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AEOLIAN PROCESSES AND THE BIOSPHERE

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[1] Aeolian processes affect the biosphere in a wide variety of contexts, including landform evolution, biogeochemical cycles, regional climate, human health, and desertification. Collectively, research on aeolian processes and the biosphere is developing rapidly in many diverse and specialized areas, but integration of these recent advances is needed to better address management issues and to set future research priorities. Here we review recent literature on aeolian processes and their interactions with the biosphere, focusing on (1) geography of dust emissions, (2) impacts, interactions, and feedbacks, (3) drivers of dust emissions, and (4) methodological approaches. Geographically, dust emissions are highly spatially variable but also provide connectivity at global scales between sources and effects, with "hot spots" being of particular concern. Recent research reveals that aeolian processes have impacts, interactions, and feedbacks at a variety of scales, including large-scale dust transport and global biogeochemical cycles, climate mediated interactions between atmospheric dust and ecosystems, impacts on human health, impacts on agriculture, and interactions between aeolian processes and dryland vegetation. Aeolian dust emissions are driven largely by, in addition to climate, a combination of soil properties, soil moisture, vegetation and roughness, biological and physical crusts, and disturbances. Aeolian research methods span laboratory and field techniques, modeling, and remote sensing. Together these integrated perspectives on aeolian processes and the biosphere provide insights into management options and aid in identifying research priorities, both of which are increasingly important given that global climate models predict an increase in aridity in many dryland systems of the world.

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1. INTRODUCTION

[2] Aeolian processes (the entrainment, transport, and deposition of sediments by wind) affect almost all aspects of the biosphere and can have important consequences for landform evolution, biogeochemical cycles, regional climate, human health, and desertification [e.g., *Goudie and Middleton*, 2006; *Field et al.*, 2010]. Aeolian landforms are found in areas where wind is the primary agent of transport (erosion and deposition) such as in arid and semiarid regions, while elsewhere the effects of aeolian processes are often masked by the effects of hydrologic processes. Aeolian erosion, the wind-forced movement of soil particles, is influenced by geological and climatic conditions and human activities [e.g., *Shao*, 2008]. Perhaps the most recognizable evidences of aeolian activity on the Earth surface are the sand dunes of different forms and sizes observed in desert and coastal environments. There is a growing body of evidence showing that aeolian processes interact with ecosystems at different scales, contributing to important biophysical feedbacks between the biotic and abiotic components of the Earth system. Even though at the global scale water is considered to be the dominant erosion agent, wind is a major erosion driver in dryland systems, which comprise almost 40% of the Earth's land surface. In these systems aeolian processes, alone or in combination with hydrological processes, are considered to be major drivers of ecosystem processes [*Field et al.*, 2010; *Ravi et al.*, 2010].

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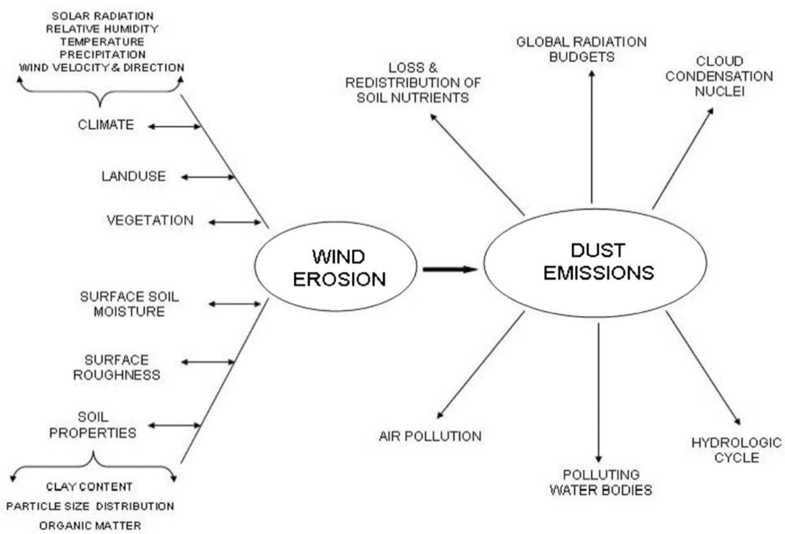


Figure 1. Conceptual diagram showing factors affecting wind erosion and the effects of wind erosion.

[3] The consequences of aeolian processes are diverse and important. At the field scale, wind erosion is associated with soil losses, while at the landscape and regional scales the entrainment and transport of atmospheric dust is associated with the redeposition of soil particles with consequent impacts on global biogeochemical cycles and climate. The effects of wind erosion and subsequent dust emissions include removal and loss of nutrient-rich topsoil particles [*Zobeck and Fryrear*, 1986; *Zobeck et al.*, 1989], abrasion and injury to growing plants [*Fryrear*, 1986a; *Armbrust and Retta*, 2000], air pollution and consequent aggravation of respiratory diseases [*Pope et al.*, 1995; *Choudhury et al.*, 1997; *Griffin et al.*, 2001], alteration of global biogeochemical cycles [*Duce and Tindale*, 1991; *Okin et al.*, 2004; *Mahowald et al.*, 2009], and changes in atmospheric radiation budgets [*Ramanathan et al.*, 2001; *Kaufman et al.*, 2002]. In arid and semiarid environments aeolian processes redistribute soil particles and nutrients [*Schlesinger et al.*, 1990; *Okin and Gillette*, 2001; *Ravi et al.*, 2007a] and thereby affect the soil surface texture and hydrologic properties [*Lyles*, 1975; *Offer et al.*, 1998; *Li et al.*, 2009a]. These changes in soil characteristics affect the composition, productivity, and spatial patterns of vegetation [*Schlesinger et al.*, 1990; *Ravi et al.*, 2007a, 2008].

[4] Atmospheric processes are notably sensitive to aerosols generated by aeolian processes. The entrained soil particles are transported in the atmosphere by advection and turbulent diffusion and are finally deposited through dry and wet removal [*Shao*, 2008]. The finer soil particles are preferentially eroded and carried away by wind over large distances [*Swap et al.*, 1996a], and large quantities of these nutrient-rich particles are deposited in the oceans and on other continents, with implications for global biogeochemical cycles [*Duce et al.*, 1991; *Okin et al.*, 2004]. Moreover, the global dust emission resulting from wind erosion is the largest source of tropospheric aerosols and affects the atmospheric radiation balance [*Tegen et al.*, 1996; *Ramanathan et al.*, 2001] as well as the nucleation and optical properties of clouds [*Rosenfeld et al.*, 2001; *Kaufman et al.*, 2005a]. These direct and indirect effects of aerosols can also lead to a weaker hydrologic cycle [e.g., *Hui et al.*, 2008], with effects on the availability of fresh water in the biosphere [*Ramanathan et al.*, 2001]. Collectively, there are several causes and effects of wind erosion (Figure 1).

