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Contemporary glacigenic inputs to the dust cycle

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ABSTRACT: The importance of glacigenic dust in the Earth's system during glacial periods is widely acknowledged. Under contemporary conditions, the world's largest dust sources are in low-lying, hot, arid regions and this is where most aeolian research is focused. However the processes of dust production and emissions are still operating in cold climate regions, particularly in proglacial areas. This paper assesses current understanding of the relationship between glacierised landscapes and dust emissions and inputs to the global dust cycle. It focuses on how elements in the glacial and aeolian geomorphic sub-systems interact to determine the magnitude, frequency and timing of aeolian dust emissions, and on feedback mechanisms between the systems. Where they have been measured, dust emission intensity and deposition rates in glacierised catchments are very high, in some cases far exceeding those in lower latitudes, however, few studies span long time scales. The impact of future glacier retreat on the balance between sediment supply, availability and aeolian transport capacity and implications for glacigenic dust emissions is also considered. This balance depends on relative spatial and temporal changes in meltwater suspended sediment concentration and wind strengths, which promote dust emissions, and patterns and rates of soil development and vegetation succession on recently-deglaciated terrain which protect sediments from deflation. Retreat of the Antarctic ice sheet could mean that in future glacigenic contributions to the dust cycle exceed those of non-glacigenic sources in the southern hemisphere. Copyright © 2012 John Wiley & Sons, Ltd.

KEYWORDS: glacierised landscape; dust cycle; ice retreat; dust deposition

Introduction

Fine particles (dust, <100 μm) play an important role in the earth–atmosphere–ocean system. Calculations of global dust emissions are very variable, but estimates of the amount entering the atmosphere each year converge around 2000 Tg, of which 1500 Tg is deposited on land and 500 Tg in the oceans (Shao *et al.*, 2011). In the terrestrial system, the erosion or deposition of fine particles affects soil development, nutrient cycling and geomorphic processes such as stone pavement formation (McFadden *et al.*, 1987; Li *et al.*, 2007; McTainsh and Strong, 2007; Muhs *et al.*, 2007). Within the atmosphere, mineral aerosols affect climate both directly, by scattering and absorbing solar and terrestrial radiation (Haywood and Boucher, 2000; Yoshioka *et al.*, 2007), and indirectly by affecting cloud properties, precipitation, cyclone intensity and acting as ice nuclei (Dentener *et al.*, 1996; Rosenfeld *et al.*, 2001, 2011; DeMott *et al.*, 2003; Foltz and McPhadden, 2008; Sun *et al.*, 2008). Dust can be an important source of iron (Fe) which in soluble form is an essential micronutrient for marine biota and can therefore impact primary productivity in oceans and the global carbon cycle (de Baar *et al.*, 2005; Jickells *et al.*, 2005; Wolff *et al.*, 2006; Maher *et al.*, 2010).

There is strong evidence for a close coupling between dust and climate that has been sustained through multiple glacial–interglacial cycles (Lambert *et al.*, 2008). During glacial periods dust emissions increase due to the combined effects of aridity, enhanced by a weakened hydrological cycle, strong tropospheric winds, a reduction in terrestrial biomass and extensive fine sediment availability. Higher atmospheric dust loadings in the past were spatially variable but likely to have been on average at least twice as high during the Last Glacial Maximum compared with at present, but estimates for this, and other glacial periods, range from two to over twenty times current amounts (Lambert *et al.*, 2008; Winckler *et al.*, 2008; Bauer and Ganopolski, 2010; Yue *et al.*, 2011). One of the legacies of this is extensive Quaternary loess (aeolian silt) deposits which cover approximately 10% of Earth's land surface (Pesci, 1990).

Under current conditions, the world's largest contemporary dust sources are in low-lying, hot, arid regions such as the Bodélé Depression, North Africa, the Taklimakan Basin, China and the Lake Eyre Basin, Australia (Prospero *et al.*, 2002). To date, the vast majority of aeolian research has focused on these prominent global dust sources in the tropics and mid-latitudes. However the processes of dust production and emissions are still operating in cold climate regions, particularly in proglacial areas, as is the deposition of fine particles to form modern loess deposits (Eden and Hammond, 2003; Hugenholtz and Wolfe, 2010). Contemporary glacigenic dust sources are typically associated with cold climates and are primarily, but not exclusively, located at high latitudes. Notably, modern dust emissions in high latitudes are not confined to arid regions and can occur in relatively humid areas of Alaska (Crusius *et al.*, 2011), New Zealand (McGowan *et al.*, 1996; McGowan and Sturman, 1997), Patagonia (Gassó *et al.*, 2010; Johnson *et al.*, 2011), Iceland (Arnalds, 2010; Prospero *et al.*, 2012) and Greenland (Bullard and Austin, 2011). Under modern conditions, although not as spatially-extensive as in subtropical environments, high latitude glacigenic dust emissions can be seasonally-intense and their proximity to marine environments may be important given many high latitude oceans are deficient in nutrients often found in aeolian dust (Nielsdóttir *et al.*, 2009; Crusius *et al.*, 2011).

Glaciers contribute to enhanced dustiness through sediment production during the grinding or abrasion by ice over bedrock or sediments to produce fine particles (glacial flour). These particles are transported by meltwater beyond the ice margin to floodplains from which they are deflated by strong glacier or ice-sheet-driven winds. Aeolian and glacial processes and landforms are widely studied individually but there are very few process studies that have tried to link activity within the glaciofluvial and aeolian systems to understand better the controls on the magnitude and frequency of dust emissions in glacierised catchments. This is despite their importance for biogeochemistry (Anderson, 2007), glacier-feedback mechanisms (Adhikary *et al.*, 2002) and proglacial landscape modification (Ballantyne, 2002). The aim of this paper is to assess current understanding of the relationship between glacierised landscapes and dust emissions and inputs to the global dust cycle. It focuses on contemporary relationships – several recent reviews cover other aspects of cold climate aeolian processes in more detail (Brookfield, 2011; Bateman, In press; Wolfe, In press).

The Glacier Landscape System

The relationship between glacierised landscapes and aeolian processes has long been acknowledged. Extensive periglacial dunefields and sand sheets, or cover sands, are clearly associated with positions of former ice margins and have been widely documented (Koster, 1988, 2005; Wolfe *et al.*, 2004; Bateman and Murton, 2006). Early notable studies include those by Högbom (1923) in Europe and Black (1951) and Bird (1967) in North America (Niessen *et al.*, 1984).

However, substantially more attention has focused on the importance of glaciers and ice sheets as generators of the material that forms the world's extensive loess deposits (Pye, 1995). While a broad range of processes and environments is now acknowledged to have contributed to Quaternary loess production (Wright *et al.*, 1998; Wright, 2001; Alenikoff *et al.*, 2008; Smalley *et al.*, 2009, 2011), glaciers and ice sheets are still considered to be the most significant source of this material globally and over the long term. In terms of impacts on the dust-cycle, the most important elements of the glacier landscape system are the processes leading to the sorting and transfer of fine sediments to a position in the landscape from which they can be entrained and transported by the wind to distal regions. In proglacial regions production and supply of sediments suitable for aeolian entrainment are closely-linked to glacier and ice-sheet dynamics (such as ice velocity, sediment load and basal thermal conditions) and hydrology (Figure 1). The glacier landscape system is not in equilibrium with contemporary conditions because climatological and hydrological boundary conditions change orders of magnitude faster than ice-sheet and landform geometries (Ballantyne, 2002; Tunnicliffe and Church, 2011). Sediment transport is therefore a function of storage as much as contemporaneous erosion and removal – subglacial/subaerial sediment re-mobilisation upstream in the catchment reflects past climate and events – however, because sediment delivery by meltwater beyond the ice margin to the outwash plain is directly influenced by contemporary conditions, such as melt and precipitation rates (Orwin and Smart, 2004; Jansson *et al.*, 2005; Cockburn and Lamoureux, 2008), the magnitude and frequency of aeolian processes are linked to modern-day glacier hydrology.

The proglacial, mobile, braided rivers that characterise many glacial outwash plains are therefore a key part of the system in that they couple glacial processes to the wider environment and facilitate the exchange of material between glacier-covered and non-glacier-covered parts of catchments (Warburton, 1999). The total fluvial sediment load is typically significantly greater in glacierised than in non-glacierised catchments. Hallet *et al.* (1996) found sediment yields from catchments containing warm-based glaciers were up to an order of magnitude greater than from steep, mountainous catchments without glaciers, and Hasholt (1996) reported sediment transport from glacierised areas of Greenland ranges from 84–1500 t km^{-2}a^{-1} compared with 1–56 t km^{-2}a^{-1} from non-glacierised areas. In Alaska the glacierised sections of the Tanana River Basin catchment were found to have suspended sediment loads 36 times higher than

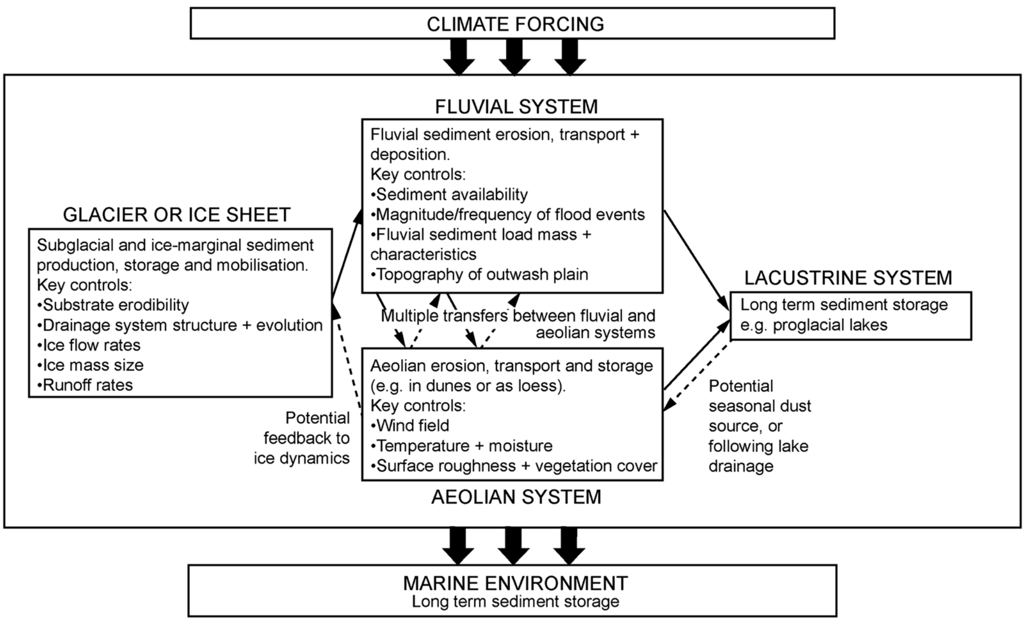

Figure 1. Simplified sediment links and storage areas within glacier-influenced landscapes. Over long timescales the marine sediment store can become a sediment source due to changes in sea level.

the non-glacierised areas (Wada *et al.*, 2011). In southern Chile suspended sediment yield is correlated with glacier cover, although the latter cannot explain the full range of variations (Pepin *et al.*, 2010). The amount of sediment produced by glaciers is closely linked to the thermal structure of the ice. Warm-based glaciers are at the pressure-melting point through the ice mass and consequently have water at the ice–bed interface. This allows rapid basal sliding and the production of fine sediments by abrasion (Hooke and Iverson, 1995). Cold-based glaciers are below pressure-melting point at the bed, which limits basal sliding and hence abrasion rates. In a comparison of glacier basins in Svalbard, Hodson *et al.* (1997) found suspended sediment yields from the warm-based Finsterwalderbreen were 710–2900 $t\,km^{-2}a^{-1}$ whereas those from the cold-based Austre Brøggerbreen were an order of magnitude less at 81–110 $t\,km^{-2}a^{-1}$.

