribbean (Prospero and Lamb,  
541 2003) and the southeastern (Prospero, 1999), eastern and northeastern US (Perry et al., 1997).  
542 The first challenge for improving data on dust particle size distributions, whether at present day  
543 or for the palaeo-record, is the need for expanded geographic and temporal coverage. The second  
544 requires instrumental convergence. Across the modern and palaeo-data aerosol communities,  
545 particle size tends to be measured using different instruments. For ice and sediment cores, for  
546 example, particle size analyses are most frequently derived from instruments using the Coulter  
547 Principle (such as the Coulter Multisizer). In contrast, the atmospheric community uses a wide  
548 range of instruments that are based on different physical principles, i.e., the mobility of aerosols

549 in electric fields, the optical properties of aerosols, aerosol aerodynamic properties. These  
550 instruments yield widely disparate results even when making measurements on identical dust  
551 aerosol populations, including African dust (e.g. Reid et al., 2003).

552

553 Notwithstanding this instrumental caveat, modern aerosol size data suggest evolution of the  
554 modal diameters for long range transport (LRT, i.e. > 1000 km) dust to a rather invariant range  
555 of ~ 3 to 3.5  $\mu\text{m}$  diameter, in contrast to values obtained in or near source regions, which tend to  
556 cluster at ~ 5  $\mu\text{m}$  diameter (Sviridenkov et al., 1993; Arimoto et al., 1997, Moore et al., 2004;  
557 Maring et al., 2003). For example, dust measured on the Canary Islands (300 km from the N.  
558 African coast) and in Puerto Rico (5000 km to the west) shows very similar particle size  
559 distributions for the < 7  $\mu\text{m}$  size range (fig. 4a). Larger particles (> ~ 7-8  $\mu\text{m}$  diameter) do  
560 display between-event differences, due to their preferential removal during atmospheric  
561 transport.

562

563 Similarly, in surface sediments along a 3600 km-long, west-east transect of the North Pacific, a  
564 transect lying ~ 4000-8000 km downwind from the E. Asian source areas (i.e. distal from any  
565 possible non-aeolian terrigenous sources), the median dust grain size barely changes, at ~ 2.8  $\mu\text{m}$   
566 to 2.4  $\mu\text{m}$  (Janecek, 1985). More recent atmospheric sampling campaigns also suggest a size  
567 range of ~ 1- 3  $\mu\text{m}$  for LRT particles in this region (e.g. Zhao et al., 2003). A simple empirical  
568 model, which sets the vertical velocity of dust particles equal to the Stokes gravitational settling  
569 velocity minus an upward velocity of ~0.33  $\text{cm s}^{-1}$ , accurately predicts changes in dust size  
570 distribution during atmospheric transport (Maring et al., 2003). Thus it appears that some  
571 atmospheric process or processes partially counteract gravitational settling along long transport  
572 pathways. It should also be noted that the transport of giant, >75  $\mu\text{m}$ , grains of dust for up to  
573 10,000 km has been reported (e.g. Pitty, 1968; Betzer et al., 1988, Middleton et al., 2001) and  
574 can not be explained in terms of simple Stokes type settling. The presence of long-lived, long-  
575 travelled dust concentrations within the atmosphere remains an unresolved challenge to our  
576 understanding of LRT processes.

577

578 Such relative observed invariance of dust size also suggests some degree of constancy in dust  
579 mineralogy, morphology and size at the regional scale (e.g. fig. 4a), in turn indicating that source

580 soils or sediments are a factor in controlling dust particle size, in addition to wind speed and/or  
581 in-transport processing (Grini and Zender, 2004). In this scenario, dust properties remain  
582 consistent when the sources also remain constant, but can be expected to change if and when the  
583 source areas change. Wang et al. (2007) report similar conclusions for modern dusts sourced  
584 from NW China. The dust grain size distributions vary from source to source but are generally  
585 trimodal, with the first peak for particles  $> 11 \mu\text{m}$ , the second for particles  $4.7 - 7 \mu\text{m}$ , the third  
586 for sub-micrometre particles ( $< 0.43 \mu\text{m}$ ). These distributions remain essentially the same for any  
587 one source whether under dust storm or non-dust storm conditions. As noted by Wake and  
588 Mayewski (1994), distinctive variations have been observed in the particle number- and mass-  
589 size distributions of dusts collected from different geographic regions, including Europe (e.g.  
590 Junge, 1963), the Sahara (Schutz and Jaenicke, 1974), the Swiss Alps (Wagenbach and Geis,  
591 1987), Colorado and Texas (Patterson and Gillette, 1976), Tajikistan (Gomes and Gillette, 1993)  
592 and Greenland (Steffensen, 1985, 1997).

593 These observational data do not support the wind speed dependence of LRT dust particle size  
594 (fig. 4b) reported experimentally by Alfaro et al. (1998) and by observation by Sow et al. (2009).  
595 These studies indicate enhancement of very fine ( $< 2 \mu\text{m}$ ) particles with increased wind speeds  
596 and surface shear velocity ( $u^*$ ), as a result of 'sandblasting' by saltating grains. *Within* an  
597 erosional event, however, the dust size distribution appears relatively insensitive to measured  
598 changes of  $u^*$ . This apparent conflict may reflect a dissonance between the typically long  $u^*$   
599 averaging time (15 mins) and significantly shorter response times of the dust emission processes  
600 (Sow et al., 2009). Engelstaedter et al. (2007), in contrast, use AERONET data to support the  
601 conventional view that larger particle sizes reflect stronger, or possibly gustier, winds.

602

603 Figure 4.

604

### 605 3.2.2 Dust radiative properties: effective particle size and mineralogy.

606 In terms of radiative effects (RE), atmospheric dust can both scatter and absorb incoming solar  
607 and outgoing long-wave radiation (direct RE), which alters the energy balance of the atmosphere  
608 (IPCC 2001b; IPCC 2007), and can in turn produce changes in atmospheric circulation (Miller  
609 and Tegen, 1998) and cloud cover (Perlwitz and Miller, 2009). The effect of dust absorptivity is  
610 to produce warming of the atmosphere, locally or regionally, and cooling of the surface. The

611 effect of scattering is to produce more widespread cooling, the result of increased upwards  
612 radiance. Dust also plays an important role in cloud microphysics, particularly with regard to  
613 numbers of cloud condensation and ice nuclei, thus influencing the formation and lifetime of  
614 clouds (indirect RE), which in turn influences the planetary albedo (e.g. Spracklen et al., 2008).  
615 Estimates from dust-cycle models of the net top of atmosphere direct RE currently span both  
616 negative and positive values, ranging from  $-0.6$  to  $+0.4 \text{ Wm}^{-2}$  (IPCC 2007). This uncertainty in  
617 both sign and amplitude of dust radiative forcing at the present day reflects both the complexity  
618 of dust distribution (including its vertical distribution, interaction with clouds, the albedo of the  
619 underlying surface) and poor understanding of the scattering/absorptivity of dust particles, in  
620 turn related to poorly known dust concentrations, mineralogy, particle size, and shape (e.g.  
621 Nousiainen, 2009). These latter properties vary regionally, depending on the source materials  
622 (soils, dry lake beds, ephemeral channels etc); and for any population of grains within a parcel of  
623 entrained dust, they also vary interdependently. Mineral dusts typically consist of an internal  
624 mixture of mineral species (and particle morphologies), often including quartz, clay minerals,  
625 iron oxides and calcium carbonate (Claquin et al., 1999), each with their own optical properties  
626 (Sokolik and Toon, 1999). Platy clay mineral particles, for example have different optical  
627 properties than equidimensional silicates (Nousiainen et al., 2009). Recent estimates in radiative  
628 forcing computations indicate that assuming that dust particles are homogenous spheres causes  
629 roughly as large an error as that associated with the uncertainties in refractive index (Kahnert et  
630 al., 2007). The latter has previously been considered the dominant source of uncertainty in dust  
631 optical properties. The radiative behaviour of dust particles also depends on their effective  
632 particle size, i.e. in relation to the wavelength under consideration. Thus, particles with diameters  
633 of around  $0.5 - 1 \mu\text{m}$  are effective in scattering solar radiation (fig. 5), relatively unaffected by  
634 particle shape or surface roughness effects. Such effects become increasingly important for  
635 wavelength-scale and larger particles (Nousiainen, 2009), including therefore those within the  $\sim$   
636  $1.5 - 3 \mu\text{m}$  size range (i.e. the reported mass median diameter for LRT dust, as above).

637

638 Figure 5

639 Another poorly quantified potential control on dust absorptivity is the presence of iron oxides/  
640 oxyhydroxides, as so-called 'free iron', i.e. often as discrete, sub-micrometre particles (figure 6),  
641 or as surface coatings on quartz grains, and, more rarely, iron oxide-clay aggregates (Sokolik and

642 Toon, 1999). Only where soil pH is low will iron oxide grains carry sufficient positive charge,  
643 and/or the clay particles carry sufficient negative charge, to interact with each other (Cornell and  
644 Schwertmann, 2003).

645

646 Figure 6.

647

648 Haematite and goethite occur as minor soil components ( $< \sim 5\%$  by mass) on a global scale,  
649 reflecting the near-surface weathering of  $\text{Fe}^{2+}$ -containing primary (lithogenic) minerals by  
650 hydrolysis and oxidation. Given that they occur in minor concentrations, both these minerals  
651 have a disproportionate radiative impact, displaying large absorbing potential at shorter  
652 wavelengths (Alfaro et al., 2004; Sokolik and Toon, 1999; Balkanski et al., 1997). The  
653 mineralogy and concentration of iron oxides in dusts is dominantly controlled by the source soil  
654 hydrology and mineralogy (Schwertmann and Taylor, 1977), subsequently by any erosional  
655 processes of grain aggregation and/or disaggregation, and finally, by any particle size changes  
656 during transport (e.g. Zhang et al., 2005). Establishing mineralogical linkages between potential  
657 sources and airborne dust is likely to be most robust when based on particle size-specific  
658 comparisons between the clay and fine silt fractions of source soils/ sediments and the fine and  
659 coarse mode of sampled dust (Maher et al., 2009; Sokolik and Toon, 1999). Compared with its  
660 'parent' soil, dust is likely to be biased towards smaller, less dense, and flatter (large surface-  
661 area-to-volume ratio) particles (Nousiainen, 2009). Haematite and goethite tend to be associated  
662 with fine ( $\ll 10\mu\text{m}$ , i.e. fine silt and clay-sized) particles, with long residence times (i.e. days) in  
663 the atmosphere and thus potentially long transport paths.

664 Haematite has been identified as a significant absorber in terms of dust radiative properties  
665 (especially when present as sub-micrometre grains, fig. 6b and c). It is very stable, is the end  
666 product of the dehydroxylation of all iron oxyhydroxides and is thus prevalent in warmer and  
667 drier regions and increasingly limited in occurrence pole-wards of latitudes  $\sim 40^\circ \text{N}$  and S  
668 (Cornell and Schwertmann, 2003). For modern dusts from arid and semi-arid regions (Claquin et  
669 al., 1999), haematite content varies between 0.6 to 3.4 % (by volume). The iron oxyhydroxide,  
670 goethite, is less restricted in its distribution. Although the red coloration of many soils and dusts  
671 often reflects the presence of haematite (highly pigmenting when it occurs as sub-micrometre  
672 grains), goethite is often volumetrically the more important. In the loess and palaeosols of the

673 Chinese Loess Plateau, for example, the wt % of goethite is 2–3 x higher than that of haematite  
674 (Balsam et al., 2007; Ji et al., 2004). Similar dominance of goethite has been reported for a  
675 limited number of dust source samples from the Sahara and the Sahel (Lafon et al., 2006). The  
676 presence of goethite can lead to large differences in the wavelength dependence of the single  
677 scattering albedo (Sokolik and Toon, 1999). The evenness of distribution of goethite in the soil  
678 matrix and resultant dusts is also likely to maximize its absorptivity. Inclusion of goethite in dust  
679 modelling thus seems essential, given its near-ubiquity in source soils.

680

681 Direct RE forcing by dust appears to result in cooling over much of the Earth's surface (fig. 7,  
682 and e.g. Perlwitz et al., 2001; Zhu et al., 2007). Conversely, over bright surfaces (such as  
683 deserts), dust can exert significant warming effects. For example, discrepancies between  
684 measured and modelled outgoing radiation over cloud-free, desert areas of north Africa indicate  
685 that dust may exert a long wave radiation forcing as great as  $50 \text{ Wm}^{-2}$  in the monthly mean at  
686  $1200 \text{ U TC}$  (Haywood et al., 2005). Similarly, for the 'great Indian desert' (Thar) area, warming  
687 of the lower atmosphere from dust absorptivity (greatest during the winter season) has been  
688 estimated at  $0.7\text{--}1.2 \text{ K day}^{-1}$  (Moorthy et al., 2007). Due to direct and indirect RE, dust can  
689 regulate and interact with the heat sources and sinks of monsoonal regions, altering the  
690 monsoonal water budget (Lau and Kim, 2006). A further complication arises from increased  
691 emissions of anthropogenic, carbonaceous aerosols, which have reduced the global annual  
692 average single scattering albedo; aerosol at present day is estimated to be approximately twice as  
693 absorbing as that in preindustrial conditions (Myhre, 2009).

694

695 Given the potential impacts of effective particle size and iron mineralogy, improved modelling of  
696 dust optical properties at solar wavelengths can only be achieved with improved data on present  
697 and past regional variations in aerosol size and shape distributions, and in iron oxide  
698 concentrations, mineralogies and mixing state. Given that haematite and goethite occur mostly in  
699 minor concentrations, their quantification in dusts and sediments requires from any method of  
700 analysis the capacity to analyse to low concentrations (i.e. below the detection levels of x ray  
701 diffraction, for example). Diffuse reflectance spectrometry and environmental magnetic methods  
702 offer potential in this regard, these measurements responding selectively to the iron oxide-



703 specific signals of colour and magnetic properties, respectively (Arimoto et al., 2002; Balsam et  
704 al., 2007; Maher et al., 2007; Watkins et al., 2007).