[5] Understanding aeolian processes is critically important for managing the world's drylands. Wind erosion is thought to be the major cause of land degradation in arid and semiarid regions of the world. On a global scale, 432.2 million (M) ha of drylands are susceptible to wind erosion [*Middleton and Thomas*, 1997], with the continent of Africa having the largest wind erosion susceptible area (159.8 M ha) followed by Asia (153.1 M ha), Europe (38.7 M ha), North America (37.8 M ha), South America (26.9 M ha), and Australasia (15.9 M ha). In the United States wind erosion is the dominant problem on about 30 M ha of land area, and the soil loss by wind erosion is estimated to be 840 million tons per year [*Soil Survey Division Staff*, 1993]. In particular, wind erosion is a serious problem in the cultivated soils in the Great Plains, where wind erosion rates are comparable or may exceed those of water erosion [*Hagen and Woodruff*, 1973; *Ervin and Lee*, 1994; *Nordstrom and Hotta*, 2004]. In this region, dramatic soil losses and dust emissions (the "Dust Bowl") were observed in particular in the 1930s [*Bennett*, 1938; *Worster*, 1979] as a result of poor land use practices in conjunction with dry climatic conditions. Similar examples of intense aeolian activity induced by the combined effect of drought and land use can also be found in other regions of the world, for example, southern and West Africa, South America, and Australia [*Gillieson et al.*, 1996; *Buschiazzo et al.*, 1999; *Bielders et al.*, 2000].

[6] Aeolian processes are also recognized as major abiotic drivers in the Earth system, and there is a growing interest in the scientific community to quantify and model the biophysical drivers and biogeochemical implications of aeolian

processes at different scales [*Field et al.*, 2010; *Ravi et al.*, 2010]. Global climate models predict an increase in aridity in many dryland systems of the world [e.g., *Seager et al.*, 2007], which may enhance aeolian processes and/or modify the hydrologic-aeolian interactions with significant feedbacks on climate and desertification [*Ravi et al.*, 2010]. Therefore, understanding the drivers and implications of aeolian processes in changing climate, disturbance (both natural and anthropogenic), and management scenarios is fundamental to environmental change research. Here we provide a state of the science review of recent literature on the breadth of aeolian research, with a particular focus on interactions and feedback loops between aeolian processes and the biosphere.

[7] In summary, aeolian processes affect the biosphere in a wide variety of contexts, including landform evolution, biogeochemical cycles, regional climate, human health, and desertification. Collectively, research on aeolian processes and the biosphere is developing rapidly in many diverse and specialized areas, but integration of these recent advances is needed to better address management issues and to set future research priorities. Here we review recent literature on aeolian processes and their interactions with the biosphere, focusing on (1) geography of dust emissions, (2) impacts, interactions and feedbacks, (3) drivers of dust emissions, and (4) methodological approaches. Geographically, we evaluate spatial variation in aeolian processes. We summarize impacts, interactions and feedbacks for aeolian processes and the biosphere in the context of (1) large-scale dust transport and global biogeochemical cycles, (2) climate mediated interactions between atmospheric dust and ecosystems, (3) impacts on human health, (4) impacts on agriculture, (5) interactions between aeolian processes and dryland vegetation, and (6) interactions between hydrologic and aeolian processes. We then review five major drivers of dust emissions in addition to climate: soil particle size, soil moisture, vegetation and surface roughness, biological and physical crusts, and disturbances. We also contrast aeolian research methods among laboratory and field techniques, modeling, and remote sensing approaches. Finally, we consider all of these findings on aeolian processes collectively to discuss management options and aid in identifying research priorities, both of which are increasingly important given that global climate models predict an increase in aridity in many dryland systems of the world.

2. GEOGRAPHY OF DUST EMISSIONS

[8] Dust emissions vary with geography. The global dust cycle [*Goudie and Middleton*, 2006] plays a major role in the delivery of iron and other elements to the oceans [*Mahowald et al.*, 2005], and there are numerous studies that show that long-distance dust transport can affect geochemical conditions on land at great distances from dust sources [*Wagener et al.*, 2008; *Pulido-Villena et al.*, 2008]. For example, Saharan dust influences the nature of soils in Gran Canaria [*Menéndez et al.*, 2007], the mountains of Cameroon [*Dia et al.*, 2006], and more remarkably in Barbados, the Bahamas, Florida [*Muhs et al.*, 2007a], and the Andes [*Boy and Wilcke*, 2008]. Indeed, one of the reasons why mineral aerosol emissions have global impacts is because of the huge distances over which dust plumes move [*Zhu et al.*, 2007]. Thus, dust from the Lake Eyre Basin (Australia) accumulates in East Antarctica [*Revel-Rolland et al.*, 2006], Saharan and Asian dust reaches North America by way of the Pacific [*McKendry et al.*, 2007; *Fairlie et al.*, 2007; *Bennett et al.*, 2006; *Zdanowicz et al.*, 2006; *Han et al.*, 2008], North American dust storms deposit fine materials on the California Channel Islands and the eastern Pacific Ocean [*Muhs et al.*, 2007b], and large amounts of Saharan harmattan (dry and dusty West African trade winds) dust are blown southward into the Gulf of Guinea [*Resch et al.*, 2007]. The Sahara is also a major source of dust deposition into the Mediterranean Sea and neighboring countries [*Santese et al.*, 2007].

[9] In recent years a clearer picture of the main source regions for dust emissions at a global scale has emerged (Figure 2). Particularly important have been data from the Total Ozone Mapping Spectrometer (TOMS) [*Prospero et al.*, 2002; *Washington et al.*, 2003; *Schwanghart and Schütt*, 2008; *Engelstaedter and Washington*, 2007]. TOMS data have indicated that many of the world's major dust sources are areas of hyperaridity, with mean annual precipitation under 100 mm [*Goudie and Middleton*, 2001]. Table 1 shows the maximum mean Aerosol Index (AI, unit less index related to aerosol optical depth (AOD)) values of major global dust sources determined from TOMS. These data have demonstrated the prime position of the Sahara and have highlighted the importance of some other regions: Arabia, Taklamakan, southwest Asia, central Australia, the Etosha and Mkgadikgadi pans of southern Africa, the Salar de Uyuni of Bolivia, and the Great Basin in the United States. Most of the major source regions are large basins of internal drainage (Bodélé, Taoudenni, Tarim, Seistan, Eyre, Etosha, Mkgadikgadi, Uyuni, and the Great Salt Lake). The nature of lake basin surfaces, including their wetness, salt crust development, and texture may be highly significant [*Reheis*, 2006; *Reynolds et al.*, 2007]. Analysis of satellite images of areas like the Makran coast of Pakistan or the ephemeral rivers of Namibia indicate that dry river beds can also be important point sources [*Eckardt and Kuring*, 2005] as are deflation plains of fine material [*Sweeney et al.*, 2007]. The combination of climatic and geomorphic factors that control the importance of dust emissions from particular areas has been exemplified for two hot spots, the Seistan Basin of Iran and Afghanistan and the Tokar Delta of Sudan [*Hickey and Goudie*, 2007]. The former is a closed basin, has large expanses of lake and deltaic sediment created by the Helmand River, occurs in a hyperarid area, and is characterized by high-velocity winds associated with topographic channeling and strong pressure gradients. The Tokar Delta also occurs in a hyperarid region, has a ready source of readily deflated silt provided by the Baraka River, and experiences strong winds, including convectional "haboobs" (a type of intense dust storm or sandstorm), associated with a major gap in the Red Sea Hills.