The products of glacial abrasion combine with sediments produced through plucking, quarrying and weathering, with the result that meltwater-transported sediments typically span a wide range of grain sizes from boulders to clay (Hooke and Iverson, 1995; Lee and Rutter, 2004) and in turn proglacial outwash plains comprise a range of different sediment sizes (although are typically deficient in gravel). Only the finer sediment fractions, typically <2000 μm, are likely to be entrained by the wind which means the most important fraction of the meltwater sediment load for dust emissions is usually that travelling in suspension. Of these finer sediments, the relatively coarse (sand-sized >63 μm) particles may form sand sheets or aeolian sand dunes proximal to the ice (Calkin and Rutford, 1974; Selby *et al.*, 1974; Gisladóttir *et al.*, 2005; Zhang *et al.*, 2005) whereas silts and clays can be transported longer distances by the wind. Aeolian transport of large particles by strong winds has been reported; for example, Lewkowicz (1998) reported wind erosion of particles up to 45 mm in diameter and weighing up to 25 g each in the Canadian Arctic, but this is not the norm.

Sediment Supply and Availability

Meltwater suspended sediment supply

The aeolian system in many proglacial regions is supply-limited, i.e. dust emissions are limited by a lack of suitable sediment. Meltwater suspended sediment transport is a supply-controlled process, requiring little energy, it therefore tends to be environmentally responsive, but less predictable than hydraulically-controlled bedload transport. While many meltwater systems have a clearly defined seasonal or diurnal discharge pattern, suspended sediment fluxes are very variable and do not have a linear relationship with meltwater discharge (Collins, 1979; Hodgkins, 1999; Jansson *et al.*, 2005). The dominant control on suspended sediment concentrations is short-term (daily, weekly, seasonal) sediment availability. At the seasonal scale, this can be linked to early season flushing, late-season exhaustion or increasing sediment supply (Fenn *et al.*, 1985; Hodgkins, 1999; Haritashya *et al.*, 2006). One of the consequences of this supply control is that suspended sediment concentration is very variable within a single catchment (Hodgkins, 1996; Hodson *et al.*, 1997; Hodgkins *et al.*, 2003; Cockburn and Lamoureux, 2008) and from region to region depending on glacier hydrology, dynamics, thermal regime and geology (Gurnell and Warburton, 1990; Orwin *et al.*, 2010) making it hard to scale-up potential sediment supply to the aeolian system from short monitoring periods. Entrainment, reworking and deposition of proglacial sediments and extra-glacial catchment inputs from runoff can also affect suspended sediment concentrations. Orwin and Smart (2004) reported a 600% increase in suspended sediment concentration between the ice-margin of Small River Glacier, Canada and the distal proglacial boundary (≈1 km), and attributed it to channel storage release and extra-channel inputs (e.g. bank collapse). McGrath *et al.* (2010) found a fourfold increase in suspended sediment in the Watson River along a transect from the Greenland Ice Sheet margin to the head of Kangerlussuaq fjord (≈30 km). The glacial outwash plain through which the Watson River flows comprises mixed sediment sizes, but contains substantial quantities of material <1400 μm in diameter (Bullard and Austin, 2011), which is available for the river to entrain as additional suspended sediment load as it flows downstream. Erosion does not always dominate; Hodgkins *et al.* (2003) found net fluvial erosion and deposition in consecutive seasons at Finsterwalderbreen in Svalbard indicating substantial aggradation can also occur.

In addition to diurnal and seasonal fluctuations, meltwater sediment concentration, particularly the fine sediment component, increases during glacier surges and is often elevated during glacier-related catastrophic floods (jökulhlaups) (Pálsson and Vigfússon, 1996; Mernild and Hasholt, 2009) (Figure 2). Old *et al.* (2005) measured suspended sediment variations in the Skaftá River, southern Iceland, during two jökulhlaups; peak suspended sediment flux during the jökulhlaups was 4650 $kg\,s^{-1}$, compared with the non-jökulhlaup flux of 190 $kg\,s^{-1}$. During the first event, >61% of the suspended sediment particles were fine silts (2–20 μm).

Seasonal fluctuations in river discharge can result in supply-limited conditions at different times of the year. Continual sub-zero temperatures may limit sediment renewal during winter due to a combination of low meltwater discharge and storage of precipitation as snow. In summer, rivers can be very high covering the whole outwash plain (Hodgkins *et al.*, 2009) or cohesion among sediments may be increased by moisture due to a high water table. The flow regime of the river will be affected by its water sources; rivers in glacierised catchments may receive water from nival melt and/or glacial melt as well as being spring-, lake- or runoff-fed.

Aeolian sediment availability

In addition to supply limitations, deflation from outwash plains may also be availability-limited, typically by surface moisture content, vegetation or lag deposits that form under strong winds. Moisture is a common limit to sediment availability in arid regions because it increases particle cohesion. Moisture contents in the range 4 to 6% have been suggested to be limits to aeolian transport (McKenna Neuman and Nickling, 1989; Wiggs *et al.*, 2004), however where evaporation is high, due

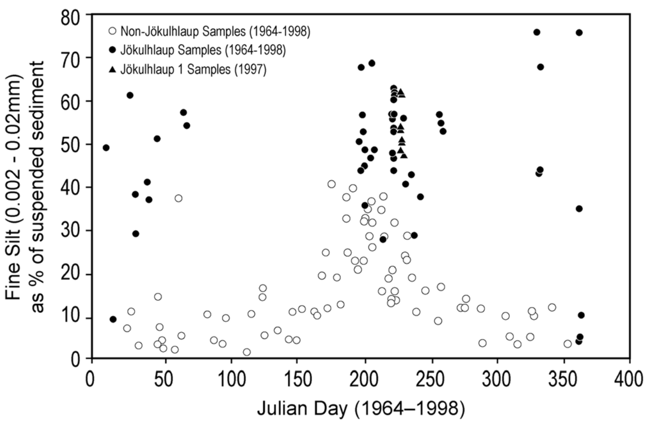

Figure 2. Particle-size distribution of suspended sediment in the Skaftá River, southern Iceland (1964–1998) compiled by Old *et al.* (2005). Reproduced by permission of John Wiley & Sons.

to high air temperatures or strong winds, the effects can be very short-lived. There are field observations of active aeolian transport occurring during rainfall events and at high levels of humidity that can be attributed to rapid desiccation of the surface layer of grains by strong winds or the effects of rainsplash ejecting particles in to the airstream (Ashwell, 1986; Jackson and Nordstrom, 1998; Marx and McGowan, 2005). The longer term role of soil moisture is usually in sustaining a vegetation cover, however, on active glacial outwash plains vegetation is often very sparse, typically <5% (Frenot *et al.*, 1998; Marteinsdóttir *et al.*, 2010).

For sediment to become available for deflation, it also needs to be deposited beyond the channel boundary, on the floodplain, on braid bars that are exposed during periods of low flow (Church, 1972; Nickling, 1978; Good and Bryant, 1985; Walker and Everett, 1991) or in ephemeral (seasonally-filled) lakes (Hugenholtz and Wolfe, 2010). Falling meltwater discharge and reduced flow velocities will deposit sediment beyond the channel and this material can either be entrained and transported by water during the next high flow event or may desiccate and be blown away by the wind (Figure 3). McGrath *et al.* (2010) measured at least five such fluctuations during the 2007 melt season in southwest Greenland, the first of which may have been responsible for depositing a 140 mm thick silt (50 µm mode) layer on the outwash plain, the rapid deflation of which was described by Bullard and Austin (2011). Deposition of fine material following high magnitude jökulhlaup events has been linked explicitly to intense dust storms in Greenland (Dijkmans and Tørnqvist, 1991) and Iceland (Prospero *et al.*, 2012). Dijkmans and Tørnqvist (1991) described aeolian deflation of a jökulhlaup deposit in southwest Greenland as follows:

> A thin layer, a few centimetres thick, of fine sand and silt had been deposited from suspension. The mean grain-size of two samples taken from the damp surface layer is 40 µm. Fair summer weather conditions caused the surface of the higher parts of the floodplain to dry up quickly, enabling eolian transport of the fine-grained material. A wind speed of only 6 m/s (measured at 2 m height) initiated small dust clouds, 10-20 m high, whereas at wind speeds of 14-18 m/s more than 100 m high dust clouds were transported from the flood plain in the direction of the sand sheets and the uplands. Due to this winnowing of the silt the mean grain-size of the residual surface material had risen to 70 µm after the dry sediment had been exposed for about one day. (p.10).