705

706 Figure 7.

707

708 4. *The palaeo-dust record for the last glacial maximum.*

709 4.1 *Dust fluxes, size distributions, mineralogy.*

710 Glacial/interglacial changes in the flux and particle size of dust have been recorded in ice cores  
711 and sediment records, both continental (windblown loess, lake sediment and peat records) and  
712 marine. Spanning the entire transect from pole to equator to pole, these dust records have shown  
713 that dust flux was higher in both hemispheres during past glacial stages (fig. 8). In light of the  
714 ongoing debates regarding modern dust supply and transport outlined above, what does the  
715 palaeo-dust record represent? Flux changes may record information on changes in the dust  
716 supply, reflecting changes at and/or of source area(s). Source areas may have expanded, through  
717 glaciogenic action (Mahowald and Muhs, 2006), emergence of continental shelves into the sub-  
718 aeral zone (Bigler et al., 2006), and (seasonally-extended) expansion of arid and semi-arid areas  
719 (Werner et al., 2002; Tegen et al., 2002). Particle size changes may reflect changes in transport  
720 and/or source. Many palaeoclimate studies have used dust size as an indicator of wind strength in  
721 source regions (i.e. larger particles indicate stronger palaeo-winds, e.g. Rea 1995) or changes in  
722 the transport pathway (smaller particles indicate longer transport pathways, e.g. Lambert et al.,  
723 2008). Clearly, these arguments only hold if there is no change in the source size distributions,  
724 and no change in the source-sink distance. However, transporting winds may have altered in their  
725 trajectory and/or speed or gustiness (*see section 3.2.1*). Transport paths may have been  
726 effectively extended or shortened by changes in the degree of rain-out en route (Yung et al.,  
727 1996; Ruth et al., 2005; Lambert et al., 2008). Deposition rates on land may have varied in  
728 response to changes in vegetation and resultant trapping of dust by increased surface roughness  
729 (Marticorena and Bergametti, 1995; Marticorena et al., 1997).

730

731 Rates of dust deposition in sediment records are most often reported (e.g. Kohfeld and Harrison,  
732 2001) as aeolian mass accumulation rates (MARs), calculated as:

733

734 
$$\text{MAR (g m}^{-2} \text{ a}^{-1}) = \text{LSR} \times \text{DBD} \times f$$

735

736 where LSR = the linear sedimentation rate (calculated as the thickness of the sediment  
737 section (m)/ the time taken for the sediment to accumulate (yrs), DBD is the dry bulk  
738 density of the sediment ( $\text{g m}^{-3}$ ), and  $f$  is the fraction of the sediment that is of aeolian origin.  
739 Figures 9a and 10a show aeolian MARs for the present day and the LGM, respectively, as  
740 collated in our updated version ('DIRTMAP3') of the original 'Dust indicators and records of  
741 terrestrial and marine palaeo-environments, DIRTMAP' database (Kohfeld and Harrison, 2001).  
742 For comparison, figures 9b and 10b show modelled present day and LGM dust fluxes  
743 (Mahowald et al., 2006). Here, we discuss LGM deposition rates (as archived in DIRTMAP3);  
744 comparisons between these palaeo-data and the model simulations are discussed in section 6.2.

745

746

747 Figure 9.

748

749

750 Figure 10.

751

752

#### 753 4.2. *The continental loess record.*

754 For the terrestrial environment, sequences of loess sediments provide key aeolian archives on  
755 every continent. High-resolution loess/palaeosol sequences can provide new understanding of the  
756 relation between aerosols and climate (e.g. Maher, 2007): they provide a record of dust  
757 accumulation over large continental areas (Kukla, 1987; Thompson and Maher, 1995); they can  
758 be dated directly (Roberts, 2008; Roberts and Duller, 2003), to obtain dust flux changes through  
759 time; and they provide mineralogical, physical and chemical information on transported dust  
760 (e.g. Prins et al., 2008; Sun et al., 2008), which in turn may affect biogeochemical cycles in the  
761 ocean. Loess sequences can act both as a sink and a source for aeolian sediments. For example,  
762 the Chinese Loess Plateau, the largest Quaternary accumulation of windblown sediment in the  
763 world, has been identified by elemental and isotopic characterization as a major dust source for  
764 the sedimentary record of the northwest and equatorial Pacific (Rea et al., 1994; Shigemetsu et

765 al., 2007; Ziegler et al., 2008). In turn, the glacial stage units of the Chinese Loess Plateau  
766 comprise extremely well-mixed dust sourced most likely by westerly transport from the western  
767 Taklimakan Desert and the northern margins of Tibet (e.g. Maher et al., 2009). The controversy  
768 over the extent of Tibetan glaciations during past glacial stages, and hence the extent of  
769 glaciogenic supply of loess, remains unresolved at present (e.g. Lehmkuhl and Owen, 2005).  
770

771 Reflecting the proximal nature of many loess deposits to their sources, loess MARs at the LGM  
772 are the highest of any recorded aeolian fluxes (fig. 10a). For North America, for example, last  
773 glacial MARs in excess of  $6,700 \text{ g m}^{-2} \text{ a}^{-1}$  have been recorded from loess sections in central  
774 Nebraska (Roberts et al., 2003). Provenance information indicates that these extremely high  
775 deposition rates reflect increased sediment supply resulting from desiccation and expansion of  
776 areas of the Great Plains, rather than from glacial production of silt-sized particles (Aleinikoff et  
777 al., 2008). In light of MARS typically ranging from  $\sim 100$ s to  $1000$ s  $\text{g m}^{-2} \text{ a}^{-1}$ , and  
778 notwithstanding their proximal and thus coarser-grained nature, loess transport on such a scale  
779 inevitably incurred deflation and entrainment of large volumes of finer particles ( $< 10 \mu\text{m}$ ), of  
780 potentially significant radiative impact. For the central Nebraskan region, for example,  $\text{PM}_{10}$   
781 MARs range from  $\sim 500$  to  $> 900 \text{ g m}^{-2} \text{ a}^{-1}$ ; such high rates of fine particle flux suggested to have  
782 caused prolonged post-LGM cooling in the central N. American region (Roberts et al., 2003).  
783 Elsewhere in N. America, loess was supplied from glacial sediment sources, downwind from the  
784 Rocky Mountains, and close to rivers issuing from the southern margins of the Late Wisconsin  
785 ice sheet. Significant loess accumulations span areas of Iowa, Illinois and Kansas; further north  
786 and west, parts of Idaho, Wyoming, Alaska and the Yukon peninsula (fig. 10a).  
787

788 For the Eurasian region of the northern hemisphere, a swathe of loess girdles Eurasia from west  
789 to east (fig. 10a), reflecting dust supply both from expanded glaciogenic and arid/semi-arid  
790 sources. For the European zone, loess MARs typically range from  $\sim 200$  to  $600 \text{ g m}^{-2} \text{ a}^{-1}$  for the  
791 last glacial stage, with some 'hotspots' of deposition reported, e.g. for the Nussloch site,  
792 Germany ( $> 2000 \text{ g m}^{-2} \text{ a}^{-1}$ , Rousseau et al., 2002) and Paks in Hungary ( $> 1000 \text{ g m}^{-2} \text{ a}^{-1}$ ,  
793 Frechen et al., 1997). Extending eastwards over the Russian steppe, loess MARs have a similar  
794 range ( $\sim 200 - 700 \text{ g m}^{-2} \text{ a}^{-1}$ ), rates then increase towards Tajikistan ( $\sim 1700 \text{ g m}^{-2} \text{ a}^{-1}$ , Frechen  
795 and Dodonov, 1998) and into the loess sequences of China. Across the Chinese Loess Plateau,

796 the increase in MARs during the last glacial stage, as estimated from median MARs across the  
797 region, is of the order of up to  $\times 5$ , i.e.  $\sim 300 \text{ g m}^{-2} \text{ a}^{-1}$  compared to an estimated  $65 \text{ g m}^{-2} \text{ a}^{-1}$  for  
798 the last interglacial stage ( $\sim 125 \text{ ka}$ , MIS5e) (Kohfeld and Harrison, 2003). However, the range  
799 of LGM MARs estimated for the Loess Plateau spans  $< 100$  to  $> 1000 \text{ g m}^{-2} \text{ a}^{-1}$ . In common with  
800 most of the world's loess sequences, major spatial and temporal (sub-millennial) variability of  
801 these glacial-stage increases in dust flux have been reported (e.g. Zhang et al., 2002). In terms of  
802 mineralogy, quartz ( $\sim 35 \%$ ), feldspars ( $\sim 20 \%$ ), clay minerals (25 %, illite and illite-smectite)  
803 and calcite (12 %) make up the very well-mixed loess (Jeong et al., 2008). Magnetic analyses  
804 additionally quantify the presence of minor amounts of haematite and goethite ( $\sim 1 \%$ ); the grain  
805 size of the haematite in the interbedded palaeosols is distinctively sub-micrometre, and confers  
806 the reddish hues to these soils (Maher, 2007).

807

808 Figure 11

809

810 A characteristic feature of many loesses is their trimodal particle size distribution, with  
811 overlapping coarse and fine size modes (fig. 11, and e.g. Wang et al., 2006; Machalett et al.,  
812 2008; Lim and Matsumoto, 2008). For the Chinese Loess Plateau, particle size end members  
813 have been identified (Prins et al., 2007) comprising a 'sandy loess' component (modal size  $63 \mu\text{m}$ )  
814 a 'silty loess' (modal size  $37 \mu\text{m}$ ), and a 'clayey loess' (modal size  $22 \mu\text{m}$ ). The two  
815 coarser end members are thought to reflect saltation and suspension of 'local' coarse dust during  
816 dust storms in spring; the finer end member inferred to represent higher level transport by  
817 background and dust storm supply. For these sequences, increases in the proportion of coarse  
818 particles has been interpreted as an increase in wind speed and thus of more intense winter  
819 monsoon circulation (e.g. Xiao et al., 1995; Maher and Hu, 2006). The increased glacial-stage  
820 MARs are also associated with increased particle size. At Luochuan, central Loess Plateau, for  
821 example, the median diameter of aeolian quartz particles is  $\sim 15 \mu\text{m}$  for the last glacial stage  
822 sediments compared with  $\sim 6 - 9 \mu\text{m}$  for the Holocene (fig. 12; Xiao et al., 1992). These paired  
823 increases in fluxes and particle size may indicate both expanded source areas and increased wind  
824 speeds, possibly coupled with a shortened transport path. Asian dust is and has been a major  
825 component of the sedimentary record of the Pacific Ocean (*see section 4.3*). Downwind of the  
826 Loess Plateau, present day estimates of dust flux range from  $\sim 6$  to  $12 \times 10^6$  tons of Asian dust

827 transported annually to the central North Pacific. Larger quantities are probably deposited over  
828 the western North Pacific, closer to the Asian sources (Uematsu et al., 1983).

829

830 Figure 12

831

832 For the southern hemisphere, significant loess accumulations exist in South America and New  
833 Zealand (fig. 10a). For the South American continent, the loess sequences of the Argentinian  
834 pampas (the central Argentinian plains) cover  $\sim 1.0 \times 10^6$  km<sup>2</sup>, with an average thickness of  $\sim 30$ -  
835 40 m, thus comprising the largest loess accumulation in the southern hemisphere. Many of these  
836 sequences have yet to be analysed in detail; age control is limited for the Tezanos Pinto  
837 Formation, traditionally considered to represent the last glacial stage but possibly extending to  $\sim$   
838 145 ka (Kemp et al., 2004). Estimated/adjusted MARs for the Pampas for the LGM are thus  
839 scarce but range from 45-145 g m<sup>-2</sup> a<sup>-1</sup> (Krohling and Iriondo, 1999, cited in Mahowald et al.,  
840 2006). Primary sources of these aeolian sediments are linked directly or indirectly to the Andean  
841 region and the *arid diagonal* located to the west. In general, the composition of the Argentine  
842 loess mirrors the mineralogical and chemical composition of Andean rocks (Gaiero, 2007).  
843 Reflecting a dominant Patagonian dust source, the southern loess deposits ( $\sim 36$ -38 °S) have  
844 similar chemical and isotopic composition and a dominance of smectite, with illite present in  
845 smaller amounts (Gaiero et al., 2004). This contrasts with the clay mineralogy of northern loess  
846 deposits ( $\sim 26$ -28 °S), dominated (Schellenberger and Veit, 2006) by illite (70–85 %), followed  
847 by smectite (10–25 %). Between the northern and southern loesses, illite slightly dominates the  
848 mineralogical composition of loess, while the coastal plains (Buenos Aires province) exhibit an  
849 approximately equal mixture of smectite, illite and kaolinite (Zarate, 2003). The loess deposits  
850 are coarse-textured. In the southern Pampas, the average sand–silt–clay content is 44 %, 26 %  
851 and 24 %, respectively, the maximum sand content (mostly very fine sand) being  $\sim 80$  % (Bidart,  
852 1996). Magnetic properties of loess from this area indicate the presence of detrital magnetite and  
853 titanomagnetite as the main ferromagnesian minerals (Orgeira et al., 2009). Northern loess  
854 deposits show fairly uniform particle size (silt loam) averaging 14 % sand, 74 % silt, and 12  
855 % clay. 80 % of the iron in the sediment is in paramagnetic minerals;  $\sim 20$  % of the iron is in  
856 hematite ( $\sim 1.6$  % of the sediment's mass), and  $\ll 1$  % of the total iron is present as  
857 magnetite/maghemite ( $\sim 0.1$  % of the sediment's total mass) (Carter-Stiglitz et al., 2006).