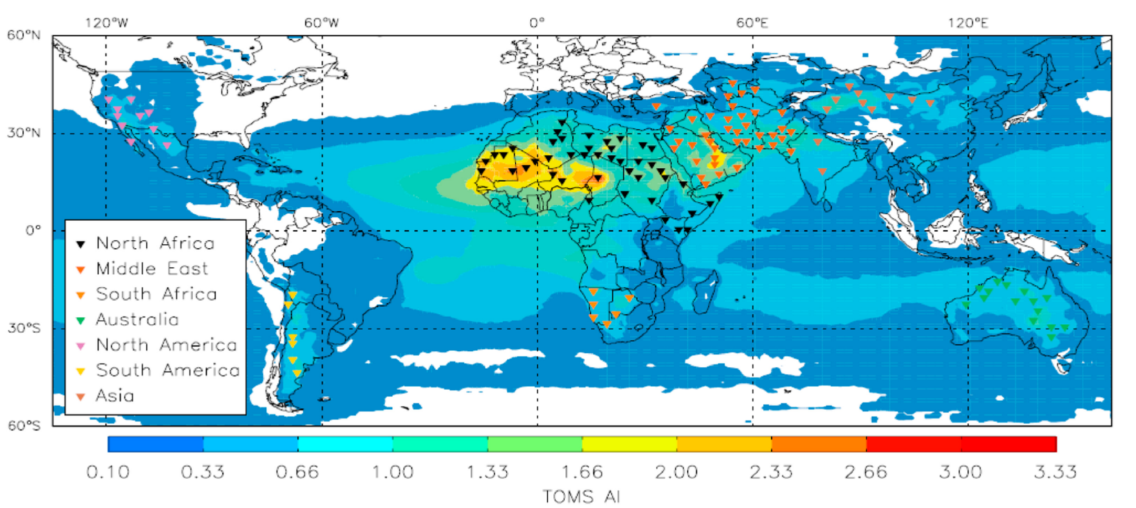


Figure 2. Location of global dust hot spots and long-term mean TOMS AI averaged over 1984–1990 [from *Engelstaedter and Washington*, 2007].

[10] When the relative importance of different regions for dust generation is considered, the importance of the Sahara is clear (Table 2). Estimates of annual global dust emissions vary between 1000 and 3000 Tg yr^{-1}. Saharan dust emissions are between 500 and 1000 Tg yr^{-1}, about half of the global total [*Engelstaedter et al.*, 2006; *Cakmur et al.*, 2006]. The next biggest source is China [*Tanaka and Chiba*, 2006], though estimates of its strength vary considerably. *Zhang et al.* [1997] suggested annual emissions of 800 Tg (comparable to the Sahara), whereas *Tanaka and Chiba* [2006] give a value of only 214 Tg and *Laurent et al.* [2006] give values between 100 and 460 Tg.

[11] Within North Africa, the main source areas for the dust were for long unclear [*Herrmann et al.*, 1999], but they are now known to include the Bodélé Depression in Niger and Chad; the Taoudenni region of southern Mauritania, northern Mali, and central southern Algeria; southern Morocco and western Algeria; the southern fringes of the Mediterranean Sea in Libya [*Bryant*, 1999; *O'Hara et al.*, 2006] and Egypt; and northern Sudan [*Brooks and Legrand*, 2000]. However, Bodélé is supreme [*Warren et al.*, 2007] and may alone be responsible for 6%–18% of global dust emissions, even though its surface area is relatively small [*Todd et al.*, 2007]. The reasons for this supremacy include the strength of the Bodélé low-level jet [*Washington and Todd*, 2005], topographic channeling by Tibesti, extreme surface wind gustiness [*Engelstaedter and Washington*, 2007; *Washington et al.*, 2006a], and the availability of large amounts of paleolake sediments, including diatomite,

TABLE 1. Maximum Mean Aerosol Index Values for Major Global Dust Sources Determined From TOMS[a]

Location	AI Value	Average Annual Rainfall (mm)
Bodélé Depression of south central Sahara	>30	17
West Sahara in Mali and Mauritania	>24	5–100
Arabia (southern Oman–Saudi border)	>21	<100
Eastern Sahara (Libya)	>15	22
Southwest Asia (Makran coast)	>12	98
Taklamakan, Tarim Basin	>11	<25
Etosha Pan (Namibia)	>11	435–530
Lake Eyre Basin (Australia)	>11	150–200
Mkgadikgadi Basin (Botswana)	>8	460
Salar de Uyuni (Bolivia)	>7	178
Great Basin of the United States	>5	400

[a]From *Goudie and Middleton* [2006].

TABLE 2. Estimates of Global and Regional Dust Emissions[a]

	% of Global Dust Emissions[b]	
Sahara, Sahel	50.7	
Central Asia	16.0	
Australia	14.5	
North America	5.2	
East Asia	4.9	
Arabia	4.2	
Others	4.5	
	Dust Emissions[c]	
	Tg yr^{-1}	%
North Africa	1430	69.0
South Africa	322	1.1
North America	9	0.4
South America	55	2.7
Asia	496	23.9
Australia	61	2.9
Total	2073	100
	Global Emissions in 1998[d]	
	Tg yr^{-1}	%
North Africa	1114	67.4
Arabian Peninsula	119	7.2
Asia	54	3.3
Australia	132	8.0
Miscellaneous	235	14.2
Total	1654	100

[a]From *Goudie and Middleton* [2006].
[b]Derived from data by *Miller et al.* [2004].
[c]Derived from data by *Ginoux et al.* [2001].
[d]Derived from data by *Luo et al.* [2003].

the surface of which is molded into large wind erosion and deflational features [*Warren et al.*, 2007; *Washington et al.*, 2006b; *Schwanghart and Schütt*, 2008].

[12] *Middleton* [1986a] has analyzed the distribution and frequency of dust storms in the Middle East and shows it to be one of the world's most important dust-generating areas. It also receives much dust from the Sahara. Dust hazes are common over the Arabian Sea, and aeolian silts have contributed to sedimentation in the Gulf of Oman, Arabian Gulf, Red Sea, and Arabian Sea [*Prins et al.*, 2000]. Dust storms reach a high frequency on the alluvial plains of southern Iraq and Kuwait. On the basis of the study of aerosol geochemistry over the Arabian Sea, *Pease et al.* [1998] have suggested that the Wahiba Sands of Oman is also a major source, while analysis of TOMS data indicates that the Oman-Saudi Arabia border is a large source that has not been picked up from ground meteorological observations [*Washington et al.*, 2003]. Also important is the eastern part of Saudi Arabia to the north of the great Rub Al Khali sand sea.

[13] With regards to southwest Asia, *Middleton* [1986b] has demonstrated that the greatest number of dust storms (on average 17–19 per year) occur at Ganganagar (northwest India) and at Jhelum and Fort Abbas (Pakistan). Dust plumes travel eastward into the Ganges Plain. As already mentioned, the Seistan Basin and the Makran coastal ranges are also dust hot spots. In the southern portions of the former Soviet Union there is a large zone where the number of dust storms exceeds 40 per year and some locations where there are more than 80, one of the highest occurrences in the world. The desiccated bed of the Aral Sea is one other major source.

[14] With respect to Chinese sources there has been debate as to the relative importance of the Taklamakan compared to the Gobi deserts [see, e.g., *Shao et al.*, 2003; *Xuan et al.*, 2004]. However, both sources are plainly very important [*Laurent et al.*, 2006] and have been responsible for the development of the great loess deposits downwind as well as for dust deposition in Korea and Japan. Attempts to identify local dust sources within the various Chinese deserts are provided by *Wang et al.* [2006], where the authors stress the importance of piedmont alluvial fans. More recently, *Zhang et al.* [2008] have used Moderate Resolution Imaging Spectroradiometer (MODIS) to identify dust source regions in both China and Mongolia and have found that the sandy lowlands of the eastern Mongolian Plateau are the principal contributors to long-range dust transport to the North Pacific.