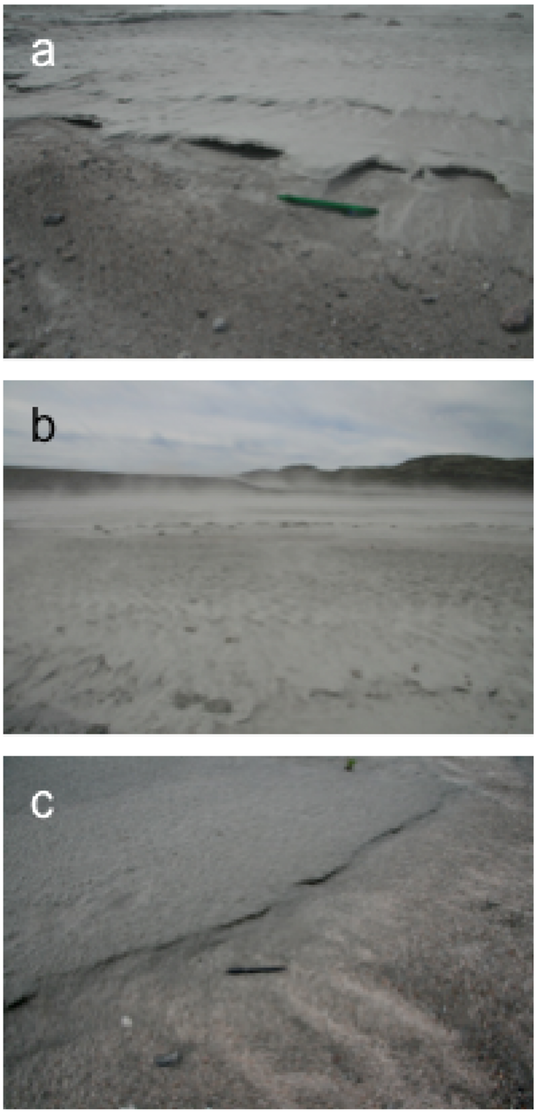

Figure 3. The Watson River floodplain, southwest Greenland. (a) Fresh deposit of silt following a meltwater flood June 2007 (modal diameter 50 µm). (b) Active deflation of the meltwater deposit. (c) Same location in August 2007, most of the silt-sized material has been deflated and a lag deposit has formed limiting further aeolian entrainment.

The winnowing process described by Dijkmans and Tørnqvist (1991) that led to a coarsening of the surface material is common and in some cases is sufficient for a lag deposit to develop that armours the surface and prevents further aeolian entrainment from occurring. Boulton and Dent (1974) described the formation of a lag surface resulting from the deflation of fine sediments from till in southern Iceland, in which the removal of the fine material led to an increase in surface coarse clast cover from 30–40% to 90% in 4 years, significantly reducing the susceptibility of the surface to wind erosion. In northeast Iceland, Mountney and Russell (2004) found that pebble lag surfaces played an important role in the structure of proglacial aeolian sand sheets. In Greenland, Bullard and Austin (2011) reported the selective removal of fine–medium silts from a flood deposit leading to the formation of a lag deposit of sand-sized particles over a period of 10 weeks (Figure 3). Such lag deposits can cause aeolian deflation to cease until wind velocities increase, or the floodplain is reworked by meltwater causing a replenishment of fine material at the surface.

Additional influences on sediment availability in cold regions include the effects of snow, frost and sublimation (McKenna Neuman and Gilbert, 1986; Law and Van Dijk, 1994; Kurosaki and Mikami, 2004). Thick snow cover can protect sediments from deflation and frost can cement fine particles preventing their erosion (Church, 1972; McKenna Neuman and Gilbert, 1986; Bristow *et al.*, 2010). Kurosaki and Mikami (2004) found that an 80% snow cover increased the threshold wind velocity for dust transport in East Asia by about 2 m s^{-1}. Alternatively frost activity may enhance surface erodibility by breaking up aggregates, limiting plant growth and moving sandy silts to the surface through freeze–thaw processes (Corte, 1971). Sublimation can also be an important mechanism for releasing frozen particles from surfaces in dry areas with cold winds and limited snow cover such as the Dry Valleys of Antarctica and southwest Greenland (McKenna Neuman, 1993; Van Dijk and Law, 2003).

Aeolian Transport Capacity

In addition to sediment supply and availability, the third key variable controlling dust emissions is aeolian transport capacity – the ability of the wind to transport sediment. The principles of aeolian sediment transport established in temperate or low latitude regions also apply to cold climate glacierised regions. For a given set of grain characteristics aeolian sediment transport is positively related to wind velocity and turbulence intensity, but negatively related to surface roughness and sediment moisture content. The threshold wind speed for aeolian transport varies with particle size, surface roughness (z_0) and turbulence, but values in the range 5–10 m s^{-1} (at 2 m above the surface; z_0 0.0007–0.002 mm) have commonly been reported from glacierised regions (Nickling, 1978; Dijkmans and Tørnqvist, 1991; Arnalds *et al.*, 2001; Marx and McGowan, 2005; Bullard and Austin, 2011). There are, however, differences between threshold wind velocities required to entrain particles under warm or cool conditions depending on air temperature, pressure and humidity, all of which affect the density of air (Selby *et al.*, 1974; McKenna Neuman, 2004). Generally for a given particle size, threshold wind velocities are lower in cold conditions and increase as temperature increases. For example McKenna Neuman (2003) calculated that the aerodynamic drag needed to entrain sand-sized particles (>200 μm) could be 30% lower in cold or high latitude areas compared with hot deserts and the mass transport rate up to 70% higher at temperatures of −40°C compared with the same wind speed at +40°C (McKenna Neuman, 2004). This difference applies to sand and there are no comparable measurements for silt and clay sized particles which are likely to behave somewhat differently; however, in areas where saltation-impact entrainment of dust (sand-blasting) is important, then this enhanced sand transport capacity may also be important for dust emissions.

Many cold climates, and particularly glacierised regions, experience very strong winds. The combination of low frictional resistance of the oceans, increase in wind speeds from the Equator to the Poles to compensate for loss of momentum derived from Earth's rotation, and strong temperature gradients between ice, land and oceans makes the Arctic and Antarctic coastal zones the windiest regions on Earth (Eldridge, 1980). Extreme wind speeds have been recorded around the margins of the major ice sheets; for example at Port Martin, Antarctica, a maximum wind speed of 90 m s^{-1} has been recorded (Périard and Pettré, 1993) while winds at Angmagssalik on the southeast coast of Greenland can exceed 50 m s^{-1} (Hedegaard, 1982). Both the Greenland and Antarctic ice sheets give rise to a cold, deep and stable boundary layer above the ice. Anticyclones dominate both, and cold air drains radially from the east central highlands of Antarctica (>4000 masl) and central Greenland (>3000 m asl) towards the coast following topographic depressions (Mather and Miller, 1967; Putnins, 1970).

Katabatic and föhn winds

Some of the strongest winds around glaciers and ice sheets are katabatic winds driven by gravity and thermal gradients. In Nansen's first crossing of Greenland (Nansen, 1890) the expedition typically travelled 10–15 km per day over the eastern and central ice sheet, but 65 km in one day on the western flank when the katabatic winds were exploited using sails rigged to sledges (Figure 4). Katabatic winds have long been associated with the aeolian transport of sediments and are important for dust storm generation and aeolian dune formation in valleys in northern Canada (Nickling, 1978; McKenna Neuman, 1990a, 1990b), Antarctica (Nylen *et al.*, 2004; Bristow *et al.*, 2010) and Iceland (Ashwell, 1966). Katabatic winds often have a strong diurnal and seasonal signature, both in terms of direction and strength. For example, at the southwest margin of the Greenland Ice Sheet, near Kangerlussuaq, in summertime during the day the air temperature over the tundra is warmer than that over the ice sheet – the thermal difference accelerates flow away from the ice reaching maximum speeds in the afternoon when the contrast is greatest (Meesters, 1994; van den Broeke *et al.*, 1994). At night the tundra surface cools more rapidly than the ice sheet and cold air over the tundra decelerates the katabatic winds. In contrast, during the winter when the tundra is snow-covered there is less thermal contrast. Air flows off the ice under the influence of gravity but because its temperature is similar to that of the tundra, it can stagnate near the ice margin (van den Broeke *et al.*, 1994).

Figure 4. Sailing a sledge using katabatic winds during the first crossing of Greenland (Nansen, 1890). Reproduced by permission of the Royal Geographical Society with the Institute of British Geographers.

In the McMurdo Dry Valleys of Antarctica, down valley katabatic winds are consistently stronger than easterly, onshore (up valley) winds from the Ross Sea/McMurdo Sound all year round, although the easterly winds are more persistent (Clow *et al.*, 1988; Doran *et al.*, 2002). Lindsay (1973) reported typical down valley winds of 11 m s^{-1} in summer compared with up valley winds of 8 m s^{-1}. In contrast, during the winter the up valley winds were much weaker (<5 m s^{-1}) than the down valley (katabatic) winds (>10 m s^{-1}). Local topography can cause significant variation in katabatic wind strength by channelling or funnelling winds along valleys, such as in the McMurdo region, but also along fjords such as on Baffin Island, Canada where Gilbert (1983) reported winter winds in excess of 40 m s^{-1}. Topography can also cause differential heating of slopes leading to spatially very variable patterns of wind speed (Doran *et al.*, 2002). The steeper the mountain or glacier slope, the stronger the katabatic wind will be (Parish and Waight, 1987) – considerable variation in ice-marginal topographic profiles therefore results in differing intensities of katabatic wind. For example, at the Burgar Oasis, Antarctica, the ice slope is 0.012 and no katabatic winds are observed (Solopov, 1967).

Although katabatic, glacier-driven winds are typically strong and capable of considerable aeolian sediment transport, their effect is often spatially-restricted. As the air flows off the ice, it is slowed down by friction over sediments or vegetation (Defant, 1951; Seppälä, 2004). Oke (1987) suggests katabatic winds associated with valley glaciers die out within 0.5 km of the glacier snout due to friction on the valley floor and the opposing force of up valley winds; katabatic winds around the Greenland and Antarctic ice sheets generally dissipate within 10–20 km of the ice margin (Tauber, 1960; Gallée *et al.*, 1995).

Föhn (aka foehn) winds are also effective agents of aeolian sediment transport in some glacierised landscapes. These occur when moist air rises over a topographic barrier such as a mountain range or high plateau, loses its moisture as precipitation and then descends on the lee side of the barrier to produce very dry, often warm, strong downslope winds. Föhn winds occur in the Dry Valleys of Antarctica reaching wind speeds of over 30 m s^{-1} and can be accompanied by an increase in air temperature of up to 40°C (Ayling and McGowan, 2006; Speirs *et al.*, 2008). Fristrup (1953) attributed much of the wind erosion in the high Arctic to föhn winds, highlighting their desiccating effect on surface moisture and the consequent increase in sediment erodibility. The föhn winds in Greenland are stronger in winter, where they can cause air temperature increases of over 20°C, than in summer. In New Zealand, wind speeds during föhn wind storms at Lake Tekapo have been recorded in excess of 30 m s^{-1} lifting particles over 1500 μm in diameter to heights over 4 m above the glaciofluvial deposits (McGowan, 1997; McGowan and Sturman, 1997).