858 Elsewhere in the southern hemisphere, loess occurs most extensively in New Zealand, notably in  
859 the lowlands surrounding formerly glaciated areas of the South Island (Eden and Hammond,  
860 2003). Glacial and floodplain sources have dominated dust production in New Zealand; its loess  
861 deposits are thicker (or possibly better documented) than in the S. American Pampean region, for  
862 example. Locally, Holocene loess has accumulated up to several metres thickness and very thick  
863 (> 10s m) LGM accumulations have been documented (Berger et al. 2002). MARs for the LGM  
864 range between 70 to 150 g m<sup>-2</sup> a<sup>-1</sup> (Eden and Hammond, 2003). Most of the New Zealand loess is  
865 quartz- and feldspar-rich. Glacially-derived loess deposits are strongly weathered (Marx et al.,  
866 2005), being clay-poor but mica-rich and relatively coarse-grained compared to dust derived  
867 from arid lands (McGowan et al., 2005).

868 Loessic sediments occur to more limited extents in other southern hemisphere countries. The  
869 'parna' of southeastern Australia, for example, is characterised by low supply and deposition  
870 rates; it dominantly comprises illite and kaolinite, carbonate and silt-sized quartz particles (e.g.  
871 Summerell et al., 2000). Accumulation rates measured at a 3m loess section in the highlands  
872 west of Sydney were relatively unchanged from 50-10 ka, at between 28 - 45 g m<sup>-2</sup> a<sup>-1</sup> (Hesse et  
873 al., 2003).

874

#### 875 *4.3. The palaeo-dust record of HNLC regions in the Pacific.*

876 As discussed above, the dust-generating regions of E. Asia appear to have expanded and  
877 contracted across a range of glacial/periglacial and arid zone environments during the LGM. In  
878 turn, these source changes have produced major changes in dust flux and dust characteristics to  
879 the downwind continental sequences of the Chinese loess and palaeosols, and thence onwards to  
880 the northwestern and equatorial Pacific.

881

882 For the northwest and the equatorial Pacific regions, there are globally important questions  
883 regarding the possible causal role of changes in ocean biogeochemistry and export production  
884 (carbon sequestration, i.e. biomass permanently removed from the carbon cycle) during the  
885 glacial and interglacial stages of the Quaternary. The importance of understanding the influence  
886 of aeolian-supplied Fe can be illustrated by recent fertilization experiments in the northwest  
887 region of the Pacific, where iron-enrichment experiments led to the greatest expansion yet  
888 observed of phytoplankton standing stocks, together with shifts to dominance of large centric

889 diatoms (Tsuda et al., 2003; de Baar et al., 2005). These two features suggest that the NW Pacific  
890 Ocean may be particularly sensitive to iron fertilization. With regard to export production (i.e.  
891 subsurface storage, rather than re-release from the ocean mixed layer), the drawdown of  
892 atmospheric carbon resulting from this experiment is estimated at ~ 150-1500 C atoms per iron  
893 atom added (*see section 3.1*; de Baar et al., 2008). The controversial question of whether changes  
894 in palaeo-dust fluxes have caused past changes in ocean productivity in the NW Pacific is also  
895 outlined here. Unequivocal identification of palaeo-productivity changes in the Pacific remains  
896 challenging in light of: differing approaches to determining sediment fluxes; varying rates of  
897 formation and dissolution of individual biogenic components (e.g. carbonate and opal; Honjo et  
898 al., 1995); and varying interpretations of radionuclide (e.g.  $^{230}\text{Th}$ ) and organic degradation  
899 products (e.g. biogenic barite). Combined, these issues have resulted in difficulties in quantifying  
900 and integrating different biological products and hence total marine export production. As  
901 discussed below, different proxies and different approaches to age models have resulted in  
902 differing interpretations of glacial-interglacial changes in palaeo-productivity.

903

#### 904 *4.3.1. Northwest Pacific dust fluxes*

905 Reported as one of the best records of aeolian flux from the NW Pacific, core V21-146 from the  
906 Shatsky Rise (37.7°N 163°E) was retrieved from an elevated core site which lies beneath the  
907 loess-transporting westerlies and provides a robust  $\delta^{18}\text{O}$  record for age control (Hovan et al.,  
908 1989, 1991). These Shatsky Rise sediments display increased dust fluxes (~ x 3.5, i.e. from ~ 2  
909 to 7  $\text{g m}^{-2} \text{a}^{-1}$ ) during glacial stages, and indicate a direct link between intervals of accelerated  
910 continental deposition of loess and enhanced dust flux to the open ocean. Variations in aeolian  
911 grain size in the Shatsky Rise (fig. 13) show forcing at Milankovitch orbital frequencies, but  
912 coarser grains (median diameter ~ 8  $\mu\text{m}$ ) show some association both with interglacial stages (at  
913 the 100 ka frequency) and with low tilt and maximal precession, i.e. associated with glaciations  
914 (Rea, 1994). Gravity core H3571 was recovered from the Hess Rise (34.9 °N 179.7 °E) some  
915 1500 km to the east of the Shatsky Rise site and ~ 3000 km from the Chinese mainland. This  
916 record also displays higher dust fluxes during glacial stages, of similar magnitude to the V21-146  
917 record. However, these two records show significant differences in the timing and magnitude of  
918 their dust flux maxima; V21-146 has maximal fluxes at the LGM, for example, while H3571  
919 displays only moderate flux values (~ 5.5  $\text{g m}^{-2} \text{a}^{-1}$ ) at this time.

920 Tracing Asian dust around its entire northern hemisphere trajectory, the Chinese loess, Pacific  
921 deep-sea record and Greenland ice cores all display large and rapid changes both in dust flux and  
922 particle size from interglacial into glacial climate stages (fig. 12). It is notable too that large dust  
923 flux changes are also evident *within* glacial stages, on sub-millennial timescales. The relative  
924 significance of changes in source (e.g. stronger and/or more frequent dust emission events)  
925 compared with changes in transport efficiency (e.g. shorter transit times and/or reduced rain-out  
926 en route) has been the subject of debate (e.g. Nilson and Lehmkuhl, 2001; Ruth, 2005; Lambert  
927 et al., 2008; Fischer et al., 2007). For the North-Greenland Ice Core Project (NGRIP) core, at the  
928 distal end of this transport path, the dust source is thought from isotopic evidence to have  
929 remained 'constant' through glacial and interglacial stages, i.e. the Taklimakan and Tengger/Mu  
930 Us deserts of W. China and Inner Mongolia, respectively (Svensson et al., 2000; Bory et al.,  
931 2003), though this is poorly expressed by models (e.g. Andersen et al., 1998). Given stationary  
932 sources, the evidence for increased dust flux all along the transport pathway, i.e. from the  
933 proximal Chinese loess sequences (e.g. Xiao et al., 1999; Ding et al., 2002), and across the mid-  
934 latitude Pacific (e.g. Hovan et al., 1991; Kawahata et al., 2000) does indicate stronger and/or  
935 more frequent winds at source (Ruth, 2005). This conclusion requires careful examination,  
936 however. Care is needed in linking the more proximal loess sequences with the distal Pacific  
937 record. First, from air mass back-trajectory analysis, present day (coarse, ~ 2- 20  $\mu\text{m}$ ) dust  
938 supply to the Loess Plateau appears dominantly controlled by ambient (non-dust storm)  
939 northwesterly surface winter winds (Zhang et al., 1999). This is in contrast to the (fine-grained)  
940 LRT dust, which is uplifted during springtime dust storms from sources such as the western  
941 Taklimakan Desert (Bory et al., 2003) and carried by the westerly jet at upper levels (~ 5- 8 km)  
942 before descending towards the northeasterly trade winds, to the south and east of the subtropical  
943 high (Zhang et al., 1999). In contrast, during glacial stages, the dust source for both the LRT dust  
944 and the Loess Plateau appears, on multi-proxy elemental and magnetic grounds, to have been the  
945 same, dominantly the western desert and Tibetan Plateau regions (Maher et al., 2009; Torii et al.,  
946 2001). Second, in terms of particle size, glacial-interglacial variations of particle size along the  
947 proposed dust trajectory - from the East Asian deserts to the Loess Plateau to the NW Pacific to  
948 Greenland - present some complications. For the Loess Plateau and for NGRIP, glacial stages are  
949 associated with increased particle size (fig. 12). Glacial stages in the NGRIP dust record are  
950 characterized by particles > 1.6  $\mu\text{m}$  in diameter; interglacials by particle sizes ~ 1  $\mu\text{m}$ .



951 For the NW Pacific (Shatsky Rise), in contrast, finer sizes (median diameter  $\sim 3 \mu\text{m}$ ) have been  
952 reported for glacial stages and coarser sizes ( $5 \mu\text{m}$ ) during interglacials (e.g. Hovan et al., 1991),  
953 whilst data collated by Zhang et al. (2007) show no clear glacial-interglacial trend (fig. 13).

954

955 Figure 13

956

#### 957 *4.3.2. Equatorial Pacific dust fluxes*

958 Until recently, consistency of glacial-interglacial records of dust flux has been difficult to obtain  
959 for the equatorial Pacific, reflecting the confounding effects of sediment redistribution by bottom  
960 water currents (e.g. Mitchell and Lyle, 2005) and inputs of volcanically-derived aerosols  
961 (Olivarez and Owen, 1991). Critically, glacial to interglacial aeolian fluxes estimated from  
962 oxygen isotope-derived age models have varied both in sign and magnitude (e.g. Rea, 1994;  
963 Murray et al., 1995). Development of the  $^{230}\text{Th}$  normalization method (Bacon, 1984; Francois et  
964 al., 1991; Francois et al., 2004) may provide a means (Lyle et al., 2007; Broecker, 2008) to  
965 calculate truly aeolian MARS. This method is based on the assumption that the rain-out rate of  
966 particulate  $^{230}\text{Th}$  sinking to the sea bed is equivalent to the known rate of  $^{230}\text{Th}$  production by  
967  $^{234}\text{U}$  decay in the overlying water column. MARS are calculated by dividing the production rate  
968 of  $^{230}\text{Th}$  in the water column by the concentrations of scavenged  $^{230}\text{Th}$  in the sediment. In  
969 tandem, terrigenous dust fluxes have been estimated for these sediments by using the ratio of  
970  $^{232}\text{Th}$  and  $^{230}\text{Th}$  as a proxy for dust flux.  $^{232}\text{Th}$  is dominantly sourced from continental dust  
971 sources (its concentration in volcanogenic materials, for example, is an order of magnitude  
972 lower), it is almost insoluble in seawater, and is not subjected to any post-depositional redox-  
973 related change. Differences in dust provenance have so far been reported to exert relatively little  
974 effect on measured  $^{232}\text{Th}$  concentrations, falling within 1 ppm of the upper continental crustal  
975 average of 10.7 ppm (Anderson et al., 2006). For an increasing number of sediment cores  
976 spanning the equatorial Pacific, recent studies making use of these techniques consistently  
977 demonstrate increased dust fluxes during glacial stages, with dust fluxes up to 2.5 x higher  
978 during the last glacial stage (Anderson et al., 2006; McGee et al., 2007; Winckler et al., 2008).  
979 Dust fluxes decrease from west to east, and towards the south, where northern and equatorial  
980 South American dust sources may contribute (Anderson et al., 2006), compared to the  
981 dominantly Asian-sourced aerosol towards the north (Ziegler et al., 2009). These new data

982 indicate a coherent response by dust generation and transport processes, from the equator to the  
983 South Pole, to glacial/interglacial climate changes through the late Pleistocene (fig. 14). The  
984 gradient in dust flux values from north to south across the equator may change through time in  
985 response to changes in the latitudinal position and/or the intensity of the intertropical  
986 convergence zone (Rea, 1994; Rea and Hovan, 1995; McGee et al., 2007; Ziegler et al., 2008).  
987 As noted by Anderson et al. (2006) and Winckler et al. (2008), the observed equatorial Pacific  
988 glacial flux increase (up to x 2.5 higher) is very different from modelled dust fluxes for this  
989 region. For example, while earlier models (Mahowald et al., 1999, Reader et al. 1999) produced  
990 unrealistically large LGM flux increases (LGM/modern ratios ranging between 12 and 105),  
991 more recent simulations result in disproportionately low LGM fluxes and also a west to east  
992 *increase* in flux (Mahowald et al., 2006).

993

994 Figure 14

995

996

997 *4.4. The palaeo-dust record of the HNLC ocean regions: Antarctic ice cores and Southern*  
998 *Ocean.*

999 *4.4.1. Antarctic ice cores*

1000 Some of the longest, most detailed palaeo-dust records have been retrieved from the Antarctic  
1001 polar region. Our interpretation of this paleo-record would be greatly facilitated by  
1002 measurements of present-day dust transport. Unfortunately, there are remarkably few dust data in  
1003 the Southern Hemisphere and, consequently, a major problem in quantifying LGM/modern dust  
1004 flux ratios for the wider Southern Ocean region is the paucity of data with which to assess  
1005 present day dust fluxes from any of the S. hemisphere land masses (*section 2.2*). Dust fluxes  
1006 recorded in Antarctic ice cores show that across the southern polar region dust flux was ~ 30 x  
1007 higher during the LGM (fig. 15). The reasons for this very large increase in dust at the LGM can  
1008 be multiple. Source regions over Patagonia were certainly extended due to drier conditions and  
1009 the decrease in sea-level that left exposed wide continental regions. In addition, the transport  
1010 paths could have been changed and the wet deposition is likely to have been considerably  
1011 reduced. These factors taken in combination or seldom can explain the differences seen in the  
1012 deposition fluxes in the ice cores (Yung et al., 1996).