[15] In the United States the greatest frequency of dust events occurs in the Great Basin, the panhandles of Texas and Oklahoma, Nebraska, western Kansas, eastern Colorado, the Red River Valley of North Dakota, and northern Montana [*Simonson*, 1995]. These areas combine erodible materials with a dry climate and high values for wind energy [*Gillette and Hanson*, 1989]. The spring months are the time of maximum dust activity [*Stout*, 2001]. Large amounts of dust, some toxic, are also blown off the bed of Owens Lake following its anthropogenically caused desiccation [*Reheis*, 1997]. A discussion of the spatial and temporal variability of dust storms in the Mojave and Colorado Plateau is provided by *Bach et al.* [1996], who identify the Coachella Valley (Southern California) as being the dustiest region. A detailed study of dust deposition in Nevada and California is provided by *Reheis and Kihl* [1995], while *Goldstein et al.* [2008] discuss the general composition of dust throughout the Southwest United States.

[16] Turning to the Southern Hemisphere, TOMS analyses indicate that there are two relatively small but clearly developed dust source areas in southern Africa. The most intense of these is the Etosha Pan of Namibia. The other center is the Mgkadikgadi Depression of Botswana [*Washington et al.*, 2003; *Bryant et al.*, 2007]. *Engelstaedter and Washington* [2007] have also reported nonlake hot spots of dust emission in southern Africa, including the Southern Kalahari (Figure 2). TOMS also identifies one area in South America where aerosol values are relatively high; this is the Salar de Uyuni, a closed basin in the Bolivian Altiplano that is located in an area with 200 to 400 mm of annual rainfall. This salt flat is not only the largest within the Andes, but is possibly the world's largest salt flat, though in the late Pleistocene it was the site of the huge pluvial Lake Tauca [*Lavenu et al.*, 1984]. The Patagonian Desert is also recognized as significant and has contributed dust to Antarctica [*Gaiero*, 2007; *McConnell et al.*, 2007].

[17] TOMS data, emission models, and ground meteorological observations have shown that Australia is not as dusty as the real dust hot spots (Tables 1 and 2). Dust emissions were low in the last few decades of the twentieth century, but have increased of late in response to drought and circulation changes. Thus, some of the late twentieth century deductions may be an underestimate of emissions [*Mitchell et al.*, 2010]. However, both at the present and in the past dust activity has contributed to sedimentation on and offshore. It is today the largest dust source in the Southern Hemisphere and in the Late Glacial Maximum contributed three times more dust to the southwest Pacific than now [*Hesse and McTainsh*, 1999]. The distribution of dust storm activity has been plotted from meteorological data by *McTainsh and Pitblado* [1987] and shows six areas of above average activity: central Australia, central Queensland, the Mallee, the eastern and western Nullarbor plains, and coastal Western Australia. Surface erodibility is indeed very important at a local and regional scale, and *Webb et al.* [2006] provide a methodology for recognizing erodible landscapes in the context of Australia. Substantial quantities of dust leave Australia in two main plumes: one that runs across the Tasman Sea toward New Zealand and another that heads westward into the Indian Ocean off north Western Australia [*Hesse and McTainsh*, 1999].

3. AEOLIAN PROCESSES: IMPACTS, INTERACTIONS, AND FEEDBACKS

[18] As noted above, aeolian processes have impacts, interactions, and feedback loops with the biosphere in a variety of contexts. Here we aggregate findings into six

categories: large-scale dust transport and global biogeochemical cycles, climate mediated interactions between atmospheric dust and ecosystems, impacts on human health, impacts on agriculture, interactions between aeolian processes and dryland vegetation, and interactions between hydrologic and aeolian processes.

3.1. Large-Scale Dust Transport and Global Biogeochemical Cycles

[19] The phenomena of long-range transport of mineral dust have been a subject of scientific inquiry in the Northern Hemisphere since at least the mid-nineteenth century [*Darwin*, 1846; *Pye*, 1984, 1987, 1995; *Goudie and Middleton*, 2006; *Pye and Tsoar*, 2008; *Goudie*, 2009]. Evidence for its study in Asia goes back even further [*Zhang et al.*, 1997]. Since the mid-1800s there has been a steady increase in the study of the long-range transport of aerosols and their local impacts.

[20] Initially much research focused on the radiative impacts of mineral aerosol deflation, transport and deposition. Broad discussions of the chemical and radiative characteristics and impacts of those aerosols have been presented elsewhere [e.g., *Buseck and Posfai*, 1999; *Buseck and Schwartz*, 2003]. Research on mineral aerosols started with observations of dustfall downwind of major source regions. Nearly coincident with those observations were studies that investigated the chemical nature of the dustfalls with additional objectives of determining the impacts on human health. Work in the 1970s–1990s began to focus upon the biogeochemical impacts of these dustfalls, both in adjacent regions as well as in far removed ecosystems well downwind of the original source site [e.g., *Duce et al.*, 1991; *Duce and Tindale*, 1991; *Swap et al.*, 1992]. With the advent of supercomputing and advanced remote sensing techniques, the use of air parcel trajectory calculations, mesoscale and global circulation models, and satellite data sets have driven the process of inquiry toward reducing the uncertainties in the temporal and spatial distributions of mineral aerosols [*Dulac et al.*, 1992; *Swap et al.*, 1996b; *Herman et al.*, 1997; *Kaufman et al.*, 2002, 2005b; *Evan et al.*, 2006], the mechanisms of transport and deposition, and their radiative and biogeochemical impacts [e.g., *Duce and Tindale*, 1991; *Tegen and Fung*, 1995; *Harrison et al.*, 2001; *Mahowald et al.*, 2008; *Tegen et al.*, 2006; *Engelstaedter et al.*, 2006; *Wagener et al.*, 2008].

[21] Perhaps the most studied mineral aerosol transport, from the Western perspective, is that from northern Africa. Long-range transport of North African mineral dust occurs over distances greater than 5000 km. It has been shown to reach northern Europe [*Reiff et al.*, 1986; *DeAngelis and Gaudichet*, 1991], the Middle East [*Levin et al.*, 1980], and North and South America [*Prospero and Carlson*, 1972; *Talbot et al.*, 1990; *Swap et al.*, 1992, 1996a; *Prospero and Lamb*, 2003]. Observations of the phenomenon have included ground-based [*Prospero et al.*, 1981; *Artaxo et al.*, 1990; *Swap et al.*, 1992], ocean-based [*Romero et al.*, 1999; *Harrison et al.*, 2001; *Stuut et al.*, 2002; *Pichevin et al.*, 2005], in situ [*Talbot et al.*, 1986, 1990], satellite-based [*Dulac et al.*, 1992, 1996; *Swap et al.*, 1996b; *Husar et al.*, 1997; *Chiapello et al.*, 1999; *Remer et al.*, 2008], and model-based assessments [*d'Almeida*, 1986; *Tegen and Fung*, 1995; *Andersen and Genthon*, 1997; *Zender et al.*, 2003; *Mahowald et al.*, 2003; *Schepanski et al.*, 2009].