Magnitude, Frequency and Timing of Aeolian Dust Emissions

The magnitude, frequency and timing of dust events are determined by sediment supply and the coincidence of aeolian transport capacity and sediment availability in potential source regions (Bullard *et al.*, 2011). This can result in distinct seasonal patterns of dust emissions. Using a 6-year record of dust concentration data, Prospero *et al.* (2012) determined that while dust emissions occur throughout the year in southern Iceland, dust activity is greatest in spring and early summer. The sources of dust are the extensive floodplains and glacial outwash plains along the south coast of Iceland (Figure 5). In this region of Iceland extra-glacial snow cover is generally low and thaws by early April. Suspended sediment supply from the rivers and meltwater streams peaks in April, with a second weaker peak in September (Gislason *et al.*, 1997; Old *et al.*, 2005). This lack of snow cover and a wide expanse of freshly deposited fine glaciofluvial sediment create ideal conditions for the spring dust storms. In the Slims River Valley, Yukon Territory, Canada, dust storms also occur throughout the year, generated by strong katabatic winds from the Kaskawulsh Glacier, but are most frequent from May to July (Nickling, 1978). These late spring, early summer months are when the river is at low stage exposing a large area of freshly deposited glaciofluvial silts and fine sands and in addition temperatures are relatively high and precipitation low, encouraging rapid drying of sediments (Nickling, 1978). Wilson (1958) observed that most dust transport on Cornwallis Island, Canadian Arctic was in spring.

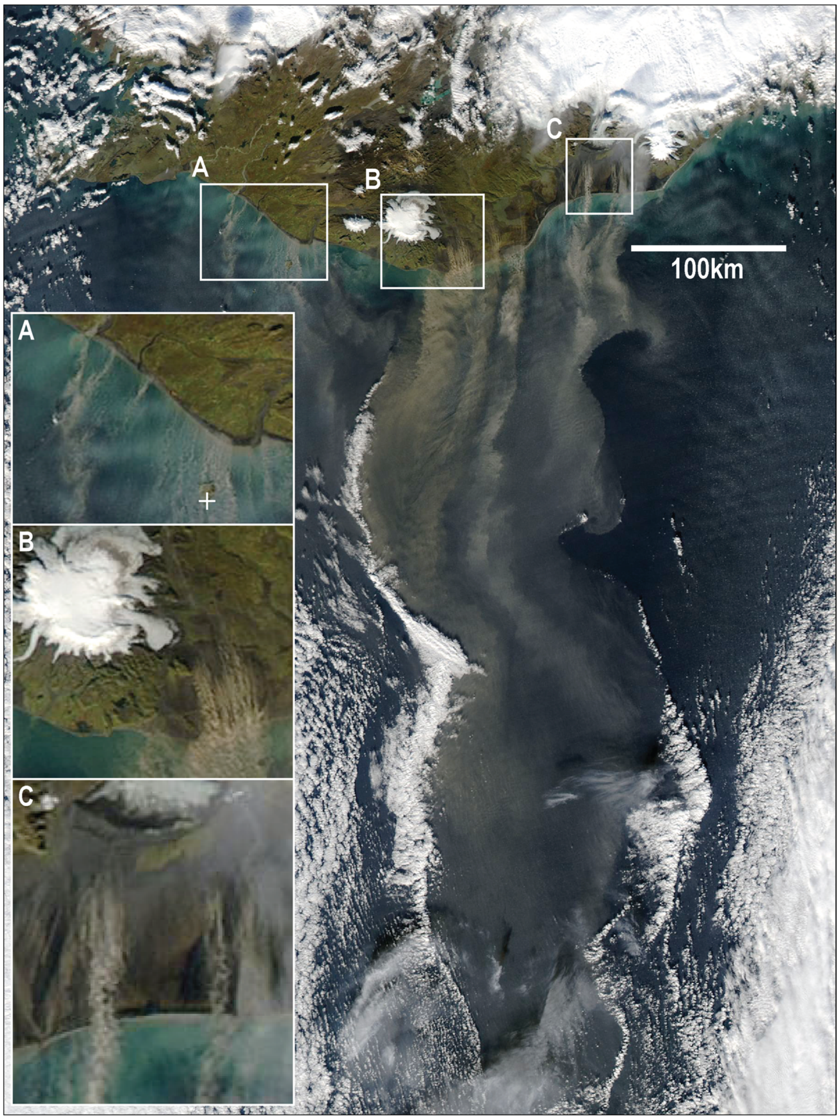

Figure 5. MODIS image 5 October 2004 showing dust plumes from glacigenic sources in southern Iceland. (a) Landeyjarsandur, cross indicates location of aerosol sampler at Stórhöfdi; (b) Mýrdalssandur; (c) Skeiðarársandur (Prospero *et al.*, 2012). Reproduced by permission of American Association for the Advancement of Science.

In contrast, in the Copper River area of Alaska, dust events predominantly occur in Autumn (Crusius *et al.*, 2011). Here, snow cover persists until early summer and the coincidence of maximum sediment exposure resulting from low river flows and strong winds does not take place until October/November (Figure 6). Figure 7 shows a remarkable series of recent plumes originating from the Copper River; Crusius *et al.* (2011) estimated that the plumes observed in November 2006 contained potentially 30–80 ktons of dust. Particle-size data and field observations around the Penny Ice Cap, Baffin Island suggest dust deposition from ice-proximal sources occurs in this region during late summer and autumn when snow cover is at a minimum (Zdanowicz *et al.*, 1998).

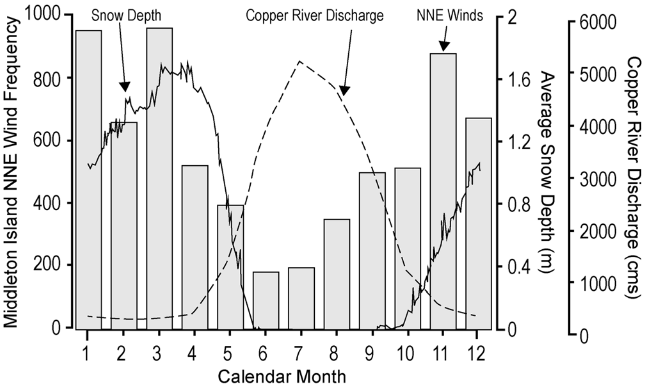

Figure 6. Monthly variation in frequency of down valley (NNE) winds, snow depth and river discharge in the Copper River catchment, Alaska (Crusius *et al.*, 2011). Glacial flour dust storms in the Gulf of Alaska: hydrological and meteorological controls and their importance as a source of bioavailable iron. Geophysical Research Letters, 38, L06602. Copyright 2011 American Geophysical Union. Reproduced/modified by permission of American Geophysical Union.

The interplay of glacial, fluvial and aeolian processes in the Kangerlussaq (Søndre Strømfjord) area of southwest Greenland was noted by Nordenskjöld (1910, 1914) and although a number of studies of aeolian activity have since been conducted in the region (Dijkmans and Tørnqvist, 1991; Eisner *et al.*, 1995; Bullard and Austin, 2011) none has used comparable techniques and sampling intervals over the course of a whole year so it is difficult to get a clear idea of the seasonal patterns of dust emissions. The glacially-fed Watson River flows from the Greenland Ice Sheet to the head of the fjord and has high flows in summer that can cover the entire outwash plain providing an input of fresh sediment; water levels decrease in the autumn and winter and low sediment loads indicate late-season exhaustion in sediment supply (Chu *et al.*, 2009). This, combined with a summer rainfall regime (most rain falls between June and September), and weak summer winds suggests aeolian activity will be lower in summer than in winter due to the limitations on sediment availability. In winter there is less precipitation and stronger winds, and sublimation is likely to make fine particles available at the surface. Snow cover is generally thin and discontinuous in the winter (< 20 cm), although thicker snow accumulation (>80 cm) can occur (Dijkmans and Tørnqvist, 1991). Observations of aeolian activity in the region suggest that sand-sized material may be reworked by wind all year round but dust storms are more prevalent during the winter months (Hobbs, 1931; Fristrup, 1953;

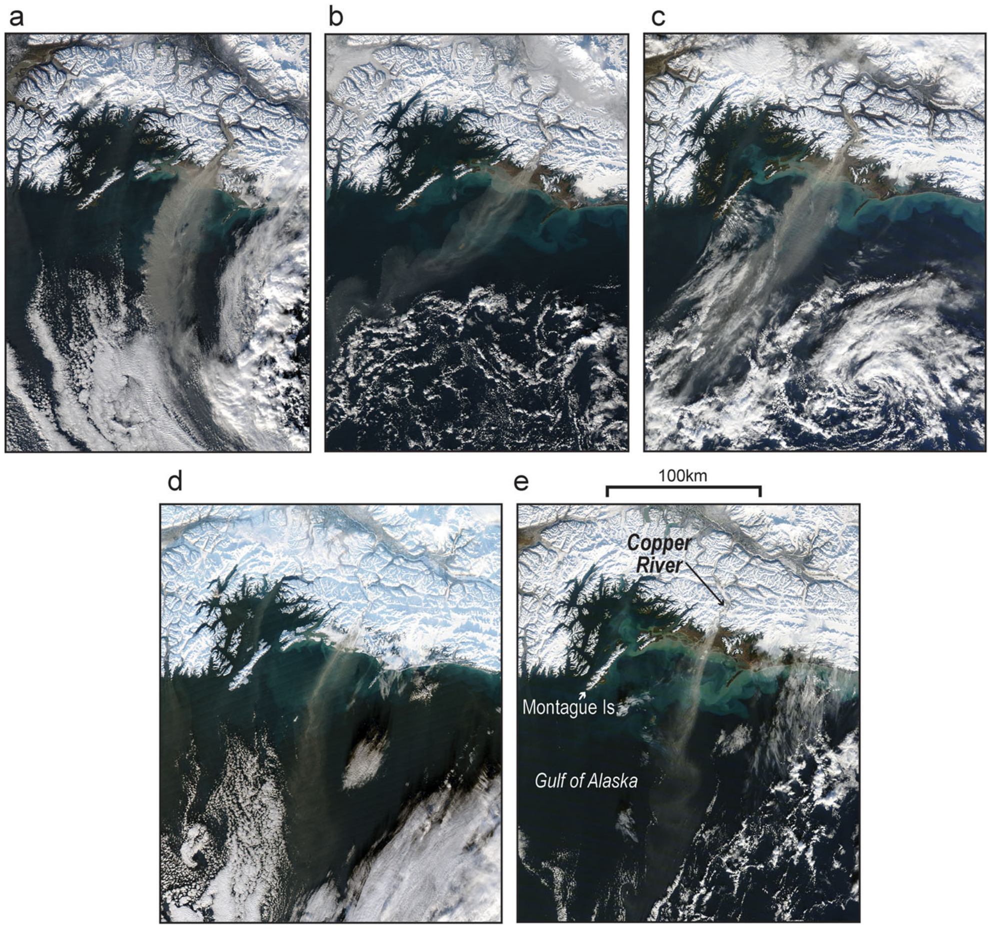

Figure 7. Dust plumes from the Copper River, Alaska. (a) 5 November 2005, (b) 6 November 2006, (c) 30 October 2009, (d) 23 December 2010, (e) 2 November 2011.