1013

1014 Figure 15

1015

1016 The growing number of ice cores retrieved from the Antarctic plateau provide long, undisturbed  
1017 dust and climate records (extending back over eight glacial cycles) representing not only the  
1018 Southern Ocean region but also sectoral differences (fig. 16) within that region (Fischer et al.,  
1019 2007). High resolution dust flux records can be compared between several locations, including  
1020 Vostok (78° 28'S 106° 48'E) (Petit et al., 1999) and the two EPICA sites (EPICA community  
1021 members, 2006), Dome C (EDC, 75°06'S, 123°21'E, in the Indian Ocean sector) and at Kohnen  
1022 station in Dronning Maud Land (EDML: 75°00'S, 00°04'E), and Talos Dome (72° 48'S 159°  
1023 06'E; Delmonte et al., 2009). For all these cores, the dust profiles are extremely similar both in  
1024 terms of total dust flux and the magnitude of the glacial/interglacial changes (Fischer et al., 2007;  
1025 Lambert et al., 2008) with high dust fluxes during glacial stages; up to 30 x higher compared  
1026 with the Holocene and earlier interglaciations. At EDC, total dust fluxes range from ~ 0.4 mg m<sup>-2</sup>  
1027 a<sup>-1</sup> at the Holocene to 12 mg m<sup>-2</sup> a<sup>-1</sup> during glacial maxima. Tight coupling of the dust and  
1028 climate records is especially evident during cold intervals of glacial stages (i.e. when  $\delta D > \sim 425$   
1029 ‰, Lambert et al., 2008). Episodes of warming are linked with reduced dust fluxes; for instance,  
1030 each of the Antarctic isotopic maxima (EPICA community members, 2006) is associated with a  
1031 decline in dust fluxes at both the EDC and EDML sites (Fischer et al., 2007), indicating  
1032 synchronous changes in the dust source regions (e.g. reduced dust supply and/or reduced wind  
1033 speeds).

1034

1035 Figure 16.

1036

1037 Comparison of the EDML and EDC records identifies that the flux of non-sea salt calcium (i.e.  
1038 as a mineral dust proxy) was 3 x higher at the EDML, Atlantic sector site; this higher flux is  
1039 attributed to its greater proximity to S. American dust sources (Fischer et al., 2007). Debate is  
1040 ongoing with regard to the relative influences of glacial-interglacial changes in source strength  
1041 and changes in dust transport processes (especially atmospheric residence time and transport  
1042 rate) on rates of dust deposition at Antarctica (e.g. see Fischer et al., 2007 compared with  
1043 Lambert et al., 2008). If source strength has been the major influence, then glacial dust fluxes to

1044 the Atlantic sector of the Southern Ocean, lying between the S. American sources and the  
1045 Antarctic region, should approach those observed in Antarctica (*see section 4.4.2*).

1046

1047 An additional if subtle complication arises with reported regional differences dust particle size  
1048 around the different Antarctic sectors (Delmonte et al., 2004). In terms of dust particle size, the  
1049 mass-size distribution of aeolian particles in east Antarctica is well sorted around a mean mass  
1050 diameter of 2  $\mu\text{m}$ , the largest particle diameter is  $\sim 5 \mu\text{m}$ , and the contribution of particles  $< 0.7$   
1051  $\mu\text{m}$  contributes  $< \sim 10\%$  of the total mass (Delmonte et al., 2004b). However, during glacial  
1052 stages, smaller particles were deposited at the EDC site, in contrast to Vostok and Dome B,  
1053 where increased particle size has been reported during glacial stages (Petit et al., 1982; Delmonte  
1054 et al. 2004).

1055

1056 Provenancing of the Antarctic dust by magnetic (Maher, 2009), elemental (Marino et al., 2008)  
1057 and isotopic, Sr and Nd, techniques (e.g. Grousset et al., 1992; Basile et al., 1997) is still being  
1058 refined, in order to extend the range of well-characterised potential source areas. As noted by  
1059 Mahowald et al. (1999), no potential source area can be discounted or verified if it has not been  
1060 sampled and characterised. It is also essential to take into account particle size effects on  
1061 mineral, elemental and isotopic signatures (figure 17a and b and e.g. Delmonte et al., 2004;  
1062 2009). For dust both at Vostok and EDC, the isotopic fields are almost identical (within the  
1063 ranges  $0.708 < {}^{87}\text{Sr}/{}^{86}\text{Sr} < 0.711$  and  $-5 < \epsilon \text{Nd} (0) < +5$ ), indicating a common dust source across  
1064 the East Antarctic region. So far, around 60 potential source areas samples from S. America (33  
1065 samples), New Zealand (15), Australia (22), southern Africa (6) and the Antarctic Dry Valleys  
1066 (6) have been subjected to isotopic analysis of their fine size fractions ( $< 5 \mu\text{m}$ ), encompassing  
1067 loess and loess-like deposits, silts and sands, fluvio-glacial sediments, soils, moraines (Delmonte  
1068 et al., 2004, 2007, 2009) and glacial lacustrine deposits (Sugden et al., 2009). At present, those  
1069 potential source areas showing partial overlap with the glacial stage ice core dusts are S.  
1070 America, New Zealand, the Antarctic Dry Valleys (fig. 17), and some Australian areas (Lake  
1071 Eyre). The overlap with Australian-sourced samples increases for interglacial stage samples  
1072 (Revel-Rolland et al., 2006; Delmonte et al., 2007; 2009; Marino et al., 2008).

1073

1074 Figure 17.

1075

1076 So far, the consensus has been that Patagonia, and possibly the Puna Altiplano area, have been  
1077 the major glacial dust sources for the Antarctic ice records (Gaiero, 2007; Delmonte et al., 2007).  
1078 Satellite observations identify these hypothesized source regions as prominent present-day dust  
1079 sources (*see fig. 3b, section 2.2*). Lowered glacial-stage sea level rendered Patagonia twice as  
1080 large as today and glacial activity was extensive; this region was most likely an intense dust  
1081 source during the LGM and previous glacial stages (e.g. Iriondo, 2000; Sugden et al., 2009).  
1082 There is no published continental record of Quaternary dust accumulation from Patagonia,  
1083 reflecting the high potential for wind erosion and poor geomorphic conditions for sediment  
1084 accumulation. Rather, loess accumulation was concentrated further to the north and east in the  
1085 pampas of Argentina (e.g. around  $\sim 30^\circ\text{S}$ ,  $\sim 60^\circ\text{W}$ ). Age control for these loess sequences has  
1086 not been well established yet, but MARs at the LGM are estimated at between  $\sim 45$  and  $145 \text{ g}$   
1087  $\text{m}^{-2} \text{a}^{-1}$  (Krohling and Iriondo, 1999). Evidence for direct links between the Antarctic records of  
1088 dust deposition and activating sources within the Patagonian region comes from chronological  
1089 and isotopic data on glacial outwash sediments fringing the Straits of Magellan (Sugden et al.,  
1090 2009). During the last glaciation, an Andean ice sheet developed between latitudes  $44^\circ\text{S}$  and  $55^\circ\text{S}$ .  
1091 Outlet glaciers flowing eastwards from this ice sheet debouched into large outwash plains,  
1092 with meltwater eventually draining into the S. Atlantic Ocean. When the outlet glaciers advanced  
1093 to terminate directly at the outwash plains around the (then emergent) Magellan Straits, large  
1094 amounts of glacial dust were supplied and available for uplift and transport by intensified  
1095 westerly winds. Conversely, when the glaciers terminated in proglacial lakes, the dust supply  
1096 was effectively ‘switched off’. Isotopic analysis (Sr, Nd) of the finest fraction ( $<5 \mu\text{m}$ ) of  
1097 Magellan glacial lacustrine sediments indicates that they may have been one of the sources of the  
1098 Antarctic dust during glacial stages (Sugden et al., 2009). Correlation between the dated  
1099 sequence of Patagonian glacial activity and the Antarctic (Dome C) dust peaks is good, although  
1100 anomalies do occur (e.g. the ice core dust peak at 30 ka has no glacial correlative in Patagonia;  
1101 and the dated sediments are lacustrine, possibly indicative of periods of glacial retreat rather than  
1102 advance). Rapid retreat of the glaciers can account for the observed rapid decreases in dust flux  
1103 during deglaciation. Notably, the deglacial decline in Antarctic dust flux precedes the reduction  
1104 in sea ice in the Southern Ocean, the warming of the Southern Ocean and the deglacial rise in sea  
1105 level (Sugden et al., 2009; Fischer et al., 2007; Wolff et al., 2006). One problem with this

1106 proposed glacial outwash dust source is the key and characteristic presence of leaf wax-derived  
1107 n-alkanes in the intervening sedimentary record of the Southern Ocean (*see section 4.4.2*); the  
1108 presence of vegetation might suggest a more low-lying source area (e.g. the emergent coastal  
1109 shelf) rather than a glacial/periglacial source.

1110

#### 1111 *4.4.2. The Southern Ocean dust record*

1112 It has so far proved difficult to quantify the flux of dust across the Southern Ocean, and thence to  
1113 Antarctica, during past glacial stages. Glacial stage sediments are characterised by increased  
1114 lithogenic MARs in all three sectors of the Southern Ocean but these sedimentary records have  
1115 diverse terrigenous contributions, including iceberg-rafted debris, current-transported sediment  
1116 and aeolian dust (Diekmann, 2007). As with the Pacific sedimentary record, significant lateral  
1117 transport of sediment by strong bottom water currents has affected the Southern Ocean records.  
1118 There are presently very few integrative studies which have used a range of independent proxies  
1119 to establish robust lithogenic and aeolian MARs, and identify glacial/interglacial changes in  
1120 export production. Martínez-García et al. (2009) present multiproxy-derived estimates of dust  
1121 and iron supply, and marine productivity, over the last 1.1My from a deep-sea sediment record  
1122 (PS2489-2, ODP Site 1090) located in the subantarctic South Atlantic (fig. 18). The initial age  
1123 model for this composite sedimentary record was obtained by oxygen isotope analysis of benthic  
1124 foraminifera (prior to tuning by correlation between the sediment sea surface temperature record  
1125 and the EDC temperature record). Using  $^{232}\text{Th}:$  $^{230}\text{Th}$  as a proxy for dust flux (in combination  
1126 with other terrigenous tracers, including long-chain n-alkanes derived from leaf waxes), these  
1127 authors report dust fluxes varying from  $\sim 1 \text{ g m}^{-2} \text{ a}^{-1}$  ( $\sim 50 \text{ mg m}^{-2} \text{ a}^{-1}$  of iron) for interglacial  
1128 periods, to  $\sim 5\text{-}7 \text{ g m}^{-2} \text{ a}^{-1}$  ( $\sim 250 \text{ mg m}^{-2} \text{ a}^{-1}$  of iron) for glacial stages (fig. 18). Assuming the dust  
1129 source to be S. America, these estimates thus indicate up to 7 x increase in the source strength for  
1130 this region. This magnitude of change contrasts with the x 20-25 increase in dust flux recorded  
1131 by the Antarctic ice cores (Lambert et al., 2008). These data suggest that increased dust source  
1132 strength, due to increased (possibly glacially-driven) dust supply and/or higher wind speeds,  
1133 might be less significant for the recorded Antarctic dust fluxes than increased efficiency of long  
1134 range transport, due to shorter transit times and/or less rain-out en route.

1135

1136 Figure 18

1137

1138 In terms of dust source, whilst very few dust flux data exist for either the present day or the LGM  
1139 for this region, present day dust fluxes seem very low (*see section 2.2*). Hence, it is probable that  
1140 sources additional to present-day southern S. American source areas were active during the last  
1141 glacial period. Recent research does indicate that southern S. America was the most important  
1142 LGM dust source for this region, as evidenced by the dominant overlap of isotopic compositions  
1143 of sediments from the Pampean loess (Gaiero, 2007), the Southern Ocean (Walter et al., 2000)  
1144 and the Antarctic dust (Delmonte et al., 2008). Modelling (Krinner and Genthon, 2003; Lunt &  
1145 Valdes, 2002) and mineralogical studies (Gaiero, 2007; Delmonte et al., 2008) indicate that ~ 80  
1146 % of dust reaching the Antarctic Plateau during cold periods was derived from Patagonia. Less  
1147 clear is the contribution of southern S. American materials to the South Atlantic Ocean. On  
1148 isotopic grounds, glacial sediments from the north Scotia Sea seem dominated by a southernmost  
1149 Patagonian source (Walter et al., 2000). However, compared with the potential southern S.  
1150 American sources, sediments from the Atlantic sector of the Southern Ocean (~41 °S, 14 °W)  
1151 display notably higher chlorite content (Diekmann, 2007). Total detrital MARs for this region  
1152 are ~15 g m<sup>-2</sup> a<sup>-1</sup> for the last interglacial stage (MIS 5, ~ 125 ka) and ~120 g m<sup>-2</sup> a<sup>-1</sup> for the LGM  
1153 (Walter et al., 2000), but the aeolian contribution to these figures is not known.