[22] Evidence is growing that biogeochemical cycles of certain far-removed ecosystems are reliant upon the deposition and biogeochemical input of mineral aerosols. For example, the deposition of northern African mineral aerosols influences the biogeochemistry of oceanic and terrestrial ecosystems [*Reichholf*, 1986; *Muhs et al.*, 1990; *Duce et al.*, 1991; *Swap et al.*, 1992, 1996a]. The iron and phosphorus contained in atmospheric mineral aerosols are important micronutrient in ocean ecosystems, perhaps contributing to fluctuations of carbon dioxide on climatic time scales [e.g., *Martin*, 1990]. Mineral aerosols are also thought to impact terrestrial biogeochemistry and atmospheric chemistry [*Duce and Tindale*, 1991; *Okin et al.*, 2004; *Mahowald et al.*, 2005, 2008]. This is especially the case regarding the supply of iron-rich aerosols, the source of 95% of which is desert mineral dust [*Piketh et al.*, 2000; *Jickells et al.*, 2005; *Mahowald et al.*, 2005, 2009]. A recent synthesis of the considerable body of work that explores the Iron Hypothesis [*Martin*, 1990] and the impact of iron enrichment on high-nutrient, low-chlorophyll (HNLC) marine surface waters, found unequivocally that the lack of iron limits the production of one third of the world's HNLC regions [*Boyd et al.*, 2007]. The strong interest in the continued study of iron addition to the HNLC waters is focused upon whether the increased phytoplankton production can contribute to potentially high rates of carbon sequestration [*Meskhidze et al.*, 2007; *Garcia et al.*, 2008; *Smetacek and Naqvi*, 2008]. Mineral dust is often considered a major iron source to HNLC regions [e.g., *Duce and Tindale*, 1991]. However, recent work by *Wagener et al.* [2008] has challenged this view: new model-based estimates of mineral dust deposition to the HNLC regions of the Southern Ocean seem to indicate that mineral dust deposition is not the predominant source of iron to these regions of the world.

[23] The long-term terrestrial impacts of this deposition are evident in the chemistry of terrestrial sediments of Caribbean and Atlantic islands [*Glaccum and Prospero*, 1980; *Muhs et al.*, 1990; *Kremling and Streu*, 1993]. The deposition of northern African mineral aerosol has also been posited as having an impact on the shorter time scales of the biogeochemical cycles of the Amazon Basin [*Reichholf*, 1986; *Swap et al.*, 1992]. *Swap et al.* [1992] found that atmospheric deposition of mineral aerosol to the Amazon Basin occurs in an episodic fashion and contributes to standing stocks of nutrients. Further, modeling studies by *Okin et al.* [2004] and *Mahowald et al.* [2008] support these earlier findings of a biogeochemical dependency of the Amazon upon the transport and deposition of aeolian aerosols.

[24] Research concerning the study of Asian mineral aerosol transport and deposition is also very rich. The geologic record of the past several million years contains evidence of the transport and deposition of aeolian sediments from Asia across much of the Northern Hemisphere,

from China to Greenland [*Biscaye et al.*, 1997; *Rea et al.*, 1998; *Guo et al.*, 2002]. Annually, over 800 Tg of Asian mineral dust are deflated from desert sources [*Zhang et al.*, 1997]. Of that total amount, roughly 50% is transported and deposited within the Asian continent, with the remainder being subject to long-range transport downwind of Asia and out over the Pacific Ocean [*Zhang et al.*, 1997]. The transport can occur over distances on the same order of those associated with the transport of northern African mineral aerosols, with the North American continent receiving this mineral aerosol from the Asian continent [*Zhang et al.*, 1997; *Eguchi et al.*, 2009]. Studies have shown that the Hawaiian rain forests receive nutrient inputs both from marine aerosols (cation inputs) and dust from central Asia (phosphorus inputs) [*Chadwick et al.*, 1999]. In the long run, these atmospheric inputs are thought to become an important source of major biological nutrients, and may sustain the productivity of these Hawaiian rain forests in highly weathered soils [*Chadwick et al.*, 1999]. Recent work has focused on the use of geochemical techniques such as Nd and Sr isotope tracers to help identify natural and anthropogenic sources of Asian dust, especially for East Asia [*Li et al.*, 2009]. *Li et al.* [2009] found that while loess from the Tibetan and Chinese Loess Plateaus tended to dominate dust aerosol composition for much of East Asia, including Japan, aerosol composition around Beijing tended to reflect an increased contribution from anthropogenically impacted terrestrial sources to the adjacent north and west.

3.2. Climate Mediated Interactions Between Atmospheric Dust and Ecosystems

[25] Climate conditions affect vegetation cover, which in turn influences the rate of dust emission. In fact, the presence or absence of vegetation affects soil susceptibility to wind erosion (see section 3.5), the main mechanism of dust entrainment into the atmosphere. The opposite is also true: dust aerosols affect climate, with a consequent effect on the biota. In fact, aerosols absorb and reflect solar radiation (direct effect), thereby reducing the radiation reaching the Earth's surface. Thus, because of this direct radiative forcing, aerosols cool the near-surface atmosphere and reduce the potential evapotranspiration. This effect partly counteracts those of climate warming driven by increasing levels of greenhouse gases [*Ramanathan et al.*, 2001].

[26] The relative importance between light absorption and scattering processes is crucial for the dynamics and thermodynamics of the atmospheric boundary layer [*Yu et al.*, 2002]. Aerosol absorptive and scattering properties depend on aerosol concentration and on the presence of carbonaceous aerosols [*Andreae*, 2001], with atmosphere warming occurring as an effect of radiation absorption by carbonaceous aerosols, and cooling resulting from the presence of nonabsorbing aerosols such as desert dust [e.g., *Ramanathan et al.*, 2001; *Kaufman et al.*, 2002]. Significant iron in desert dust can, however, cause atmospheric warming through absorption in the visible and near-infrared (VNIR) wavelengths [*Sokolik et al.*, 1993]. Aerosols also affect cloud microphysical processes (indirect effect); aerosol particles act as cloud condensation nuclei (CCN). An increase in aerosols results in a partitioning of moisture into a larger number of droplets within the cloud, with the effect of increasing cloud surface area and the scattering of solar radiation [*Twomey et al.*, 1984; *Chuang et al.*, 2002]. As a result, both the lower atmosphere and the Earth's surface (first indirect effect) become cooler [*Albrecht*, 1989]. Thus, both the direct and the first indirect radiative effects of dust aerosols result in a cooling of the land surface, which in turn causes a decrease in convection and evapotranspiration. Aerosol's effect on convection is expected to be crucially important to the regional water balance in many dryland regions, which rely on convective precipitation [*Nesbitt and Zipser*, 2003]. At the same time, the larger concentration of CCN leads to a very inefficient condensation and coalescence of cloud droplets, due to the competition for water vapor arising among the large number of CCN existing within the same cloud (second indirect effect). High CCN concentrations can completely shut down precipitation in clouds with top temperatures higher than −10°C [*Rosenfeld et al.*, 2001; *Ramanathan et al.*, 2001; *Kaufman et al.*, 2002; *Hui et al.*, 2008]. It is still unclear to what extent mechanisms of rainfall suppression (second indirect effect), which are known to have a substantial impact on precipitation from shallow clouds [*Rosenfeld et al.*, 2001], would affect deep convective clouds [*Fuentes et al.*, 2008], though dust has been implicated in reducing convection that drives hurricanes in the North Atlantic [*Lau and Kim*, 2007]. The overall response of the global climate to atmospheric aerosols is a reduction in surface solar radiation, hence in latent heat flux and evapotranspiration [*Ramanathan et al.*, 2001]. The decreased evapotranspiration is expected to be balanced by a decrease in rainfall, with the effect of slowing down the hydrological cycle.

[27] At a regional scale, the effect of aerosols on the hydroclimatology of dryland regions is expected to be different; because of the high evaporative demand and of the limited soil moisture availability typical of these environments, the aerosol impact on evapotranspiration (direct and the first indirect radiative effects) might be limited. However, mechanisms of rainfall suppression (second indirect effect) could have a substantial impact on the water cycle [*Rosenfeld et al.*, 2001] and could explain the interannual persistence of droughts in some arid and semiarid regions [*N'Tchayi Mbourou et al.*, 1997; *Nicholson*, 2000; *Hui et al.*, 2008; *Cook et al.*, 2009].