Engels, 2003). Engels (2003) measured summer dust deposition of 2–8 g m^{-2} a^{-1} compared with winter values of 1.97–4.33 g m^{-2} a^{-1}. Snow cover was high during Engel's (2003) study and the lack of a clear seasonal dust flux pattern in this area has been attributed to the high interannual variability of snow cover affecting sediment availability (Willemse, 2000). A similar cycle of year round aeolian reworking of coarse aeolian material but substantial sand and dust storms in winter associated with low meltwater flow, strong winds and sublimation has also been reported by McKenna Neuman and Gilbert (1986) and Gilbert (1983) in Arctic Canada.

By comparison with subtropical regions, there have been few studies of high latitude dust emissions, but those that have been made suggest the quantity of material entrained is often comparable, or even higher than in lower latitudes. For example, Nickling (1978) measured rates of 4.5–52 g m-w^{-1} s^{-1} for individual dust events in Yukon Territory, Canada, while in Iceland aeolian transport rates of 56 g m-w^{-1} s^{-1} for moderate dust storms have been recorded (Arnalds *et al.*, 2001) – values from lower latitudes typically range from <1 g m-w^{-1} s^{-1} to 30 g m-w^{-1} s^{-1}. Over longer periods, Church (1972) estimated that 1600 g m-w^{-1} of sediment was transported across the Lewis River sandur, Baffin Island, in July 1965 of which 70% occurred in just 3 days. Bullard and Austin (2011) recorded an average of 700–7100 g m-w^{-1} day^{-1} over a 7-day period near Kangerlussuaq, Greenland. However, for many of these studies the measurements include coarser sand-sized particles that are likely to be retained locally as well as the finer silts and clays that may be transported further afield. Arnalds (2010) suggested that dust emission rates in parts of Iceland, which have been calculated in the range 1000–5000 g m^{-2} (Kjaran *et al.*, 2006), were comparable with those in the Bodélé Depression, Chad ('the dustiest place on Earth'; Warren *et al.*, 2007) but acknowledged that the dust source area in Chad is considerably larger than that in Iceland.

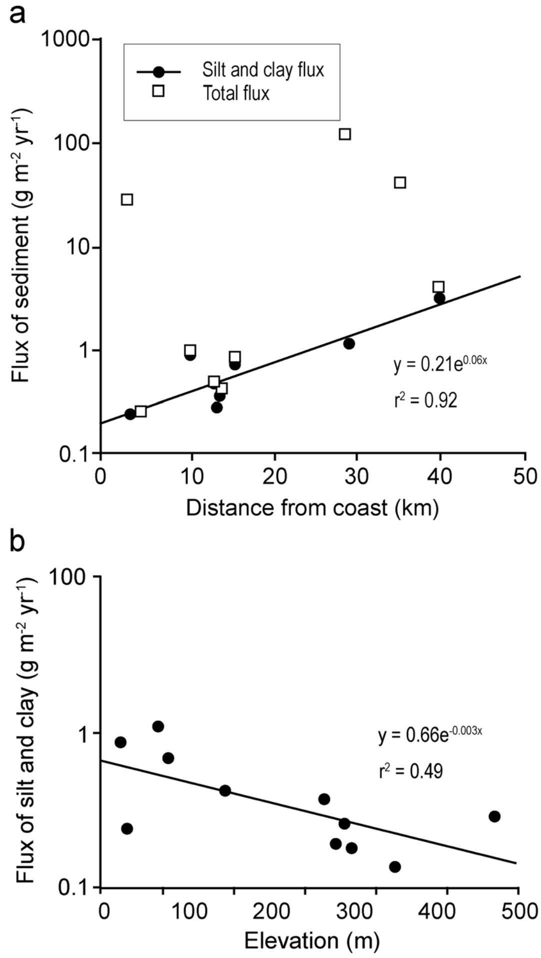

Figure 8. Dust flux data from Antarctica (Lancaster, 2002). (a) Changes in sediment flux with distance up the Taylor Valley. (b) Changes in silt-clay flux with elevation for various sites in the McMurdo Dry Valleys.

Dust Deposition

Comparative rates of dust deposition

Dust deposition occurs worldwide and overall deposition rates are higher close to source with both the quantity and particle-size of sediments decreasing with transport distance; the deposited material can make a major contribution to soil development in high latitudes (Muhs *et al.*, 2004; Anderson, 2007; Arnalds, 2010). Topography can also play an important role in determining deposition patterns with higher rates of deposition (and typically coarser sediments) in the source valleys and lower rates (and finer sediments) at higher elevations (Lancaster, 2002; Bullard and Austin, 2011) (Figure 8). Lawrence and Neff (2009) found dust deposition rates in 52 areas ranged from 0.05–450 g m^{-2} yr^{-1}. Comparing dust deposition rates from different studies is not straightforward because they are measured in different ways and over varying time periods. Assumptions have to be made as to the likely consistency of deposition through time, and most long records suggest dust deposition is highly variable both temporally (seasonally and inter-annually) and spatially (Bory *et al.*, 2002; Ayling and McGowan, 2006). Of the studies examined by Lawrence and Neff (2009), 52% were based on records lasting 2 years or less. Table I shows a selection of dust deposition rates from glacierised and non-glacierised regions and suggests that rates in the former exceed the highest included in Lawrence and Neff's (2009) survey. In his review of dust sources and deposition patterns in Iceland, Arnalds (2010) points out that dust deposition rates in central and southern Iceland are higher than any in Lawrence and Neff's (2009) list with even those Icelandic areas classified as 'low deposition rate' (13–26 g m^{-2} yr^{-1}) ranking alongside the top 20 highest worldwide. It is worth noting, however, that the main Icelandic zone of high dust deposition coincides not only with glacially-active regions but also with the active volcanic zone; consequently some of the source areas are likely to be primary volcanic deposits rather than glacigenic sediments (cf. Thorarinsdottir and Arnalds, 2012). Even more extreme are the dust deposition values recorded by Hugenholtz and Wolfe (2010) in the Athabasca River Valley, Canada, which reach over 33,000 g m^{-2} yr^{-1}. At this site, the glacially-fed Athabasca River causes seasonal filling and deposition of fine sediments in Jasper Lake. During low water levels the sediments are deflated and result in high local dust deposition rates. The extreme values are due to sediment trapping in nearby coniferous forest and proximity to source (<100 m).

Primarily these high dust deposition rates in glacierised regions reflect the fact that most measurements are taken close to source and would be classified as 'local' (<10 km from source) or 'regional' (10–1000 km), however in some cases dust deposition reflects a combination of sources. For example Lawrence and Neff (2009) classify dust deposition on the Penny Ice Cap (0.05 g m^{-2} yr^{-1}) as from a global source (probably continental North America or central Asia; Zdanowicz *et al.*, 1998), however, it is likely that both local/regional and global dust are deposited on the ice (Zdanowicz *et al.*, 2000). Similarly, snow pit samples on Berkner Island ice cap, Antarctica include dust deposits from a variety of sources including not only Patagonia and Australia but also local sources in Antarctica such as the Dry Valleys and other ice-free areas (Bory *et al.*, 2010). Dust deposited on the central Greenland Ice Sheet has been attributed to Asian sources (Tegen and Rind, 2000), but dust from Icelandic sources can also reach

Table I. Dust deposition in selected locations

Location		Dust deposition $g\,m^{-2}\,yr^{-1}$	Reference
Iceland	Very deposition areas	569-807	Arnalds, 2010
	High deposition areas	256-391	Arnalds, 2010
	Medium	54-117	Arnalds, 2010
	Low	13-26	Arnalds, 2010
Greenland	Lake Anna catchment, Kangerlussuaq	2-8	Engels, 2003
Canada	Athabasca River Valley	24-33120	Hugenholtz and Wolfe, 2010
	British Columbia	13.1	Owens and Slaymaker, 1997
	Penny Ice Cap	0.05	Zdanowicz *et al.*, 1998
Antarctica	McMurdo Dry Valleys	0.24-2.96	Lancaster, 2002
		6.48*	Ayling and McGowan, 2006
New	Lake Tekapo	257	McGowan *et al.*, 1996
Zealand	West Coast	26.4	Marx and McGowan, 2005
China	Desert regions	450	Zhang *et al.*, 1998
Libya	Sahara	82-276	O'Hara *et al.*, 2006
Israel	Negev Desert	17.7-205.6	Littmann and Gintz, 2000, Offer and Goossens, 2001
Australia	NSW	37.6	McTainsh and Lynch, 1996
USA	Nevada and California	11.7	Reheis and Kihl, 1995
Local	0-10 km from source (n = 11)	50-500	Lawrence and Neff, 2009
Regional	10-1000 km from source (n = 28)	1-50	Lawrence and Neff, 2009
Global	>1000 km from source (n = 13)	0-1	Lawrence and Neff, 2009

*Extrapolated from summer values.

this location (Drab *et al.*, 2002; Prospero *et al.*, 2012). In the ablation zone around the western and northern margins of the Greenland Ice Sheet supraglacial sediments comprise a mixture of fine sediments exposed at the surface by ablation and direct dust deposits. On the western margin, Wientjes *et al.* (2011) excluded an Asian source for this latter material and suggested it was transported on to the ice by deflation from the surrounding local tundra and ice-free regions. Similarly, Bøggild *et al.* (2010) concluded that the relatively coarse nature of the surface dust deposits (27% silt, 7% fine sand) indicated a local source of material blown up on to the ice. Atkins and Dunbar (2009) describe the deflation of material from 'dirty ice' (formed when sediments incorporated into the ice by adfreezing reach the surface by ablation) in Antarctica and suggest an input of 7.8–24.5 $g\,m^{-2}\,yr^{-1}$ to the ocean floor from aeolian transported material of which 30–40% is <25 µm in diameter.