1154

1155 For the other S. hemisphere land masses, the Murray-Darling basin was a major Australian dust  
1156 source in the LGM, exporting dust towards the south and southeast (Hesse, 1994; Hesse and  
1157 McTainsh, 2003). With dust storm frequency ~ 50 % higher (McTainsh and Lynch, 1996), the  
1158 LGM flux of Australian dust to the mid-Tasman Sea (160 °E, 40 °S) is estimated at ~ 3 x  
1159 Holocene levels (Hesse, 1994). An additional LGM sediment source was the ancestral Lake Eyre  
1160 basin (Dulhunty, 1982). Australian dust may have been a long-term contributor to the Southern  
1161 Ocean and the southwest Pacific (Stancin et al., 2006, 2008) but the record is obscured by poor  
1162 carbonate preservation in the deep Southern Ocean basins and reworking of sediments by strong  
1163 Antarctic circum-polar circulation. In terms of particle size, LGM dust transported to the Tasman  
1164 Sea from Australia (Hesse and McTainsh, 1999) comprises a minor coarse silt component  
1165 (modal diameter 20-25 µm) and dominant fine silt and clay (< 4 µm). Compared to the present  
1166 day, there is little change either in the mode or the size of the coarser particles, indicating little  
1167 change in the strength of the zonal westerlies over Australia at the LGM. The palaeo-dust

1168 contained within Tasman Sea sediments contains fine-grained haematite/goethite, the  
1169 concentration of which varies directly with the dust concentration (Hesse, 1997). Thus, the flux  
1170 of fine-grained haematite/goethite to the Southern Ocean and southwest Pacific and from  
1171 Australia was greater in the LGM and previous glacials than in the Holocene and previous  
1172 interglacials. (The volumetric flux of these minerals, however, has yet to be determined).  
1173 Although New Zealand appears to be an insignificant source of dust to the Pacific sector of the  
1174 Southern Ocean today (Prospero et al., 2002; fig. 3), it represents a (poorly documented) potential  
1175 contributor of dust to the palaeo-Southern Ocean. Given its significant terrestrial accumulations  
1176 of loess, New Zealand may have been a significant source of dust to the southwest Pacific and  
1177 Southern Ocean at least during glacial intervals of the Late Quaternary. While the geochemical  
1178 fingerprint of New Zealand dust overlaps glacial and interglacial Vostok dust signatures (figure  
1179 16; Delmonte et al., 2004, 2009), it is likely that New Zealand, like Australia, contributes more  
1180 to the Pacific than other sectors of the Southern Ocean. It is notable that New Zealand's Taupo  
1181 volcano has been excluded as a potential aerosol source for the Antarctic ice cores from Dome C  
1182 ( $75^{\circ}06'S$   $123^{\circ}21'E$ ) and Vostok ( $78^{\circ}28'S$   $106^{\circ}46'E$ ), on the basis of its relatively high  $SiO_2$   
1183 content (Narcisi et al., 2005). The absence of Taupo-sourced tephra from the last 200 ka of the  
1184 Antarctic ice cores suggests that dust transport from the New Zealand/Australian region to  
1185 Antarctica is inefficient (Delmonte et al., 2004; Narcisi et al., 2005). Yet Taupo volcanic dust has  
1186 been found in Greenland ice cores (Zielinski et al., 1996); it seems unusual that there could have  
1187 been no transmission of dust, at least to subantarctic Southern Ocean waters, given the extreme  
1188 size of some eruptions. For example, in an eruption dated at  $\sim 26.5$  ka, Taupo emitted  $\sim 1.4 \times 10^{18}$   
1189 g of tephra (Mason et al., 2004), i.e. 1000 x the current global dust flux.

1190

1191 Given the current scarcity of present day or palaeo-dust flux and provenance data for the  
1192 southern S. hemisphere region, model simulations remain to be validated. For the LGM,  
1193 modelling of southern hemisphere dust deposition (e.g. Lunt and Valdes, 2002) suggests up to 15  
1194 x greater flux of South American dust in the Atlantic sector of the Southern Ocean (compared  
1195 with up to 7 x flux as reported by Martinez-Garcia et al., 2009), and > 90 % contribution of S.  
1196 American dust to deposition in the Atlantic and Indian Ocean sectors. In the Lunt & Valdes  
1197 (2002) study, the modelled LGM flux of Australian dust exceeds the S. American value  
1198 (presumably reflecting the greater area of the Australian continent) but the Australian dust is



1199 dispersed over the entire Pacific sector of the Southern Ocean (Lunt and Valdes, 2002). Other  
1200 modelled fluxes (Mahowald et al., 1999, 2006) indicate LGM/modern ratios in the Southern  
1201 Ocean ranging from 0.5 to 2 (with dust sources as modelled for the present day) and 20 to 50  
1202 (with modelled expansion in dust sources at the LGM).

1203

#### 1204 *5. Palaeo-productivity changes during the LGM in the HNLC regions*

##### 1205 *5.1. Northwestern and equatorial Pacific.*

1206 In the context of iron fertilization, did the measured LGM enhancement of flux of Asian dust  
1207 overcome iron limitation in the HNLC regions of the northwestern and equatorial Pacific,  
1208 resulting in significant drawdown of atmospheric CO<sub>2</sub>? Export production has previously been  
1209 reported to have been globally higher during the LGM, with just one exception, the Southern  
1210 Ocean (Kohfeld et al., 2005). Some studies have identified an association between glacial  
1211 climate stages and increased productivity in the Pacific Ocean. For example, for core H3571  
1212 from the NW Pacific, biogenic opal and organic carbon were both reported to have increased  
1213 during glacial stages (Kawahata et al., 2000). Similarly, for sediments retrieved from the eastern  
1214 equatorial Pacific, Pedersen (1983) identified peak sedimentary organic carbon content (C<sub>org</sub>) in  
1215 association with the LGM. This study used the iodine:C<sub>org</sub> ratio to rule out enhanced preservation  
1216 as a causal factor and thus ascribed the peak C<sub>org</sub> to a significant increase in productivity during  
1217 the LGM. Using biogenic barite fluxes as a palaeo-productivity indicator and conventional  
1218 stratigraphically-derived MARs, Paytan et al. (1996) also identified ~ 2 x higher export  
1219 production during glacial stages for central and eastern equatorial Pacific cores.

1220

1221 More recent studies reporting increased glacial stage export production, in the western equatorial  
1222 and subarctic Pacific, include those of Zhang et al. (2007) and Amo and Minagawa (2003),  
1223 respectively. Zhang et al. (2007), using a mixture of micropalaeontological and foraminiferal  
1224 isotopic ( $\delta^{13}\text{C}$ ) proxies from Hole 807A (Ontong Java Plateau, 3°36.42N, 165°37.49E), find that  
1225 productivity has been greater during glacial stages and has gradually increased since MIS13 (~  
1226 530 ka). They also find correlation between their palaeoproductivity record and the dust flux  
1227 record in northwestern Pacific core V21-146 (37°N 163°E), from the Shatsky Rise (Hovan et al.,  
1228 1991). Zhang et al. (2007) thus suggest that dust transported from East Asia may be significant  
1229 for enhancement of biological productivity in the western equatorial Pacific during glacial times.

1230 Similarly, Amo and Minagawa (2003) estimated temporal changes of organic carbon fluxes  
1231 during the last 130 ka from marine and terrigenous sources for sediments at the Shatsky Rise  
1232 (33.3 °N, 159 °E). In contrast with the minor input of terrestrially-derived carbon,  $C_{\text{marine}}$  was  
1233 estimated at > 86% of the total organic carbon for the whole core, reaching highest accumulation  
1234 rates (9.2 mg C cm<sup>-2</sup> ka<sup>-1</sup>) in the LGM, and next highest rates in earlier glacial periods. From  
1235 their alkenone-derived SST reconstructions, Amo and Minagawa (op. cit.) show that although  
1236 cold sea surface temperatures (SSTs) prevailed at their Shatsky Rise site throughout MIS 2-4 (~  
1237 20 – 7- ka), export production peaked at the LGM, when aeolian dust flux to the site was also at  
1238 a maximum. It should be noted that both this study and that of Zhang et al. (2007) use  
1239 conventional  $\delta^{18}\text{O}$ -derived chronologies for calculation of MARs.

1240

1241 However, recent work on equatorial Pacific sediments has contradicted these findings. Revising  
1242 their previous conclusions, Paytan et al. (2004) reported that with normalisation of sediment  
1243 accumulation rates using <sup>230</sup>Th, the apparent glacial increases in Pacific productivity either  
1244 disappeared or even reversed. They conclude that higher rates of biogenic flux during glacial  
1245 stages result not from greater export production but from increased lateral influx of sediment,  
1246 transported by bottom water currents. This problem of sediment focusing appears to be  
1247 widespread across the equatorial Pacific region, as indicated by ratios in excess of 1 for  
1248  $\frac{^{230}\text{Th}_{\text{measured}}}{^{230}\text{Th}_{\text{expected}}}$  (water column production). In another approach, using calcite  
1249 accumulation rates (excess <sup>230</sup>Th-normalised and corrected for dissolution), Loubere et al. (2004)  
1250 also estimated significantly reduced export production in the eastern equatorial Pacific (from 0.1  
1251 to 3.1 °S and 84.65 – 95.65 °W) at the LGM, i.e. ~ 30-50 % lower than during the Holocene.  
1252 They attribute this reduction to lowered nutrient supply from the Southern Ocean via the  
1253 equatorial undercurrent. This magnitude of productivity reduction is echoed by Bradtmiller et  
1254 al.'s (2006) analysis of <sup>230</sup>Th normalized opal fluxes and <sup>231</sup>Pa:<sup>230</sup>Th ratios in eleven equatorial  
1255 Pacific cores; Holocene opal burial rates those of the late glacial period by 35 %. Rather than  
1256 systematic reduction in productivity during glacial stages, Anderson et al. (2008) report that  
1257 <sup>230</sup>Th-normalized barite fluxes (equatorial core, TT013-PC72, 140 °W) exhibit no clear glacial–  
1258 interglacial pattern of variability (fig. 19). Similarly, whilst acknowledging the absence of any  
1259 ‘ideal’ palaeo-productivity proxy, Ziegler et al. (2008) identify no relationship between enhanced  
1260 flux of Asian dust and ocean productivity, estimating similar export production rates at the LGM

1261 as at present day (for five equatorial cores retrieved at 140 °W). Nor do they find any relationship  
1262 between palaeo-productivity and potential Fe supply by upwelling from the equatorial  
1263 undercurrent.

1264

1265 For the subarctic, northwestern Pacific (ODP 882, 50.3 °N, 167.5 °E), Jaccard et al. (2005) report  
1266 reduced export production during glacial stages (defined using a  $\delta^{18}\text{O}$ -derived age model), based  
1267 on use of 'biogenic barium' as a palaeoproductivity proxy (i.e. sedimentary barite concentrations  
1268 normalized by detritally-sourced aluminium). They attribute this glacial stage reduction in  
1269 productivity to increased glacial ocean stratification, and reduced nutrient supply from  
1270 upwelling. Their findings are supported by Shigemitsu et al. (2007) who report increased aeolian  
1271 fluxes of loess during glacial stages but reduced marine productivity for another western  
1272 subarctic Pacific core (50°N, 164.9°E) . Similarly, low glacial productivity (succeeded by  
1273 increased deglacial productivity) is reported by Gebhardt et al. (2008), using biogenic opal and  
1274 chlorins as proxy indicators for another piston core (51.27°N, 167.7 °E) near to ODP Site 882.

1275

1276 At present, therefore, where sedimentation rates have been normalized by excess  $^{230}\text{Th}$ , evidence  
1277 suggests there has been no enhancement of ocean productivity and export production in the  
1278 equatorial Pacific during the last glacial stage (fig.19). For the NW Pacific, varying with choice  
1279 of proxy, contradictory evidence currently exists. As suggested by Gebhardt et al. (2008),  
1280 productivity changes may have differed in time and space across the open ocean. However,  
1281 selection and interpretation of productivity proxies in the Pacific region remain an ongoing issue  
1282 and, as noted by Ziegler et al. (2008), it may be problematic to infer a change in total  
1283 productivity from a change in any one particular biological component.

1284

1285 Figure 19.

1286

1287 *5.2. Southern Ocean.*

1288 In terms of export production, Martínez-García et al. (2009) report good correlation between  
1289 glacial/interglacial changes in total organic carbon and the estimated fluxes of dust, indicating  
1290 enhanced, and possibly dust-related, marine productivity in the subantarctic eastern Atlantic  
1291 sector of the Southern Ocean during glacial stages (fig. 18). However, despite the high fluxes of

1292 dust recorded by the Antarctic ice cores during the LGM, glacial increases in TOC appear much  
1293 greater during previous glacial stages. For example, while the LGM is characterised by ~ 2 x  
1294 increase in TOC, increases of up to ~ 28 x are observed during MIS 10 (~ 360 ka) and earlier  
1295 glacial stages. One feature common to all the glacial intervals recorded in this composite record  
1296 is that increases in export production (i.e. above the average interglacial value) only occurred  
1297 when the ice core record of CO<sub>2</sub> concentration fell below ~ 230 ppmv (i.e. > 50 ppmv lower than  
1298 during interglacial stages).

1299

1300 Figure 20

1301

1302 Enhanced export production in the glacial subantarctic has previously been reported, based on a  
1303 range of radionuclide proxies, for each sector of the Southern Ocean, but varying spatially across  
1304 the sectors (Kumar et al., 1995; Pondaven et al., 2000; Dezileau et al., 2003; Chase et al., 2001,  
1305 2003). At the present day, biogenic opal accumulates at similar rates in each sector, in the  
1306 permanently open-ocean zone of the southern part of the Antarctic Circumpolar Current (ACC),  
1307 with productivity maxima south of the modern Polar Front (fig. 20). During the Last Glacial  
1308 Maximum, opal accumulation rates decreased in the southern part of the ACC at similar rates in  
1309 all three sectors, and increased at regionally different rates in the northern part of the ACC, with  
1310 maxima in the Atlantic sector (Diekmann, 2007). This suggests that limitation on primary  
1311 productivity was overcome to a greater extent in the Atlantic than in the Indian and Pacific  
1312 sectors (fig. 19). For the LGM, thorium-normalised lithogenic MARs were ~ 4 x higher in the  
1313 Atlantic than in the Pacific sector (Chase et al., 2003). Glacigenic and bottom water-transported  
1314 sediments are likely, however, to have contributed most of the fine-grained terrigenous material  
1315 to this sector (Diekmann, 2007; Maher and Dennis, 2001; Latimer and Filipelli, 2001; Tagliabue  
1316 et al., 2009).

1317

1318 Figure 20

1319

1320 *6. Synthesis.*

1321 *6.1 LGM dust fluxes and changes in ocean productivity?*

1322 It is clear from the palaeo-data that dust fluxes to the HNLC regions of the world ocean were  
1323 higher during the LGM than at the present day. The key question arises of whether or not these  
1324 increased mineral fluxes resulted in significant changes in marine productivity, export production  
1325 (i.e. carbon sequestration, biomass that is permanently removed from the carbon cycle) and,  
1326 consequently, drawdown of atmospheric CO<sub>2</sub> during the last glacial stage.

1327

1328 As summarized above, a growing body of evidence suggests that there has been no enhancement  
1329 of ocean productivity and export production in the equatorial Pacific during the LGM. For the  
1330 NW Pacific, the evidence appears contradictory. The differences in data interpretation for this  
1331 latter region may reflect heterogeneity in ocean productivity through time and space (Gebhardt et  
1332 al., 2008), but choice and understanding of productivity proxy also remain an ongoing issue.  
1333 Further, the relationships between productivity and export of carbon remain poorly resolved for  
1334 this region. Data reported by Hays (2009), for example, indicate lower productivity at the LGM  
1335 than in the Holocene but higher LGM export of carbon due to significant re-organization of the  
1336 epipelagic food web, in particular the glacial dominance of more productive, deep-living (> 200  
1337 m water depth) radiolarian species (especially *Cycladophora davisiana*), as previously reported for  
1338 the Sea of Okhotsk (e.g. Hays and Morley, 2004).