[28] An interesting hypothesis of land-atmosphere dynamics (Figure 3) capable of sustaining drought conditions considers the role of aerosols [*Nicholson*, 2000; *Ramanathan et al.*, 2001; *Rosenfeld et al.*, 2001]: (1) the atmospheric dust content strongly depends on soil moisture and land cover conditions in the previous seasons; (2) it affects the hydrologic cycle, with the ability of decreasing the likelihood of rainfall occurrence; and (3) this reduction in precipitation increases the surface dryness and the rate of dust emission. This feedback mechanism between rainfall and dust aerosols [*Bryson and Baerreis*, 1967] recognizes the importance of surface soil moisture and vegetation

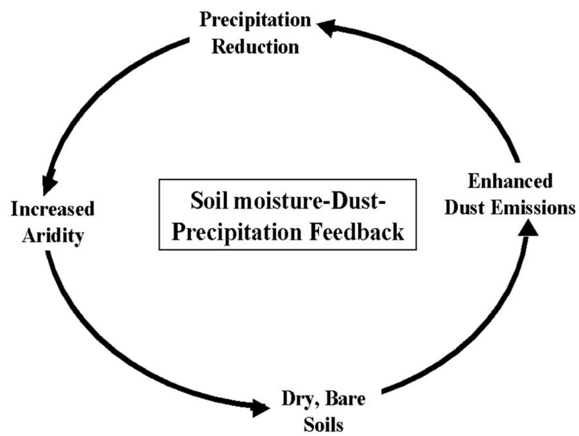


Figure 3. Schematic representation of possible feedbacks between dust emissions and ecohydrological processes.

conditions in the coupling with the atmospheric boundary layer, though this coupling does not act through water vapor fluxes as in other atmosphere-biosphere feedbacks [e.g., *Bonan*, 2002] but through soil erosion and dust emissions. It is still unclear whether this mechanism can affect the regional water balance and explain drought persistence in some dryland regions such as the Sahel [*Nicholson*, 2000; *Hui et al.*, 2008].

[29] Atmospheric dust also reduces the solar irradiance reaching the surface. This effect, combined with lower soil moisture values resulting from the decrease in precipitation, leads to smaller rates of evapotranspiration. However, the effect of atmospheric dust on photosynthesis and transpiration is in general not trivial. In fact, atmospheric aerosols affect the partitioning of photosynthetically active radiation into direct and diffuse irradiance. In densely vegetated regions an increase in diffuse radiation associated with atmospheric dust and other aerosols may enhance light penetration through the canopy, thereby favoring photosynthesis and transpiration [*Gu et al.*, 1999].

3.2.1. The Case of West African Sahel

[30] Instrumental records available from the West African Sahel indicate that in this region the rainfall regime exhibits patterns of multidecadal persistence. Dry anomalies have been observed almost continuously since 1968, while wet anomalies have occurred in almost every year during the 1950s and most of the 1960s [*Nicholson*, 2000]. Because large-scale climatic forcings (such as the sea surface temperatures (SSTs)) do not exhibit similar multidecadal patterns, it has been argued [*Nicholson*, 2000] that local dynamics of land-atmosphere interactions could explain the long-term persistence of droughts through positive feedbacks arising from land surface processes. A contributing factor to the drought persistence observed in this region could be precipitation suppression by mineral aerosols, which in turn could enhance land degradation, soil aridity, and dust production. This positive feedback could sustain an initially dry climatic anomaly. Other studies have shown that rainfall anomalies in the Sahel are indeed determined by SST variability, but are enhanced by land-atmosphere interactions [*Giannini et al.*, 2003]. Thus, Sahel rainfall responds mainly to the (ocean-forced) large-scale atmospheric circulation. However, positive feedbacks emerging from land-atmosphere interactions may enhance the persistence of rainfall anomalies initiated by anomalous SSTs. Early studies on the relation between dust emission and precipitation can be found in the work by *N'Tchayi Mbourou et al.* [1994, 1997], who used visibility records to show how the increased frequency of dust storms in the last decades is associated with the trend of decreasing precipitation. *Prospero and Nees* [1986] found that Sahelian dust measured at a downwind site is negatively correlated to rainfall in the previous rainy season [*Nicholson*, 2000]. The hydrological significance of these relations was investigated by *Hui et al.* [2008], who found a negative relation between daily rainfall amounts from raingage records in the Sahel and values of aerosol optical thickness (AOT) (a proxy for atmospheric aerosol concentrations) detected at upwind locations (Figure 4). A detailed and quantitative assessment of the physical processes affecting the role of the dust-precipitation feedback in West Africa is nevertheless still missing, in that it is still unclear whether any possible dust-induced decrease in precipitation is due to the effect of dust on convection or on cloud microphysics.

3.2.2. The Dust Bowl Drought

[31] Intense dust emissions occurred in the 1930s in the southern Great Plains of North America (from western Kansas to the Texas panhandle). These emissions resulted from low vegetation cover due to poor land management practices and drought [*Bennett*, 1938; *Worster*, 1979; *Schubert et al.*, 2004]. The conditions that occurred in this geographical region during this historical period are referred to as the Dust Bowl. Despite its clear association with dust emissions, until recently the dependence of the Dust Bowl drought on atmospheric dust has remained poorly understood. In fact, most studies have investigated the effect of drought on dust storms and aerosol concentrations, while the impact of atmospheric dust on drought persistence has been clarified only in the last few years [*Schubert et al.*, 2004;

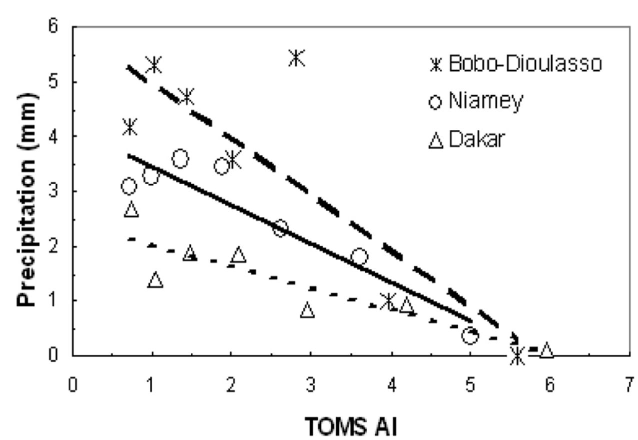


Figure 4. Negative relation between daily rainfall and AI using data from three rain stations in the Sahel (R^2 = 0.81, 0.87, 0.78) [from *Hui et al.*, 2008].

Cook et al., 2009]. Model simulations indicate that the drought was controlled both by anomalous sea surface temperatures and by regional dust emissions, which contributed to the amplification of the intensity and spatial range of the drought [*Cook et al.*, 2008]. By forcing a general circulation model with the sea surface temperatures from the Dust Bowl period, *Cook et al.* [2008] was able to reproduce the severity and spatial pattern of the 1930s droughts in the U.S. Great Plains. The dust emissions that resulted from human induced land degradation during the Dust Bowl period are thought to have amplified the drought [*Cook et al.*, 2009].