It is difficult to determine what proportion of glacigenic dust is transported and deposited a long distance from source (>10 km), in part due to a lack of field data and also due to the difficulty of tracking dust plumes from regions with substantial cloud cover (Gassó and Stein, 2007; Gassó *et al.*, 2010; Thorsteinsson *et al.*, 2011; Prospero *et al.*, 2012). Many dust events affect only the local area as sediments are relatively coarse and katabatic winds can be very localised. The depositional record (loess) in many areas shows a rapid decline in sand and coarse silt content and an increase in the proportion of fine particles downwind of fluvioglacial sources as coarser wind-blown materials settle out of the airflow (Muhs *et al.*, 2004; Dijkmans and Tørnqvist, 1991: Figure 9) but how much material is transported higher and further in the atmosphere is largely unknown. Dust emissions from the Icelandic sandur plains can affect air quality in Reykjavik over 100 km away (Thorsteinsson *et al.*, 2011) and some dust plumes from glacigenic sediments are very extensive. For example, the dust plumes examined by Crusius *et al.* (2011) all extended over 350 km across the Gulf of Alaska while the dust plume in Figure 5 extends more than 500 km over the Atlantic Ocean. Icelandic dust has been recorded over 1000 km from source in central Greenland (Drab *et al.*, 2002) and Ireland (Ovadnevaite *et al.*, 2009). From the glaciated and volcanic Jan Mayan Island in the Arctic Ocean, King (1939) wrote 'storms of volcanic sand are common on the eastern shores of

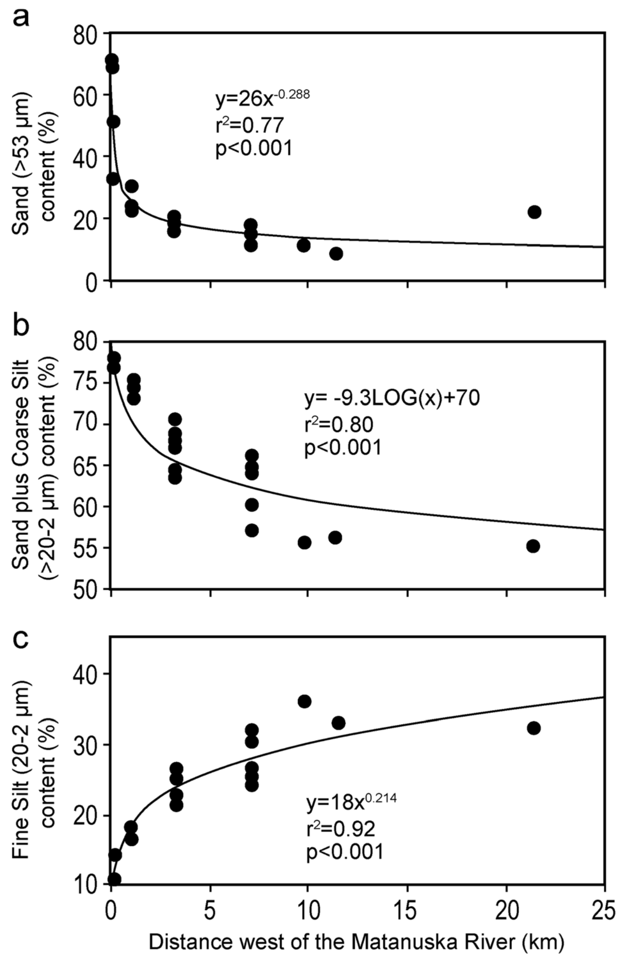

Figure 9. Percentage changes in sediment content of unaltered loess with distance from the Matanuska River, Alaska: (a) sand >53 µm; (b) sand and coarse silt >20 µm; (c) fine silt 2–20 µm (Muhs *et al.*, 2004). Reproduced by permission of Elsevier.

Jan Mayan, ships several miles out to sea occasionally collecting inches of it on their decks. Volcanic sand becomes an inevitable constituent of each meal eaten on the island!'. Dust deposition from such plumes will contribute to marine sediments as well as to atmospheric and potentially terrestrial processes.

Impacts of glacigenic dust deposition

The impacts of dust deposition from tropical and mid-latitude sources have been widely discussed in the literature and many also occur as a result of glacigenic dust deposition. These include contributions to soil development, both in terms of texture and soil biogeochemistry, landform development and vegetation, through the supply of nutrients. The focus in this section is on glacigenic dust impacts in the marine system and the cryosphere; impacts on soil development are discussed later in the paper.

The potential impact of iron-bearing dusts on marine ecosystems is the subject of considerable study and debate (Maher *et al.*, 2010; Bouttes *et al.*, 2011). The 'iron hypothesis' (Martin, 1990) is that dust input to the oceans decreases the effects of iron limitation on primary productivity; resultant CO_2 uptake and export of biogenic carbon from surface to deep ocean waters could explain part of the glacial–interglacial difference in atmospheric CO_2 levels. This effect is likely to be most important in the High Nitrate Low Chlorophyll (HNLC) oceans, which include the northwest subarctic Pacific Ocean and the Southern Ocean (Harrison *et al.*, 1999). Although large dust plumes have been observed over these oceans there are very few measurements of dust concentration and iron content (Mahowald *et al.*, 2009). Crusius *et al.* (2011) estimated that a single dust event in 2006 transported 25–80 ktons of dust and deposited 30–200 tons of soluble iron in the surface waters of the Gulf of Alaska (Figure 7b). Glacial suspended sediments have been found to contain significantly higher amounts of iron than some desert dust sources (Schroth *et al.*, 2009) and any changes in the quantity of dust entrained from glacial sources may therefore also impact marine ecosystems (Schroth *et al.*, 2011).

Dust deposition also affects the cryosphere; typically light dust deposition ($<0.2\,kg\,m^{-2}$) on snow or ice decreases albedo, increasing melt rates and runoff (Peltier and Marshall, 1995; Adhikary *et al.*, 2002; Bøggild *et al.*, 2010) whereas more dense dust deposition can insulate the snow/ice from solar radiation slowing the rate of ablation. Adhikary *et al.* (2000) conducted experiments on the relationship between 'dust' (actually 150–350 μm fine–medium sands) and ablation rates at Lirung Glacier in the Nepal Himalayas and suggested that a layer of dust ≤ 1.33 mm thick would enhance ablation rates while a thicker layer would insulate the surface; the critical threshold dust concentration was $\approx 600\,g\,m^{-2}$ (Figure 10). Singh *et al.* (2000) found a 2 mm layer of dust caused an increase in the degree day factor (used to indicate melt rate) of $0.77\,mm°C^{-1}\,day^{-1}$ and $0.6\,mm°C^{-1}\,day^{-1}$ for deposits on snow and ice respectively. A number of studies have indicated that dust deposits in the cryosphere can increase not only the quantity, but also the timing of runoff (Painter *et al.*, 2007; Oerlemans *et al.*, 2009) although the relative timing of dust versus snow deposition can significantly impact the effect of the dust layer (Ledley and Pfirman, 1997). It is likely that differences in dust source characteristics (such as mineralogy, colour, shape), as well as the thickness of deposit and particle-size distribution after transport will cause considerable variation in the impacts of dust on melt rates, as will redistribution of the sediments on the ice surface by supraglacial runoff.

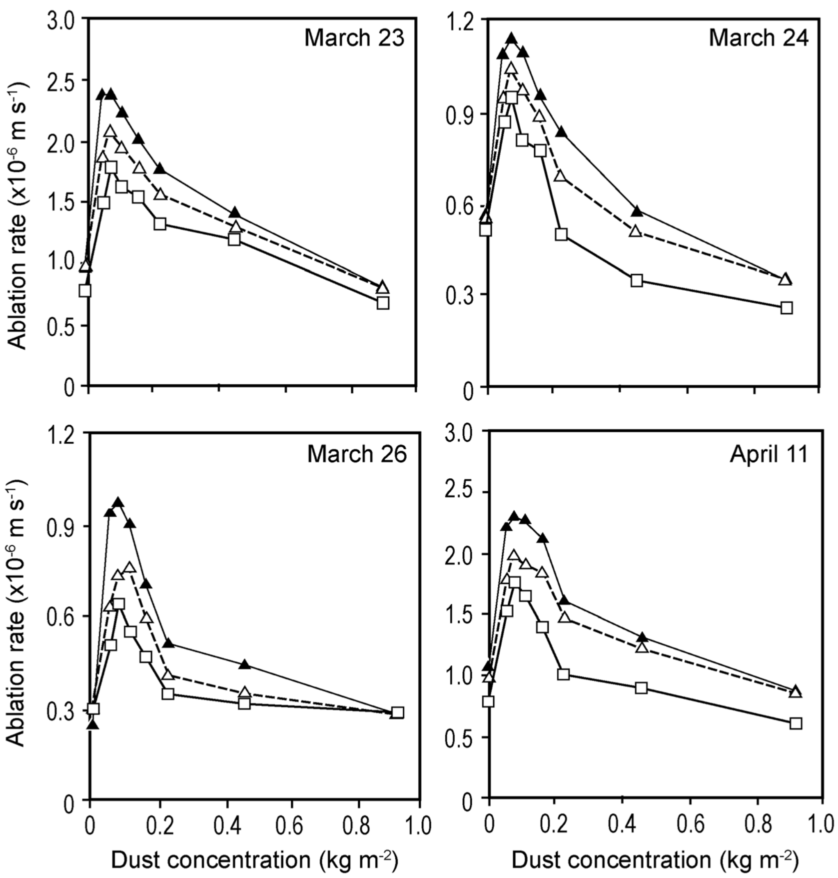

Figure 10. Daily average snow ablation rate ($m\,s^{-1}$ water equivalent) versus dust concentration. Open squares, open trianges and filled triangles indicate observed, estimated with aggregation of dust particles, estimated without aggregation of dust particles respectively (Adhikary *et al.*, 2002). Reproduced by permission of John Wiley and Sons.

Glacigenic Dust Emissions and Glacier Retreat

Under contemporary conditions many ice sheets and glaciers are retreating and predicted to continue retreating in the foreseeable future in response to climate warming. Taking the three main themes of sediment supply, availability and transport capacity, each is expected to change in the future with potential impact on glacigenic dust emissions. In the short-term (decades) sediment supply is likely to increase due to terrestrial ice retreat in response to rising global temperatures and associated predicted increases in suspended sediment loads. However, this needs to be considered against potential decrease in sediment availability due to vegetation colonization of outwash plains and weakened katabatic winds due to lower ice profile gradients.

Changes in sediment supply to floodplains

Glacigenic dust emissions have been linked to suspended sediment supply to the floodplain by glacier runoff. Several models of glacier and ice sheet retreat suggest that this will be associated with an initial increase in runoff over the next few decades followed by stabilisation, then a decrease to runoff levels lower than at present (Björnsson and Pálsson, 2008; Huss *et al.*, 2008; Jóhannesson *et al.*, 2012). The period of enhanced runoff is likely to vary depending on the size and other characteristics of the ice mass. For example Flowers *et al.* (2005) predicted that Vatnajökull, the largest ice cap in Iceland, will be 12–15% smaller in terms of area within 100 years and that the maximum increase in glacier-derived runoff will be approximately 25% after 130 years. The reduction in ice cap size and increased runoff may result in a changed subglacial drainage pattern as drainage divides migrate, and this may have an impact on meltwater and jökulhlaup routing and the location of proglacial lakes (Flowers *et al.*, 2005). Dust emission areas may also change depending on which sectors of the outwash plains remain active. On large ice caps and ice sheets individual outlet glaciers may undergo retreat and associated increase in runoff on shorter timescales (Aðalgeirsdóttir *et al.*, 2011).