1339

1340 For the Southern Ocean, coupling between dust flux and export production is inferred by  
1341 Martinez-Garcia et al. (2009), who note that the observed increase in export production during  
1342 glacial stages is non-linear, rising only slowly in early glaciation stages before accelerating to  
1343 peak at glacial maxima. They attribute this exponential pattern to dust-driven iron supply, with  
1344 dust flux also rising exponentially as global ice volume increases (Ridgwell and Watson, 2002;  
1345 Lambert et al., 2008). These data would indicate that dust-forced increases in Southern Ocean  
1346 export production can induce significant CO<sub>2</sub> drawdown but only during glacial maxima, and  
1347 thus can account for ~ 40 -50 ppmv drawdown of CO<sub>2</sub> (i.e. as much as half of the total  
1348 interglacial-glacial change of ~ 100 ppmv). The apparent coherence between the productivity  
1349 proxy (alkenones) and the record of change in atmospheric CO<sub>2</sub> is notable (fig. 21c). Intervals  
1350 where the relationship between dust flux, productivity and CO<sub>2</sub> break down (e.g. ~ 220 ka) seem  
1351 quite rare but may provide some clue to the required additional causal factors in CO<sub>2</sub> drawdown.  
1352 The coupling of dust flux and CO<sub>2</sub> in the reverse direction (i.e. glacial to interglacial) is less

1353 clear. Rothlisberger et al. (2004) note that only small CO<sub>2</sub> variations accompanied large  
1354 variations in the flux of non-sea salt-Ca<sup>2+</sup> (as a proxy for dust) during glacial Antarctic warm  
1355 events A1 to A4. They estimate that decreased dust deposition to the Southern Ocean during  
1356 glacial to interglacial transitions accounts for not more than a 20 ppmv increase in CO<sub>2</sub> (with  
1357 changes in the North Pacific adding <8 ppmv CO<sub>2</sub>).

1358

1359 As yet, productivity proxies have yet to be quantified; there is no linear relationship between, for  
1360 example, the Southern Ocean alkenones concentrations and palaeo-productivity. It is possible,  
1361 however, to estimate some upper and lower bounds on the extent of productivity increase needed  
1362 to account for the timing and fall in atmospheric CO<sub>2</sub>, as recorded in the Antarctic ice cores (Box  
1363 1).

1364

1365 Similar calculations can be made for other waters. For example, if the drawdown of CO<sub>2</sub> was  
1366 attributable to the HNLC waters of the NW Pacific as well as the Southern Ocean, then at ~15 %  
1367 of the global NPP, the estimated glacial changes to NPP and/or pump efficiency would be  
1368 proportionally smaller. Regardless of whether interglacial-glacial changes in atmospheric pCO<sub>2</sub>  
1369 are due to a change in pump efficiency, a change in NPP or a combination of the two, the  
1370 simplest case is for pump efficiency in the Southern Ocean to be towards the upper end of the  
1371 probable range (i.e. closer to 20 % than to 1 %). Lack of observational data (as opposed to  
1372 modelled estimates) prevents further resolution of the terms here and we suggest this is a high-  
1373 priority goal for future research.

1374

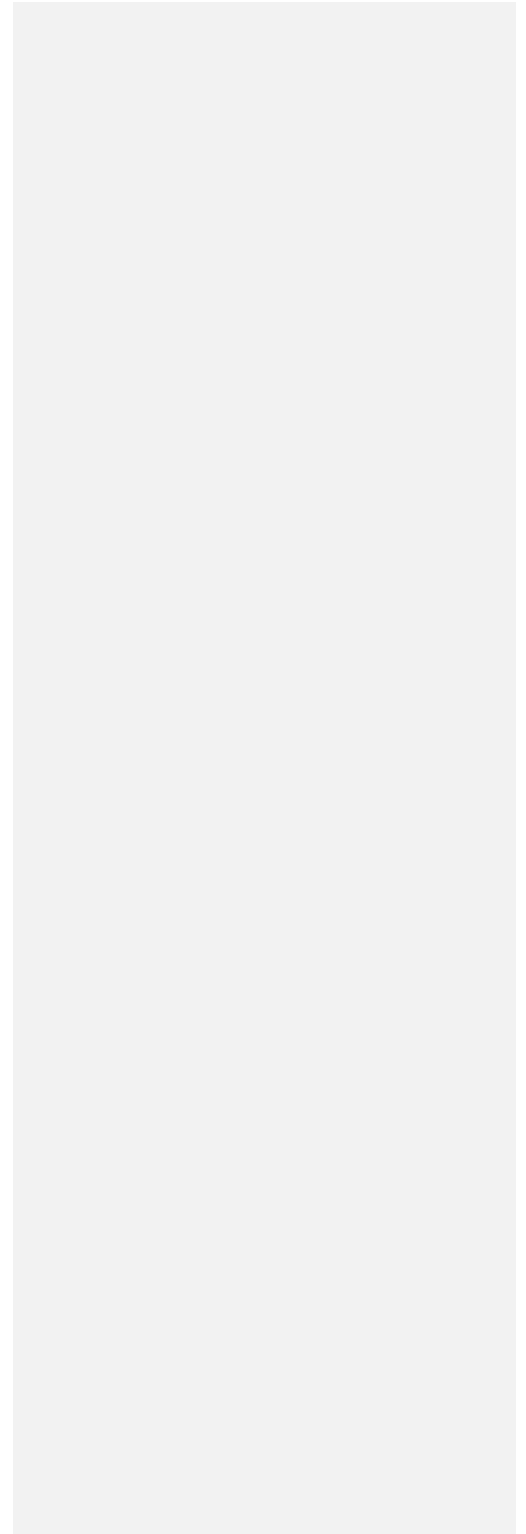
1375 Figure 21.

1376

1377 However, whereas Martinez-Garcia et al. (2009) infer that dust-forced iron fertilization has been  
1378 operating in the subantarctic region over at least the last 1.1 My, the spatial distribution of LGM  
1379 increases in ocean productivity does clearly vary with sector. LGM opal accumulation rates  
1380 increased in the northern part of the ACC but were greatest in the Atlantic sector of the Southern  
1381 Ocean, coincident with highest detrital sediment accumulation rates. Rather than of aeolian  
1382 origin, much of this terrigenous flux is likely to have been of glaciogenic and/or bottom water-  
1383 transported origin (Diekmann, 2007; Maher and Dennis, 2001). Further, even within this sector -

1384 much of which lies directly downwind of the S. American dust source -, it is suggested that areas  
1385 of enhanced LGM productivity existed as rather localized 'hotspots' (Anderson et al., 2008), and  
1386 that these 'hotspots' may reflect localized iron inputs, such as reactive (i.e. more bioavailable),  
1387 nanoparticulate iron from iceberg rafting (Raiswell et al., 2008).

1388



BOX 1: Estimations of changes in net primary productivity and/or efficiency of the biological pump at glacial terminations.

1. Global net primary productivity (NPP) is  $\sim 105 \text{ Pg C per year}$ , with  $\sim 50 \text{ Pg}$  of this in the oceans (e.g. Field et al., 1998). The efficiency of the biological pump is contentious (e.g., Archer and Johnson, 2000; Moore and Doney 2007; Jin et al., 2008); estimates span a wide range from  $< 1$  to  $20 \%$  of NPP being exported, depending both on the location and definition of 'exported'. Thus, modern carbon export may lie between  $0.5 - 10 \text{ Pg C y}^{-1}$ .
2. From the onset of glaciations, it takes at least 10,000 years for  $\text{CO}_2$  to drop by 100 ppm, such a decrease representing  $\sim 220 \text{ Pg carbon}$ , or  $\sim 0.02 \text{ Pg C y}^{-1}$ .
3. Assuming that the decrease in atmospheric  $\text{CO}_2$  was due solely to enhanced export from the Southern Ocean, bounding conditions can be established by considering two end member cases. The first end member case (see 4. below) is when NPP changes and pump efficiency is constant; the second end member case (see 5. below) is when NPP is constant and the pump efficiency changes.
4. Modern primary production in the Southern Ocean (waters south of  $50^\circ\text{S}$ ) is  $\sim 2 \text{ Pg C y}^{-1}$  (Arrigo et al., 2008). At the termination of a glacial stage, with a  $1 \%$  pump efficiency, total export =  $0.04 \text{ Pg C y}^{-1}$ , i.e. the 'baseline' export of  $1 \%$  of the modern NPP ( $2 \text{ Pg}$ ) =  $0.02 \text{ Pg C y}^{-1}$ , plus the  $0.02 \text{ Pg C y}^{-1}$  'extra export' required to reduce atmospheric  $\text{CO}_2$  to its observed glacial value. This would require NPP at the glacial termination to have been  $4 \text{ Pg C y}^{-1}$ , i.e. a *100 % increase over the modern NPP value*.
5. Conversely, at a glacial termination, with  $20 \%$  pump efficiency, the total carbon export of  $(0.4 + 0.02) \text{ Pg C y}^{-1}$  would have required an NPP of  $2.1 \text{ Pg C y}^{-1}$ ; i.e. just a *5% increase over modern NPP*.
6. In the case of constant NPP (at  $2 \text{ Pg C y}^{-1}$ ), the total required glacial export in the Southern Ocean of  $0.04 \text{ Pg C y}^{-1}$  indicates a pump efficiency of  $1.3 \%$  ( a  $30 \%$  increase over a baseline  $1\%$  efficiency), while total export of  $1.22 \text{ Pg C y}^{-1}$  gives a pump efficiency of  $20.3 \%$  ( $\sim 2 \%$  increase over a baseline  $20 \%$  efficiency).



1391 In summary, changes in Southern Ocean marine productivity, during the later stages of the LGM,  
1392 may have contributed up to half the recorded glacial fall in atmospheric CO<sub>2</sub>. These changes may  
1393 have been linked directly and causally with increased S. American dust fluxes, but are also likely  
1394 to reflect iron supply from other detrital sources.

1395

#### 1396 *6.2 LGM dust fluxes and changes in radiative forcing?*

1397 Dust flux data retrieved from sediment archives (e.g. figs. 8, 10a, 14) demonstrate the major,  
1398 global increase in ‘dustiness’ at the LGM, with at least some evidence of coherent MAR  
1399 increases between both hemispheres, even at sub-millennial timescales. From the terrestrial loess  
1400 sequences in particular, very high dust fluxes indicate marked increases in dust source ‘strength’,  
1401 i.e. dust supply increasing due to significant expansion of source areas. Provenance studies of  
1402 loess shows that this expansion reflects desiccation and expansion of arid and semi-arid areas  
1403 (such as the American Great Plains, the Russian steppe), as well as continental- (northern  
1404 hemisphere) and regional-scale (southern hemisphere) glaciation, acting to supply freshly ground  
1405 rock particles for deflation and transport at and beyond the ice sheet margins (Hughes, 1992).  
1406 Modelling studies also indicate that expansion of dust source areas is required to simulate the  
1407 observed LGM fluxes. Using linked terrestrial biosphere, dust source, and atmospheric transport  
1408 models to simulate the dust cycle, Mahowald et al. (1999), for example, show that the simulated  
1409 increase in high-latitude dustiness requires the expansion of unvegetated areas, especially in the  
1410 high latitudes and central Asia. This expansion results from increased aridity and lowered  
1411 atmospheric CO<sub>2</sub>, indicating the impact of vegetation feedbacks in modulating the atmospheric  
1412 dust sources and fluxes. Visual comparison of the LGM modelled and measured dust fluxes (fig.  
1413 10), however, shows that while the dust cycle models simulate far-field dust deposition  
1414 moderately well, they frequently under-estimate dust loadings close to dust sources, and over the  
1415 continents in general, sometimes by up to an order of magnitude. Some of this discrepancy has  
1416 been resolved by a twofold approach: firstly, by incorporating dust into a GCM and secondly, by  
1417 including in dust cycle models a number of glaciogenic dust sources, designated by inversion  
1418 from the palaeo-data (fig. 10b, and Mahowald et al., 2006). An important caveat is that dust  
1419 cycle models do not commonly represent particles at the coarser end of the size distribution (> 10  
1420 µm). Present day observations of dust are often similarly constrained; many aerosol studies use  
1421 cascade samplers with a cut-off at 2 µm. Thus, in making comparisons between model outputs

1422 and dust observations and/or archives (e.g. sediment cores), it is important that size fractions are  
1423 matched (e.g. Mahowald et al., 2006). Nevertheless, current dust cycle models have yet to  
1424 adequately compute deposition fluxes in regions including or in the vicinity of dust sources (as  
1425 above). The converse problem is that the models appear to over-estimate dust fluxes in the  
1426 Southern Ocean, where dust deposition (to the Kerguelen Plateau) is up to 2 orders of magnitude  
1427 lower than in most current models (Wagener et al., 2008).

1428

1429 Complementing these increases in supply of dust, the enhanced transport of that dust can readily  
1430 be invoked, given – for example - that the Southern Ocean westerly winds are likely weaker  
1431 and/or shifted equatorward during glacial periods compared to interglacial periods (Toggweiler  
1432 et al., 2006). However, enhanced transport can be ascribed not only to more vigorous  
1433 atmospheric circulation. Changes in transport path and removal mechanisms are also likely to  
1434 have occurred, e.g. reduced rain-out (leading to increased atmospheric residence times),  
1435 shortening of trajectory from an expanded source, and/or shifting of the atmospheric jet. Such  
1436 transport factors may also have resulted in regional-scale variations in dust particle size,  
1437 rendering subsequent inferences of wind speed changes subject to some degree of caution. Based  
1438 on changes in dust particle size, LGM increases in wind speed have previously been inferred for  
1439 northern and central China, Greenland, NW Africa, southern S. America and Antarctica.  
1440 However, more complicated particle size/transport path relationships are evident, for example,  
1441 for the dust record of the NW Pacific (despite its ‘intermediate’ position between the Chinese  
1442 loess and the Greenland ice cores), and indeed across the Antarctic Plateau.