3.3. Impacts on Human Health

[32] Numerous studies have demonstrated the significance of the relationship between soil quality and human health and have shown that degradation of soil, either through excessive soil loss or pollution, can have significant ramifications on ecological functioning, agricultural productivity, and human health [*Pimentel and Sparks*, 2000; *Toy et al.*, 2002; *Lal*, 2001; *Lal et al.*, 2003; *McNeill and Winiwarter*, 2004; *Montgomery*, 2007]. Soil erosion processes transport and redistribute soil nutrients, soil organic materials, and sequestered contaminants, all with potential negative consequences to human health. This redistribution can impact human health, either indirectly (e.g., through agricultural productivity and diversity) or directly through inhalation of enhanced levels of fugitive dust, which could be contaminated with hazardous chemical, biological, or radiological contaminants. Another significant public health hazard arises during dust storms that may suddenly and seriously reduce air quality and visibility [*Skidmore*, 1994; *Blackburn*, 2006].

[33] Deleterious relationships between human health and wind erosion have been attributed to exposure to atmospheric dust [*Leathers*, 1981; *Griffin et al.*, 2001]. More locally, wind erosion, especially from neighboring agricultural lands, can cause significant increases in concentrations of respirable dust [*Saxton et al.*, 2001], which have been shown to increase respiratory distress and death rates [*EPA*, 2004a]. Though significant links exist between health effects and particulate mass concentrations of respirable particles (i.e., particles having aerodynamic diameters less than 10 μm and 2.5 μm, the so-called PM-10 and PM-2.5 standards), less is known about the specific organic and inorganic constituents in the air that cause the health detriment or the biological mechanisms for the health response [*EPA*, 2004a, 2004b]. Some of the primary constituents of the respirable dust include sulfate, nitrates, ammonia, sodium chloride, carbon, mineral dust, and water.

[34] Studies around the world have shown that pathogenic microorganisms and toxic chemicals can pose a public health threat through airborne transport [*Griffin et al.*, 2001]. A classic example is the case of the "Valley Fever" (Coccidioidomycosis), a fungal disease endemic to the arid regions in the Western Hemisphere [*Leathers*, 1981; *Kolivras et al.*, 2001]. Valley Fever outbreaks, common during the dry periods of the year, are linked to dust storms [*Kolivras et al.*, 2001]. The infection can occur when the spores of the soil-dwelling fungi become airborne during dust storms and are inhaled by humans and animals [*Kolivras et al.*, 2001]. Similar incidences of *meningococcal meningitides*, a bacterial disease, in the sub-Saharan Africa have been associated with drought and dust storm activity [*Griffin et al.*, 2001]. Another example is the case of *Escherichia coli* bacteria, observed both in airbone and settled dust in Mexico City, Mexico [*Rosas et al.*, 1997]. Certain strains of these bacteria are known to cause serious food poisoning in humans. Further, inhalation of dust containing toxins produced by microorganisms, such as endotoxins and mycotoxins produced by bacteria and fungi, respectively, are known to cause diseases and deaths [*Griffin et al.*, 2001] in humans.

[35] Wind erosion and transport of suspended dust from lands containing contaminated surface soil can cause increases in exposures to the airborne chemical and radioactive contaminants in the dust [*Anspaugh et al.*, 1975; *Larney et al.*, 1999; *Griffin et al.*, 2001; *Whicker et al.*, 2006]. In the case of dust storms originating from the Aral Sea, widespread use of agrochemicals has resulted in high concentration of chemical pollutants such as organophosphate and organochlorine pesticides [*O'Malley and McCurdy*, 1990; *Hooper et al.*, 1998; *Griffin et al.*, 2001]. The transport of these toxic chemicals during dust storms is a serious threat to human health. Further, this toxic dust pollutes the water bodies, thereby reducing the available sources of drinkable water and causing the accumulation of toxins in aquatic food sources [*Griffin et al.*, 2001].

[36] Even in environments with low rates of wind erosion, disturbances to the land, either through climate change or human interaction, can significantly increase wind erosion and atmospheric dust levels [*Tegen et al.*, 1996; *Whicker et al.*, 2008]. As noted, during the Dust Bowl, poor agricultural practices coupled with drought led to failed crops and massive dust storms [*Schubert et al.*, 2004; *Gill and Lee*, 2006]. The Dust Bowl illustrated how these combined environmental disturbances may lead to numerous health-related issues including increased prevalence of respiratory distress associated with the windblown dust. We are now seeing similar Dust Bowl-like situations in other countries, especially in portions of China [*Normile*, 2007]. More recently, wild fires combined with persistent dry conditions in the forests in Bryansk in Russia, which were contaminated by the Chernobyl disaster of 1986, escalated fears of transport of radioactive particles through enhanced postfire dust emissions, even though the actual health risks were determined to be very small [*Kelland*, 2010].

[37] The relationships between wind erosion, health risk, and environmental disturbance have also been noted in studies at U.S. Department of Energy sites where low levels of chemical and radioactive contamination are sequestered in surface soils [*National Academy of Science*, 1989]. These studies found that rates of wind and water erosion increased following ecosystem disturbances such as forest fires and

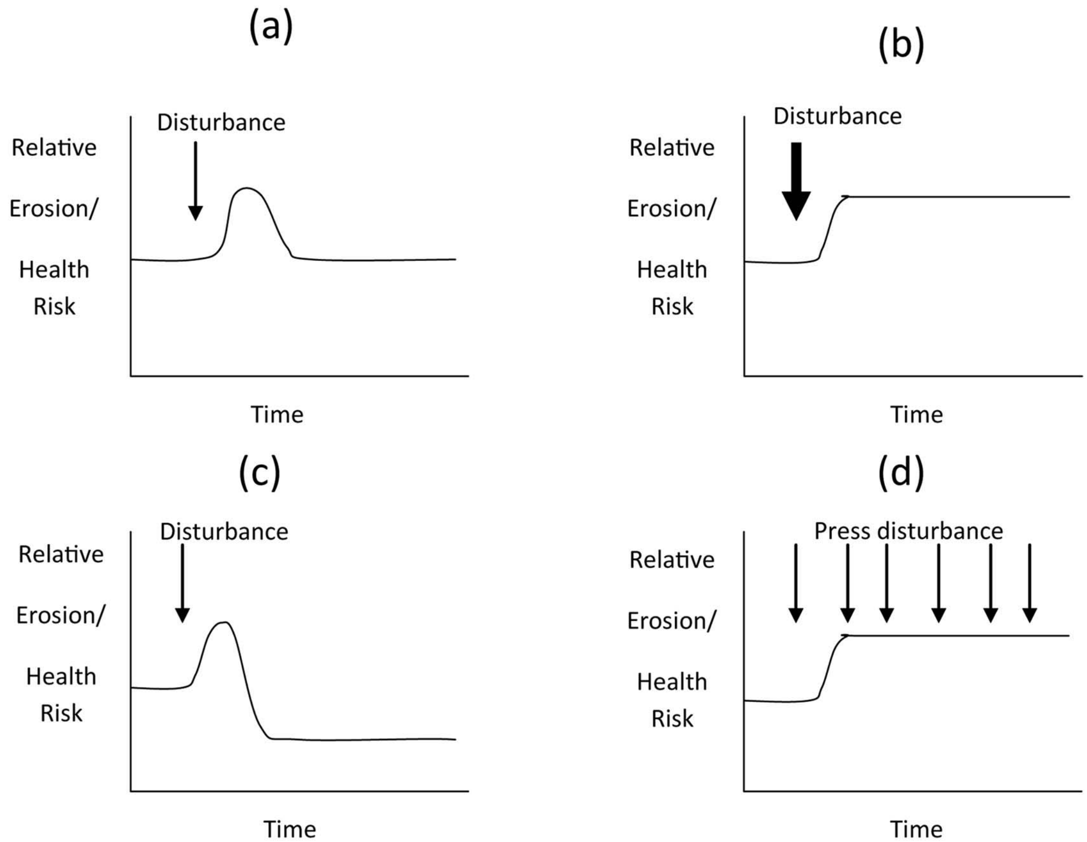


Figure 5. Conceptual model outlining the relationship between soil erosion and health risk for the following scenarios: (a) single disturbance with full ecosystem recovery, (b) single, large disturbance with altered ecological end state, (c) single disturbance with enhanced ecosystem recovery, and (d) a press disturbance with an altered ecological end state. Disturbance patterns and ecosystem response are based on the work by *Paine et al.* [1998] [from *Whicker and Breshears*, 2011].