Meltwater sediment concentration is not necessarily correlated with discharge so a predicted increase in runoff may not be associated with an increase in sediment supply to the outwash plains. Some studies of enhanced glacier ablation and past glacier retreat have suggested that increased suspended sediment transport can occur (Leonard, 1997; Menounos and Clague, 2008; Koppes *et al.*, 2010: Figure 11). For example, Stott and Mount (2007) found suspended sediment load from the Torrent du Glacier Noir, France was 3–4 times higher during an unusually warm ablation period in 2003 compared with the 2°C cooler ablation period in 2004; Hodgkins *et al.* (2003) documented net erosion of a proglacial plain in Svalbard during a warm, high runoff year and net aggradation in a cold, low runoff year. Yet others have found no relationship between advance and retreat patterns and changes in meltwater sediment yield. For example, Hallet *et al.* (1996) examined suspended sediment yield over a period of 25 years in Norway during which the glacier retreated for the first 9 years, maintained a stable position for 10 years and then advanced for the remainder of the time. Glacier advance or retreat had very little influence on suspended sediment yields. Similarly, Schiefer *et al.* (2010) found that suspended sediment concentrations in a glacierised catchment in British Columbia, Canada were extremely variable and hard to predict even with a long available record because events such as short-term rapid glacier recession, a landslide and heavy rainfall affected sediment supply. Sustained rates of increased suspended sediment will depend on factors such as catchment characteristics and subglacial sediment storage but are hard to predict especially where subglacial reorganisation of drainage patterns occurs and may cut off sediment supply, or mobilise new stores. High dust emissions in some proglacial regions have been associated with catastrophic outburst floods; Evans and Clague (1994) suggest that in a warmer climate there may be an increase in the frequency of jökulhlaups due to increases in air temperature, melt rates and runoff, but that the magnitude of the floods may decrease due to changes in ice dynamics. Predicted increase in volcanic activity associated with glacier retreat (Hall, 1982; Pagli and Sigmundsson, 2008; Carrivick *et al.*, 2009) may also increase jökulhlaup frequency where this is a trigger. The range of short-term impacts influencing seasonal and inter-annual variations in suspended sediment load means that sediment fluxes and glacier 'extent' are probably only correlated at the longest timescales. Over such longer timescales, there may be links between regional scale atmospheric circulation characteristics (such as phases of the North Atlantic Oscillation or Arctic Oscillation), runoff and suspended sediment loads. For example the Lawler *et al.* (2003) 20-year study of variability of river flow and sediment fluxes in three catchments in southern Iceland found that suspended sediment loads decreased from 1973–1992 in response to reduced discharge in spring and autumn, in turn caused by cooling linked to a positive North Atlantic oscillation index.

While long-term trends of suspended sediment delivery to the floodplain associated with glacier retreat may be hard to predict, one widely observed phenomenon could have a substantial impact on the supply of sediments to the aeolian system – proglacial lake formation. Proglacial lakes trap meltwater sediment and prevent it being transferred to other parts of the geomorphological system, they may also cause river incision in glacial outwash plains downstream which reduces the possibility of replenishing floodplains with fine sediment, decreasing the likelihood of dust emissions (Sugden *et al.*, 2009: Figure 12). Glacierised catchments in Patagonia are thought to have been significant sources of dust in the southern hemisphere during glacial periods and there are substantial deposits of glacio-fluvial sediment in the region (Clapperton, 1993), however contemporary suspended sediment loads are relatively low (Pepin *et al.*, 2010) in part due to the formation of proglacial lakes which trap sediments preventing them from reaching downstream floodplains (Gaiero *et al.*, 2003). As a result, under contemporary conditions dust derived from volcanic sources may be more important than glacial sources in southern South America (Gaiero *et al.*, 2004). The occurrence and size of proglacial lakes is increasing with glacier retreat in parts of Iceland, Svalbard (Schomacker and Kjaer, 2008), the Himalayas (Komori, 2008) and Canada (Schiefer and Gilbert, 2008). Schiefer and Gilbert (2008)

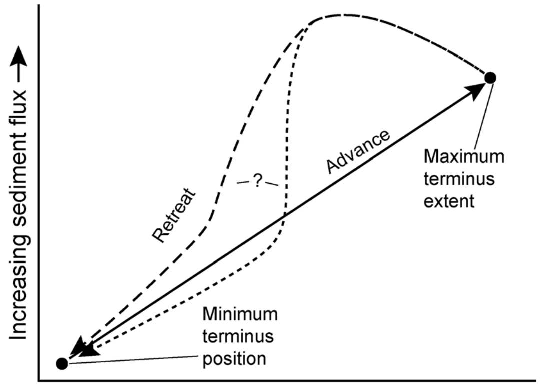

Figure 11. Conceptual model for changes in sediment flux from a glacier during a cycle of glacier advance and retreat (Jansson *et al.*, 2005). As the glacier retreats fresh sediment is exposed in the forefield increasing sediment flux. Reproduced by permission of John Wiley & Sons.

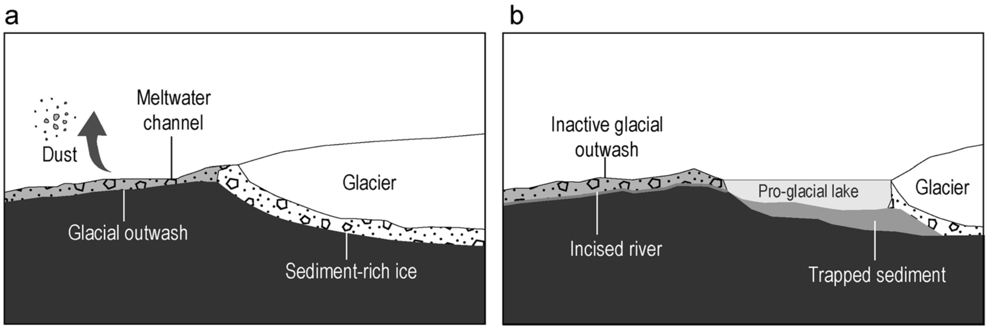

Figure 12. Impact of proglacial lake formation on dust emissions based on Sugden *et al.* (2009). (a) No proglacial lake, meltwater supplies sediment to the floodplain which dries and is entrained by the wind. (b) Proglacial lake traps sediment and outwash plain becomes inactive (Ackert, 2009). Reproduced by permission of Nature Publishing Group.

describe the formation of Silt Lake, a proglacial lake which formed after 1947 when the Lillooet Glacier (Canada) retreated and by 1973 had an area of 0.75 km^2. When sediment budgets in the lake were monitored in 2005, by which time it had shrunk to 0.49 km^2 due to sandur progradation, it was found that during peak glacial melt the lake trapped approximately 70% of the suspended sediment preventing it from reaching the outwash plains downstream. In Iceland, the ice-marginal lakes of Vatnajökull have been expanding since 1995 as many glaciers change from having land-based termini to being lake-terminating glaciers. (Schomacker, 2010). Jökulsárlon, the proglacial lake at Breiðamerkurjökull, expanded from 15 km^2 in 2000 to >21 km^2 in 2009 and is thought to trap virtually all the meltwater sediments; this trapping has been associated with downstream river incision and coastline retreat (Jóhannesson and Sigurđarson, 2005). Large quantities of sediment may build up in proglacial lakes, but if further glacier retreat (or an event such as a jökulhlaup) leads to restructuring of proglacial meltwater systems this can lead to drainage of the proglacial lake and exposure of large areas of fine sediment to deflation.

Postglacial soil development and vegetation recovery

It is unclear how sediment supply to the aeolian system by meltwater may change in future and there are also likely to be changes in sediment availability. Receding glaciers will expose extensive new areas of loose sediments including large quantities of fine silts and sands increasing the potential surface area of dust emissions (Gisladóttir *et al.*, 2005). Flowers *et al.* (2005) simulated the response of Vatnajökull to warmer temperatures and suggested that at a rate of 2°C of warming per century, the ice cap would lose 25% of its area in 130–150 years. This would potentially expose an additional 2000 km^2 of sediments to the wind although this may be a short-lived sediment source given that, unless replenished or reworked by meltwater, deflation will lead to the development of a protective lag deposit. In addition, as ice retreats and temperatures warm, vegetation would be expected to recolonise previously glacierised areas and soil development will occur, increasing surface roughness and protecting sediments from deflation.

Matthews (1992) provides a comprehensive review of the ecology of recently deglaciated terrain, highlighting the spatial and temporal variability and complexity of interactions among factors controlling soil development and vegetation succession in the wake of ice retreat. Both are slow processes; soil development has been observed to occur as rapidly as 50 (A-horizon) to 100 (B-horizon) years (Ugolini, 1968; He and Tang, 2008; Dümig *et al.*, 2011) but often takes considerably longer. Studies from the Jotunheimen-Jostedalsbreen region of southern Norway suggest that although soil horizons may be identifiable after a few hundred years, full development of different horizons takes considerably longer (Matthews, 1992). Rates of soil development are controlled by environmental factors including the nature of the parent material, topography, climate and vegetation (Matthews, 1992). Fine-textured but poorly-sorted deposits develop soils more rapidly than coarser, well-sorted materials, largely due to the higher moisture holding capacity of the former, which tends to favour more rapid soil development on unsorted tills in moraine slopes than on outwash plains. Moraine crests typically have poor soil development due to a combination of snow cover in winter, moisture deficit in summer and exposure to strong winds which can cause aeolian erosion of the silt fraction (Hall and Shroba, 1993; Applegarth and Dahms, 2004). The input of glacigenic dust to soils, in the form of loess layers, can also influence soil development, typically slowing or interrupting the process, as has been observed in New Zealand (Rodbell, 1990) and Alaska (Muhs *et al.*, 2004). Climate affects soil development primarily through the combined impact of moisture and temperature. In Alaska, soil development at Glacier Bay, which has a maritime climate (mean annual temperature *c.* 5°C, annual precipitation *c.* 1900 mm) is occurring more rapidly (recognizable podzols after *c.* 150 years) than on the Robson Glacier forefield, British Columbia which has a cooler, drier continental climate (mean annual temperature *c.* 3°C, annual precipitation *c.* 400–500 mm) and where surfaces with the same exposure date as those in Alaska have considerably less mature soils (Tisdale *et al.*, 1966; Sondheim and Standish, 1983). Soil development in recently deglaciated areas is predicted to become more rapid under warmer temperatures (Cannone *et al.*, 2008).