1443

1444 Intensified cyclonic activity in northern middle latitudes has been modelled for the LGM (e.g.  
1445 Shin et al., 2003; Bush and Philander, 1999) although with some southerly displacement ( $\sim 3^\circ$ )  
1446 of the core of the jet. Depending on the reconstructed ice sheet topography (Peltier, 1994), the  
1447 Laurentide ice sheet also splits the jet stream in the above-cited models; such a split might have  
1448 resulted in a more directly coupled transport path from the Chinese dust sources to the Greenland  
1449 ice sheet.

1450

1451 In radiative terms, models have indicated rather different outcomes arising from increased glacial  
1452 dust fluxes. For example, enhanced glacial fluxes of dust may have increased albedo resulting in

1453 marked net cooling (Claquin et al. (2003); Mahowald et al. (2006)), in turn reducing  
1454 convective intensity and precipitation (Miller et al, 2004), and diminishing poleward transport  
1455 of heat and moisture (Ivanochko et al., 2005). In direct contrast, Overpeck et al. (1996) suggested  
1456 that high atmospheric dust loadings during glaciations may have resulted in episodic, regional-  
1457 scale warming (of the order of  $\sim 5$  °C), downwind of the major Asian and ice-margin dust source  
1458 regions. The modelled degree of warming was greater at progressively higher latitudes, reaching  
1459 peak values (+2.4 °C) for high-albedo (snow- and ice-covered) areas with high atmospheric dust  
1460 loadings. Over such bright surfaces, the effects of dust absorption may dominate those of  
1461 scattering, thus reducing short wave backscattering and enhancing long wave radiation (*see*  
1462 *section 3.2.2*). The model simulations reported by Overpeck et al. (1996) also show glacial-stage  
1463 warming patterns which differ markedly from the pattern of the applied, dust-induced radiative  
1464 forcing. These differences likely reflect the interactive role of feedbacks, as atmospheric  
1465 pressure, circulation and cloud patterns are altered, in modifying the regional energy balance.  
1466 These authors note that these dust-driven radiative impacts may be under-estimated by the  
1467 model, given the spatial and temporal variability of flux changes traced by the palaeo-dust  
1468 record. Nor did the simulations include the radiative effects of deposited dust on snow and ice,  
1469 which may significantly amplify warming effects (Peltier and Marshall, 1995; Krinner et al.,  
1470 2006). Sub-millennial variability in dust fluxes may in turn reflect feedback-induced oscillations;  
1471 for example, in the case of the Asian region, through modifying land-ocean thermal gradients,  
1472 resultant monsoonal intensity and extent and type of vegetation cover in the semi-arid zone.  
1473 Similarly, dust-induced warming in the Southern Ocean region may have caused glacial retreat in  
1474 the southern S. America dust source areas, cutting off dust supply as glaciers terminated and  
1475 deposited sediment in proglacial lakes rather than as unvegetated debris on outwash plains  
1476 (Sugden et al., 2009).

1477

1478 Across a range of palaeo-dust records, it appears that dust fluxes peak just prior to  
1479 Dansgaard/Oeschger oscillations, before and during Heinrich events, and just before glacial  
1480 terminations (Rothlisberger et al., 2002). The Southern Ocean records suggest that up to half of  
1481 the glacial-interglacial change in atmospheric CO<sub>2</sub> can be accounted for by dust-driven changes  
1482 in marine productivity. It seems possible that radiative forcing by dust can provide an additional

1483 key feedback, producing warming above and beyond the extensive bright surfaces – whether ice  
1484 and snow or bare soil and desert - of the glacial world.

1485

1486 *7. Conclusions.*

1487 Dust is important because not only can it affect climate, but the generation and transport of dust  
1488 is itself extremely sensitive to climate (fig. 1, fig. 22). For the future, dust cycle models predict  
1489 large changes in aeolian transport from the continents to the oceans over coming centuries, in  
1490 response to anthropogenic climate change. It is thus important and timely to evaluate the extent  
1491 to which changes in dust fluxes and properties may modulate or amplify global warming.

1492 Because spatial and temporal coverage of data on dust properties and emissions is limited at the  
1493 present day, most of the currently reported dust budget values are based on transport models. The  
1494 utility of these models and simulations is limited by uncertainties in our knowledge and  
1495 understanding of dust sources and characteristics, and of atmospheric transport and dust removal  
1496 and reaction processes (fig. 22). For example, regional-scale variations in dust composition,  
1497 shape, particle size and surface roughness introduce significant uncertainties to the problem of  
1498 modelling dust radiative forcing. At present, direct radiative forcing by dust appears to result in  
1499 cooling over much of the Earth's surface; bright surfaces (such as deserts and ice and snow) are  
1500 the exception, where dust can exert significant warming effects. *Improved modelling of dust*  
1501 *optical properties at solar wavelengths can only be achieved with improved data on present and*  
1502 *past regional variations in aerosol size and shape distributions, and in iron oxide*  
1503 *concentrations, mineralogies and mixing state.*

1504

1505 Other types of dust cycle model are being developed in order to estimate the biogeochemical  
1506 effects of dust transport to the oceans, through the supply of otherwise limited nutrients to  
1507 marine phytoplankton. For example, an observational dust proxy (ocean surface concentrations  
1508 of dissolved aluminium), a biogeochemical element cycling ocean model and a global dust  
1509 entrainment and deposition model have recently been used in semi-independent combination to  
1510 constrain dust deposition to the oceans (Han et al., 2008). However, even with major recent  
1511 expansion of the ocean Al database, the present day spatial and temporal dust proxy coverage  
1512 remains inadequate for resolving global dust climatology. This scarcity of data on dust fluxes  
1513 and properties is an issue to be addressed with new field campaigns. In the case of ocean Al, new

1514 sampling transects across the Pacific and Indian Oceans (within the CLIVAR and GEOTRACES  
1515 programs), will begin to address one aspect of this problem. However, the most robust estimates  
1516 of past and present dust fluxes are likely to be achieved only with use of multiple observational  
1517 datasets (at optimized spatial resolution) to constrain dust cycle models (Cakmur et al., 2006).  
1518 The DIRTMAP3 data (figs. 9a and 10a) clearly delineate those regions which remain severely  
1519 under-represented in terms of any observed dust fluxes for the present day and/or the LGM. Such  
1520 under-represented sites include much of the southern hemisphere, both terrestrial and marine.  
1521 Even for those sites encompassed by DIRTMAP3, very few data exist regarding the key dust  
1522 properties associated with the known sources and fluxes. *Enhanced collaboration between the  
1523 modelling (radiative and biogeochemical) and data-gathering scientific communities can  
1524 optimize design of new field campaigns to obtain key observational data both for the present day  
1525 and for past time- (and climate-) slices.*

1526  
1527 In order to achieve the required spatial and temporal information on dust fluxes, mineralogy,  
1528 particle size and shape, effective and economically-feasible means of analysis are required. In  
1529 terms of mineralogy, rapid elemental analysis using x ray-based techniques such as particle  
1530 induced x-ray emission (PIXE), x ray fluorescence and automated scanning electron microscopy  
1531 (with energy dispersive x ray analysis, e.g. 'QEMSCAN'), continue to evolve in terms of sample  
1532 preparation, and lowering of analysis cost. For the iron-bearing components of dust, often  
1533 nanoscale in size and present in minor or trace concentrations, non-destructive magnetic methods  
1534 may provide both a sensitive and fast means of analysis. Regarding particle size, the first  
1535 challenge for improving data on dust particle size distributions, whether at present day or for the  
1536 palaeo-record, is the need for expanded geographic and temporal coverage. The second requires  
1537 a degree of instrumental convergence between the atmospheric and oceanic communities. In  
1538 particular it is notable that in the oceanic and ice-core community particle sizes are commonly  
1539 measured with instruments based on the Coulter Principle (such as the Coulter Multisizer); such  
1540 instruments are rarely used in aerosol research programs. Thus it is difficult or impossible to  
1541 relate the size distributions measured by different research groups.

1542  
1543 Not only is this a problem for those measuring dust fluxes but also in turn for the modelling  
1544 community. If the cut-off size of the instruments used to measure dust fluxes is poorly defined,

1545 then meaningful comparison subsequently with modelled dust fluxes is also hindered. *Thus, if*  
1546 *one wants to say with confidence that a model is able to reproduce the main characteristics of*  
1547 *the dust deposition fields at present and during the LGM period, then a strategy for measurement*  
1548 *and treatment of dust particle size distributions is needed which is agreed and implemented*  
1549 *across both the data-gathering and dust modelling communities.*

1550

1551 *Similarly, just as convergence in analytical methods is at least desirable across the aerosol and*  
1552 *palaeo-dust research communities, similar discussions between the aerosol, biological and*  
1553 *palaeo-dust communities might aid development of the most appropriate analytical approaches*  
1554 *to assessing the solubility and bioavailability of iron-bearing dusts.*

1555

1556 The large changes in dust emissions and transport seen from the palaeo-dust record may reflect a  
1557 variety of processes (fig. 22): changes in sources and source conditions; changes in vegetative  
1558 cover; sub-aerial erosion of emergent continental shelves; deflation from periglacial deposits;  
1559 variations in wind speed and gustiness; changed wind patterns linking sources to deposition  
1560 areas; changes in deposition along the dust transport path. For the HNLC regions of the world  
1561 ocean, the palaeo-dust record indicates increased dust fluxes at the LGM. The equatorial Pacific,  
1562 for example, saw dust fluxes ~2.5 x higher during the last glacial stage, with dust fluxes  
1563 decreasing from west to east, and towards the south. However, where sedimentation rates have  
1564 been normalized by excess <sup>230</sup>Th, no enhancement of ocean productivity and export production  
1565 has been observed in the equatorial Pacific during the LGM. The deep-sea record of the  
1566 northwest Pacific, and the Greenland ice cores, display large and rapid changes both in dust flux  
1567 and particle size from interglacial into glacial climate stages. It is notable too that large dust flux  
1568 changes are also evident *within* glacial stages, on sub-millennial timescales. Contradictory  
1569 evidence currently exists for productivity and export changes at the last glacial stage in the  
1570 northwest Pacific region, interpretations vary depending on the type of proxy used and its  
1571 interpretation. *More information is needed on the reorganization of marine ecology which results*  
1572 *from glacial changes in ocean temperature and subsequent effects on export production.*

1573

1574 Dust flux data remain notably scarce in the southern hemisphere, both for the present day and for  
1575 the past. *There is an urgent need for dust flux measurements in the Southern Oceans so that we*

1576 *can improve our understanding of dust processes in this region and to facilitate the development*  
1577 *of atmosphere-ocean models.* One practical approach to obtaining improved data coverage at the  
1578 present day would be to install air samplers on the commercial and/or research vessels travelling  
1579 between the Antipodes, South Africa, Antarctica and on islands in the Southern Ocean. For the  
1580 past, the sedimentary record recently reported by Martinez-Garcia et al. (2009) appears to  
1581 indicate up to 7 x higher dust flux at the LGM compared with the present day. This is in contrast  
1582 to the much larger increase (x 30) recorded by the ice cores of the Antarctic Plateau. In terms of  
1583 productivity, the Atlantic sector of the Southern Ocean appears to have responded most to these  
1584 increased LGM dust fluxes, with increased export production having occurred slowly during the  
1585 early glacial period before rising to a peak at the glacial maximum. This temporal pattern  
1586 suggests that dust-forced increases in Southern Ocean export production can induce significant  
1587 CO<sub>2</sub> drawdown but only during glacial maxima, and thus can account for ~ 40 -50 ppmv  
1588 drawdown of CO<sub>2</sub> (i.e. as much as half of the total interglacial-glacial change of ~ 100 ppmv).  
1589 Other drivers must be responsible for the early glacial decrease in atmospheric CO<sub>2</sub>, i.e. from ~  
1590 280 ppmv during the interglacial stage to the mid-glacial level of ~ 230 ppmv. It is possible to  
1591 estimate some upper and lower bounds on the extent of the productivity increase in a) the  
1592 Southern Ocean and b) the Southern Ocean and NW Pacific combined, in order to account for  
1593 the timing and fall in atmospheric CO<sub>2</sub>, as recorded in the Antarctic ice cores. These estimates  
1594 suggest that the efficiency of the biological pump in HNLC waters is towards the upper end of  
1595 the range (i.e. closer to 20 % than to 1 %).