Soil development and recolonisation of vegetation are closely-coupled (Matthews, 1992), and like soil development, vegetation can be slow to establish on glacier forefields. On the active floodplain rate of recolonisation of vegetation will reflect the magnitude and frequency of reworking; topography and microclimate also combine to produce complex spatial patterns of vegetation succession (Tisdale *et al.*, 1966; Cannone *et al.*, 2008; Garibotti *et al.*, 2011). Cryptogams and biological crusts are the first, and sometimes only, colonisers of deglaciated terrain for the first few decades, and even after 50 years percentage cover is unlikely to exceed *c.* 50% (Matthews, 1992). Cannone *et al.* (2008) determined that surfaces deglaciated 6–11 years ago had, on average, a cover of only 2–3% of mosses and vascular plants although locally this could be up to 35%. Tisdale *et al.* (1966) found the moraines in front of the Robson Glacier, British Columbia had developed a moss and lichen cover of <4% after 50 years, increasing to 34% after 160 years. Cryptogams can be vulnerable to both wind and water erosion and as consequence may provide little protection against surface reworking during floods or strong winds. Herbs, dwarf-shrubs and trees typically follow in the succession. The rate of recolonisation may also be affected by aeolian processes. Enhanced aeolian sand transport around glaciers in Iceland has been observed to bury and, in some cases, destroy vegetation cover (Alho, 2003; Gisladóttir *et al*., 2005) and Smith and Dugmore (2006) have suggested that deflation lag surfaces may be resistant to rapid colonisation by vegetation but these can be reworked by increased runoff (increasing potential aeolian sediment supply). Thawing permafrost and warmer temperatures will also promote the desiccation of surface deposits through increased infiltration and enhanced evaporation.

Retreating glaciers and wind regime

The relative importance of different drivers of aeolian transport such as synoptic versus katabatic winds affect how much dust is entrained and how far it is transported. In Antarctica, climate models indicate circumpolar westerlies are likely to decrease in strength as the ozone hole recovers (estimated as 2065), however during some seasons wind strengths are likely to increase, affected by climate warming. Perhaps more importantly katabatic winds, which field studies suggest are crucial for aeolian entrainment in some regions, are rarely incorporated into global models due to the wind field scale, although they have been included with some success in regional climate models (Jourdain and Gallée, 2011). The diurnal and seasonal strengths of katabatic winds are predicted to change in association with changing ice-slope, proglacial vegetation and temperature-melting gradients (van den Broeke *et al*., 1994; van de Wal *et al*., 2005). These changes are very dependent on catchment and retreat characteristics and some glacier slopes steepen as they retreat due to thinning in the ablation zone whereas others become less steep (Willis *et al*., 2012) with implications for wind strengths and potential dust entrainment (Zdanowicz *et al*., 2000). Under future climate scenarios, a reduction in the intensity of katabatic winds and possible increase in the magnitude and frequency of ice-wards (synoptic) airflow combined with higher dust loadings from increased meltwater deposition may trigger rapid melting of ice in a positive feedback system (Peltier and Marshall, 1995). Oerlemans *et al*. (2009) documented a decrease in glacier albedo due to dust deposition from local sources that increased melt rates causing the removal of about 3.5 m of ice (equivalent to a temperature rise of 1.7 K) in 4 years. The dust comprised both mineral and biogenic components with the mineral dust stimulating the growth of algae.

The changing relative importance of glacigenic dust emissions

Quaternary records indicate that glacigenic dust emissions were considerably higher in the past when ice cover was more extensive than at present. During the Last Glacial Maximum (LGM) ice covered around 25% of Earth's land area (*c*. 37 million km^2). Potential dust source areas – a function of aridity, reduced vegetation cover and sediment supply – were more extensive at the LGM in both the northern and southern hemispheres. In the northern hemisphere, potential dust source areas extended across northern Alaska and northern Canada, were considerably more extensive in northeast Asia than at present, and were generally spatially associated with maximum ice limits (Mahowald *et al*., 1999). In the southern hemisphere potential dust source areas were more extensive in South America, particularly in the Andean and Patagonian regions, and in mainland Australia and southern Africa, although the latter were associated with aridity rather than glacigenic sources (Hall, 2004). Without explicitly taking in to account glacigenic sources, Mahowald *et al*. (2006) suggest the total global dust source area at the LGM was up to 35% larger than at present (based on vegetation cover and CO_2 fertilization effects) with total dust loadings up to 60 Tg m^2. When selected glacigenic dust sources in Europe, Siberia, North America and the Pampas region of South America are factored in, modelled global dust loadings for the LGM are predicted to have been 70–80 Tg m^2, with dust source strengths 4–5 times higher in the northern hemisphere than in the southern hemisphere (Mahowald *et al*., 2006). Under contemporary conditions only *c*. 10% of Earth's land masses are ice-covered but there are marked differences between the hemispheres. Around 2.3% of northern hemisphere land is ice-covered compared with 25.9% of land in the southern hemisphere – figures which reflect both the smaller area of land available in the southern hemisphere (less than half that in the north) and the expanse of the Antarctic ice sheet. Northern hemisphere dust emissions remain higher than those in the southern hemisphere and are dominated by north African sources; estimated total current global dust loadings are around 30 Tg m^2 (Mahowald *et al*., 2006).

During the LGM, meltwater suspended sediment flux from small ice masses is estimated to have been in the range 3000–6000 Tg yr^{-1}, three times higher than that from the Antarctic and Greenland ice sheets combined (Raiswell *et al*., 2006). At present, small ice masses are still estimated to contribute the highest levels of meltwater suspended sediment flux – 1100 Tg yr^{-1} compared with 300 Tg yr^{-1} from the two ice sheets (Raiswell *et al*., 2006). However, these small ice masses represent a small fraction of the current global ice cover (<2%) so in the future the biggest potential change in glacigenic contributions to the dust cycle will be associated with the retreat of the two main ice sheets. The extent of ice cover in Antarctica at the LGM has not yet been fully determined, however, ice extended off the present shore around most of the continent with only very limited ice-free coastal areas (Ingólfsson, 2004). Any meltwater sediment supply from the Antarctic ice sheet is therefore likely to have been deposited directly in to the marine system. For these sediments to become available to the dust cycle, they need to be deposited on land where desiccation and deflation can occur. At present 98% of the Antarctic continent is ice covered and most sediment flux still directly enters the marine system; a greater proportion of outlet glaciers from the Greenland ice sheet are land-terminating. With continued retreat, it is likely that both ice sheets will discharge more meltwater suspended sediments in to fluvial systems and floodplains rather than the oceans making large-scale dust emission from the proglacial zones in Greenland and Antarctica possible.

It is difficult to predict what the magnitude of these dust emissions might be, although they are unlikely to exceed those at the LGM. Raiswell *et al*. (2006) estimated total contemporary glacial meltwater suspended sediment flux to be 1400 Mt yr^{-1} of which 300 Mt yr^{-1} was attributed to the Antarctic and Greenland ice sheets (and all of which was assumed to reach the ocean). If the two ice sheets were to contract in size and become land-terminating then this sediment could become available to the aeolian system – 300 Mt yr^{-1} is the equivalent of total contemporary dust emissions from Asia, or twice that of contemporary emissions from the Americas and Australia combined. In reality, not all of this sediment would directly enter the dust cycle as a substantial quantity would remain in the fluvial system and be transported to the oceans, however climate warming is also likely to enhance melt rates, runoff and possibly meltwater suspended sediment concentrations which could increase glacigenic sediment supply to the dust cycle at these high latitudes, and there is likely to be a considerable time lag before soil development and vegetation cover are adequate to prevent dust emissions. There are too many uncertainties to suggest whether the relative importance of glacigenic dust sources could exceed that of non-glacigenic sources globally in the next few hundred years. Within the southern hemisphere, dust emissions from the Antarctic continent could possibly exceed combined emissions from South America, Australia and southern Africa although the quantity of dust emissions from these sources may also change in response to both climate change and land use pressures. The Kalahari dunefield in southern Africa has been cited as a possible new southern hemisphere dust source (Bhattachan *et al.*, 2012) but this depends on reactivation of the dunefield which is a matter of some debate (Thomas *et al.*, 2005; Ashkenazy *et al.*, 2012). Predicted impacts of climate change in central Australia (CSIRO, 2007) suggest drier conditions which, in combination with more intense rainfall and associated flooding, are likely to increase dust emissions in areas that are currently supply-limited (Bullard *et al.*, 2008), but there are no estimates of the magnitude of any increase.

Conclusions

Contemporary glacigenic dust sources have been under-represented, or more commonly ignored, in many reviews of global dust emissions and aerosol impact (Goudie, 2009; Carslaw *et al.*, 2010; Ravi *et al.*, 2011; Shao *et al.*, 2011). There are several possible reasons for this apparent neglect. First, there have been no attempts to quantify systematically the global expanse of contemporary glacigenic, high latitude and cold climate dust sources. At present they cover a relatively small area of Earth's surface when compared with temperate and sub-tropical sources and so are generally assumed not to be significant in terms of total contemporary global dust emissions. However, where they have been studied, emission intensity and deposition rates are very high; it is unclear what the total contribution of glacigenic dust is, however, its impact is more likely to be determined by the timing of dust emissions and the feedback processes triggered following deposition rather than by absolute emission rates. Second, some examples of measurements of dust emissions and deposition in high latitudes have been cited here, however, many field-based studies to date have been restricted to the summer months. It is likely that in some high latitude regions dust activity will be higher during the fall–winter months and has therefore been under-estimated. Third, although space-based remote sensing has been demonstrated to be one of the best tools for dust detection, it has proved to be a poor performer in high latitude environments (Gassó and Stein, 2007; Gassó *et al.*, 2010; Crusius *et al.*, 2011; Thorsteinsson *et al.*, 2011) due to the excessive cloudiness and poor viewing conditions commonly found. By failing to, or under-detecting, dust events from satellite observations, it is very difficult to track their transport pathways, estimate sediment loading or constrain model estimates of deposited dust over the oceans.

Long-term global-scale modelling of changes in atmospheric dust loading under different climate conditions has shown that inclusion of inferred glacigenic dust sources improves the simulation for the Last Glacial Maximum (Mahowald *et al.*, 2006). Global dust models are important for understanding and predicting the influence of the dust cycle on global climate, however, there are many research questions to be addressed before the impacts and responses of glacigenic dust sources to present and future scenarios can be parameterised effectively in such models. These include understanding the relative importance of different drivers of aeolian transport such as synoptic versus katabatic winds but also the magnitude, frequency and seasonal timing of dust emissions across the high latitudes, which affect when and how much material enters the atmosphere. Although the total quantity of contemporary high latitude, glacigenic dust may be small compared with that from low latitude, warm arid regions, its potential impact when deposited, particularly on ice or in the oceans is substantial given the importance of feedbacks within the Earth system.

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