1596

1597 Finally, it seems clear that enhanced collaboration between the diverse scientific communities -  
1598 observational, modelling, biological – can only expedite resolution of the key gaps which exist in  
1599 our understanding of dust sources, processes and climate impacts (fig. 22). Targeted and cross-  
1600 calibrated collection of field data, both for the present day and past time- and climate-slices,  
1601 seems a fundamental imperative. The aim must be to achieve not only robust measurement  
1602 and/or estimation of dust fluxes across currently data-sparse regions, but also enhanced  
1603 information on dust mineralogy, size, shape and surface roughness. Such data are essential both  
1604 to benchmark model simulations and indeed to frame new dust, climate and biogeochemical  
1605 models, including transient simulations and mesoscale modelling. *Until we have a good*

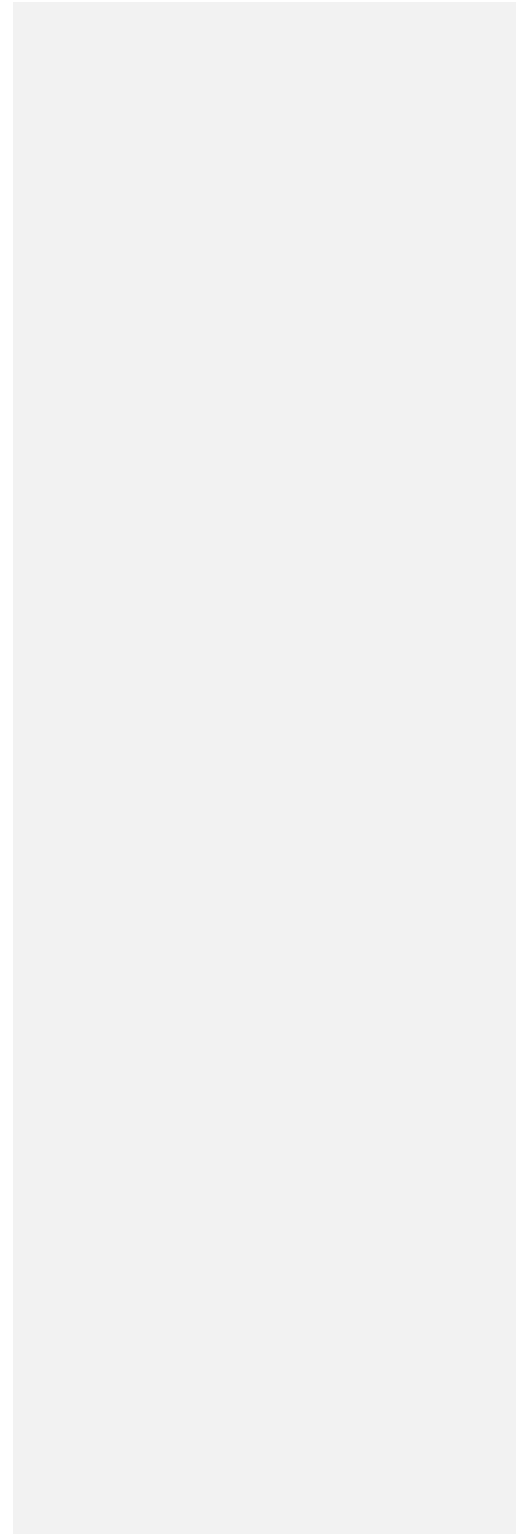
1606 *understanding of present-day processes, we will not be able to adequately address these*  
1607 *processes either in the palaeo- record or with regard to the future impacts of dust on climate.*

1608

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1616





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2714

2715 Figure captions.

2716 Figure 1. Schematic of the hypothetical glacial dust-atmospheric CO<sub>2</sub>-climate feedback system.  
2717 Different components of the Earth system can directly interact in three possible ways: positive  
2718 influences (an increase in one component resulting directly in an increase in a second  
2719 component) are indicated by red arrows; negative influences are indicated by black arrows; or no  
2720 influence. An even number (including zero) of negative influences within any given closed loop  
2721 give rise to a positive feedback, operation of which tends to amplify any initial perturbation (e.g.  
2722 the ice-albedo feedback). An odd number of negative influences gives rise to a negative  
2723 feedback, which tends to dampen any perturbation. Primary interactions are indicated by solid  
2724 lines, additional interactions dotted lines. Four main (positive) dust-atmospheric CO<sub>2</sub>-climate  
2725 feedback loops exist in this system. 1. Dust supply → ocean export production → xCO<sub>2</sub> →  
2726 temperature → ice volume → sea level → dust supply (four negative interactions). 2. Dust  
2727 supply → ocean export production → xCO<sub>2</sub> → temperature → hydrological cycle → vegetation  
2728 → dust supply (two negative interactions). 3. Dust supply → ocean export production → xCO<sub>2</sub>  
2729 → temperature → hydrological cycle → dust supply (two negative interactions). 4. Dust supply  
2730 → ocean export production → xCO<sub>2</sub> → temperature → ice volume → dust supply (two negative  
2731 interactions). From Ridgwell, 2002.

2732

2733 Figure 2. Dust deflation and entrainment processes (modified from Pye, 1987).

2734

2735 Figure 3. a) Remotely sensed aerosol optical depth measurements (Husar et al., 1997); b) The  
2736 global distribution of major dust sources and dust transport paths based on the Total Ozone  
2737 Mapping Satellite (TOMS) absorbing aerosol index (AAI) (Prospero et al., 2002). The figure is a  
2738 composite of monthly mean TOMS AAI frequency of occurrence distributions computed using a  
2739 threshold of 1.0 in the North Africa – Middle East – Asian dust belt and 0.7 everywhere else.  
2740 The arrows show the main transport paths over the oceans. The arrow size is not indicative of the  
2741 magnitude of the transport nor the distance that dust is carried. Some major points of note:

2742 During the late boreal spring, summer and fall, huge quantities of African dust are carried  
2743 to the western Atlantic and into the Caribbean and the eastern United States; also, from sources  
2744 in the Arabian Peninsula and southwest Asia to the Arabian Sea and Indian Ocean. Dust is also  
2745 carried across the Mediterranean to Europe and the Middle East.

2746 Over the Atlantic during the boreal winter and early spring, dust from the Sahara and the  
2747 Sahel is carried to South America; large quantities are also carried on occasion to the mid-  
2748 latitude Atlantic and into Europe, as indicated by the north-looping arrow.

2749 The large arrow in the North Pacific shows the main direction of large dust storms carried  
2750 from Asia every boreal spring, in some cases penetrating to North America; the dust originates  
2751 from a variety of sources including arid regions in Inner Mongolia (cross-hatched area) which do  
2752 not show up well in the TOMS product.

2753 The dust sources in the Southern Hemisphere are considerably smaller and less active  
2754 than those in the Northern Hemisphere. The most active are those in Argentina, southern Africa  
2755 and Australia.

2756

2757 Figure 4. a) Normalised size distributions of dust measured on Tenerife, Canary Islands (July,  
2758 1995) and in Puerto Rico (July, 2000) using the identical instrument, a TSI APS 3310. The  
2759 Tenerife site is ~ 300 km from the N. African coast while Puerto Rico is 5000 km to the west.



2760 The transit time for dust outbreaks is about 5 – 7 days; b) wind tunnel data (Alfaro et al., 1997)  
2761 indicating smaller modal particle sizes entrained and transported at higher wind speeds.

2762  
2763 Figure 5. Effect of the mass median diameter on dust deposition flux and on optical depth of the  
2764 particle size distribution. Optical depth expresses the amount of radiation removed by scattering  
2765 or absorption. Values are normalized to a size distribution with a mass median diameter of 4.5  
2766  $\mu\text{m}$  and a sigma value of 2.1 (Reid et al., 2003, Table 1). A log-normal distribution with a  
2767 constant width is assumed (Balkanski et al., 2007).

2768  
2769 Figure 6. Transmission electron micrographs of discrete, pedogenic ‘free iron’ grains, a)  
2770 typically acicular particles of goethite ( $\alpha\text{FeOOH}$ ), scale bar = 200 nm, b) more equidimensional  
2771 particles of haematite ( $\alpha\text{Fe}_2\text{O}_3$ ) - the smaller, electron-opaque particles -, essentially  
2772 unassociated with the larger, clean flakes of kaolinite, scale bar = 1  $\mu\text{m}$ ; c) ultrafine haematite  
2773 crystallite, soil sample, Morocco, scale bar = 100 nm. Micrographs from Maher et al. (2003),  
2774 Jones (in Schwertmann, 1991), and Lieke (pers. comm.), respectively.

2775  
2776 Figure 7. a) Atmospheric warming from absorption of radiation by dust; b) resultant net surface  
2777 cooling. Integration of satellite and ground-based data for 2002 (from Chung et al., 2005).

2778  
2779 Figure 8. Changes in glacial/interglacial dust concentrations over a pole-to-pole transect,  
2780 spanning Greenland (the NGRIP ice core record, dust mass, ppm/10, Ruth, 2005, Svensson et al.,  
2781 2000), the northern North Atlantic (Core T88, 48.3 °N, 25.0°W, high-field magnetic dust proxy,  
2782 Maher, unpub. data), equatorial Atlantic (ODP Site 633A, 1.28 °N, 11.98 °W, terrigenous %,  
2783 deMenocal et al., 1993) and Antarctica (EPICA Dome C, dust mass, EPICA Community  
2784 members, 2004).

2785  
2786 Figure 9. a) Measured aeolian MARs for the present day (for MAR data, see Appendix 1, and for  
2787 MARs, age, site and quality control data, see the updated ‘DIRTMAP3’ database (Maher and  
2788 Kohfeld, 2009, [http://www.lec.lancs.ac.uk/research/LU\\_themes/inqua\\_working\\_group.php](http://www.lec.lancs.ac.uk/research/LU_themes/inqua_working_group.php)); b)  
2789 modelled LGM MARs (Mahowald et al., 2006).

2790  
2791 Figure 10. a) Measured aeolian MARs for the LGM (for MAR data, see Appendix 1, and for  
2792 MARs, age, site and quality control data, see the updated ‘DIRTMAP3’ database (Maher and  
2793 Kohfeld, 2009, [http://www.lec.lancs.ac.uk/research/LU\\_themes/inqua\\_working\\_group.php](http://www.lec.lancs.ac.uk/research/LU_themes/inqua_working_group.php)); b)  
2794 modelled LGM MARs (Mahowald et al., 2006 and Mahowald, pers. comm).

2795  
2796 Figure 11. Typically trimodal particle size distribution of Chinese loess (Wang et al., 2007).

2797  
2798 Figure 12. Time series of North Greenland Ice core Project (NGRIP) and Chinese loess data,  
2799 synchronized on the NGRIP SS09 age scale (Ruth et al., 2003). Ice core data:  $\delta^{18}\text{O}$ , dust mass  
2800 and mean particle size (diameter, lognormal mode of volume distribution); loess data: quartz  
2801 mean diameter, from the central Loess Plateau site, Luochuan (Xiao et al., 1999).

2802  
2803 Figure 13. Particle size records for the mid-latitude Pacific over several glacial-interglacial  
2804 stages (redrawn from Hovan et al., 1991 and Zhang et al., 2007; ‘v’ in panel 3 = volcanic ash).

2805

2806 Figure 14. Time series of dust flux to the central equatorial Pacific (TTN013-PC72; 0.1°N,  
2807 139.4°W) and to the EPICA Dome C Antarctic ice core (from Winckler et al., 2008).  
2808  
2809 Figure 15. Dust fluxes at the Vostok and EPICA Dome C Antarctic ice cores (redrawn from  
2810 Lambert et al., 2008).  
2811  
2812 Figure 16. The sectors of the Southern Ocean, together with the frontal system of the Antarctic  
2813 Circumpolar Current, and the locations of the deep sea sediment and Antarctic ice cores referred  
2814 to here (modified from Diekmann, 2007).  
2815  
2816 Figure 17. Isotopic characterization (Nd:Sr) of Antarctic dusts and suite of potential source  
2817 areas, a) as reported for < 5µm potential source samples (Delmonte et al., 2004), and b) a  
2818 refinement of plot a; progress definition of the Patagonian isotopic signature and inclusion of  
2819 new samples from Australia (Revel-Rolland et al., 2006) and the Puna/Altiplano of South  
2820 America (Gaiero, 2007; Delmonte et al., 2009) to explain the non-Patagonian isotopic signature  
2821 found in the Antarctic dust.  
2822  
2823 Figure 18. Comparison of the Southern Ocean and Antarctic ice core records over the last 1.1Ma.  
2824 (a) Temperature reconstruction from EPICA ice cores (black) (Jouzel et al., 2007) and alkenone-  
2825 based SST from site PS2489-2/ODP1090 (red). Marine Isotope Stages (MIS) are shown for  
2826 reference. (b) Atmospheric CO<sub>2</sub> concentrations from the EPICA ice cores. Dashed line indicates  
2827 the CO<sub>2</sub> level when productivity starts to increase above the average interglacial value. Filled  
2828 area illustrates CO<sub>2</sub> concentrations below 230 ppmv. Glacial terminations are shown for  
2829 reference. (c) EPICA Fe flux (blue) and site PS2489-2/ODP1090 Fe flux (red). (d) EPICA  
2830 Insoluble dust (light brown) and site PS2489-2/ODP1090 long-chain odd carbon-numbered n-  
2831 alkanes (C25-35) mass accumulation rate (MAR) (blue). (e) Site PS2489-2/ODP1090 C37  
2832 alkenones MAR (green) and TOC MAR (Diekmann and Kuhn, 2002) (black). Shaded areas  
2833 highlight the high-productivity intervals, when alkenone MARs are 3 x the average interglacial  
2834 value. (Redrawn from Martinez-Garcia et al., 2009).  
2835  
2836 Figure 19. Time series of proxy records for palaeo-productivity and dust flux for the equatorial  
2837 Pacific through several glacial-interglacial cycles (Redrawn from Anderson et al., 2008).  
  
2838 Figure 20. <sup>230</sup>Th-normalised fluxes of biogenic opal in the three sectors of the Southern Ocean  
2839 (from Diekmann, 2007).  
  
2840 Figure 21. a) Global δ<sup>18</sup>O stack (Lisiecki & Raymo, 2005) versus n-alkanes concentrations from  
2841 Southern Ocean site PS2489-2/ODP1090; marine isotope stages shown; b) Temperature  
2842 reconstruction from EPICA ice cores (Jouzel et al., 2007) and alkenone-based SSTs from  
2843 PS2489-2/ODP1090; c) atmospheric CO<sub>2</sub> concentration from the EPICA ice cores (Luthi et al.,  
2844 2008; Petit et al., 1999; Siegenthaler et al., 2005) versus alkenones concentration, PS2489-  
2845 2/ODP1090.  
2846  
2847 Figure 22. Schematic of the sources, process and impacts of dust on climate, a) at the present  
2848 day, and b) at the LGM, with key gaps in data and understanding marked by ‘?’ on the diagrams.  
2849  
2850

2851 Table captions.

2852

2853 Table 1. Comparison of a range of modelled present day dust fluxes (mean annual, by region).  
2854 The unit is Tg/yr, numbers in brackets = the % of the annual mean global emission flux. (From  
2855 Tanaka & Chiba, 2006).

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