tree thinning, and the increase in soil erosion rates was associated with increases in radionuclide concentrations in sediment in water and airborne dust [*Johansen et al.*, 2003; *Whicker et al.*, 2006]. Based on these data and the work of *Paine et al.* [1998], qualitative relationships have been proposed linking environmental disturbance, erosion of contaminated soil, and human risk (Figure 5) [*Whicker and Breshears*, 2011]. Environmental disturbances are expected to result in increased soil erosion following the disturbance, and then the erosion rates are hypothesized to follow the ecological recovery trajectory.

3.4. Impacts on Agriculture

[38] Wind erosion is a serious problem affecting agricultural areas in many arid and semiarid regions [*Lal*, 1994; *Dregne*, 1995; *Bielders et al.*, 2000; *Buschiazzo et al.*, 1999; *Gomes et al.*, 2003; *Nordstrom and Hotta*, 2004; *Shi et al.*, 2004]. Aeolian processes affect crop production mainly by altering soil resources (loss and redistribution) and by mechanical injury to crop plants. Wind erosion winnows the finer, more chemically active components of the soil, especially nutrients affecting plant growth (Figure 6) [*Lyles*, 1975; *Sterk et al.*, 1996; *Stetler et al.*, 1994; *Van Pelt and Zobeck*, 2007]. For example, cotton (*Gossypium hirsutum* L.) lint weights and kenaf (*Hibiscus cannibinus* L.) stem weights were reduced by 40% and sorghum (*Sorghum bicolor* (L.) Moench) grain yields were reduced by 58% in a study of a severely wind-eroded field in west Texas [*Zobeck and Bilbro*, 2001]. In addition to soil fertility degradation, the disproportionate loss of soil organic carbon (SOC) [*Van Pelt and Zobeck*, 2007] and soil fines may affect soil water infiltration and holding capacity, further affecting soil productivity in crop and rangelands. More recently it has been speculated that fine dust particles deposited on leaf surfaces may affect leaf physiological processes and may act as desiccants reducing the drought tolerance of plants [*Burkhardt*, 2010].

[39] In source fields, moving soil particles may sandblast crop plants and can seriously damage a seedling stand (Figure 7) [*Armbrust*, 1968; *Fryrear and Downes*, 1975; *Skidmore*, 1966; *Armbrust and Retta*, 2000]. A partially damaged stand often requires economically risky decisions concerning replanting [*Fryrear*, 1973]. However, for certain crops and certain growth stages, sandblast injury may result in increased rates of growth in surviving plants [*Baker*, 2007]. According to *Farmer* [1993], the deposition of

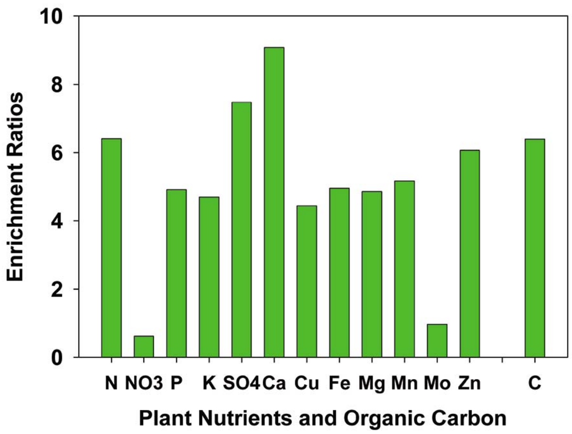


Figure 6. Fine dust enrichment ratios ((chemical species in dust)/(chemical species in source soil)) for selected plant nutrients and organic C [from *Van Pelt and Zobeck*, 2007]. The dust samples were taken from the air cleaner of a tractor used to till fields of Amarillo fine sandy loam. These fields frequently erode and release plumes of fugitive dust during wind storms. The study site was near Big Spring, Texas.

windblown soils on crops decreases their value and hinders processing. In certain parts of the world, however, agronomic ecosystems depend on the nutrient inputs from deposited dust [*Sterk et al.*, 1996]. Most wind-eroded soil is deposited very near the source field [*Okin et al.*, 2001a, 2001b; *Hagen et al.*, 2007]. Deposition of wind-driven sand along field margins, especially along weedy fence lines and in drainage ditches, results in costly, recurring maintenance tasks for landowners and government authorities.

[40] The development of land for production agriculture is often accomplished by total removal of native vegetation and at least some smoothing of the land surface, leading to the increased susceptibility of the soil to wind erosion. Conventional cropland tillage practices that lead to the increased susceptibility of the surface to wind erosion and dust emissions include plowing, leveling beds, planting, weeding, fertilizing, cutting and baling, spraying, and burning. As noted in section 3.2.2, the problem of wind erosion on cropped ground became very obvious in the semiarid Great Plains of North America during the drought years of the 1930s [*Worster*, 1979; *Baumhardt*, 2003]. Development of the disk plow and tractors enabled the cultivation of vast expanses of this former grassland. When the rains and the crops failed, the soil was left unprotected, and America's worst environmental disaster of the twentieth century, the Dust Bowl, resulted [*Baumhardt*, 2003]. During this period, wind erosion of rangeland and cropland reached an estimated 20 million hectares [*Hurt*, 1981], and a single "black blizzard" of this period is estimated to have resulted in soil being removed from the Earth's surface and mixed into the atmosphere at loading rates exceeding 2000 Mg km^{-3} [*Woodruff and Hagen*, 1972].

[41] One of the results of the Dust Bowl was the formation of the U.S. Department of Agriculture-Soil Conservation Service (USDA–SCS), later renamed the Natural Resources Conservation Service (NRCS), and funding for soil conservation research. As a result of the research and the efforts of the SCS then and NRCS recently, wind erosion today is much less than in previous years [*Stout and Lee*, 2003]. In spite of the downward trend, wind erosion on United States' cropland was 776 million tons in 2003 (U.S. Department of Agriculture–Natural Resources Conservation Service, National Resources Inventory data, 2007). In a study of summer fallow on the northern Great Plains, *Larney et al.* [1995] found that the topsoil loss in 1 year of measured erosion would require 17 years to replace at the fastest rate of soil development reported. The National Food Security Act of 1985 was passed by the U.S. Congress in an effort to insure future food production and readily available food in the United States. Part of that legislation specifies that annual soil loss rates greater than 10 Mg ha^{-1} are unacceptable and place the individual producer in a noncompliant state [*Bunn*, 1997, 1998, 1999].

3.5. Interactions Between Aeolian Processes and Dryland Vegetation

[42] Deserts inherently are patchy. Specifically, vegetation-covered areas in deserts are often interspersed with patches of bare ground. The patchiness is manifested at a wide range of spatial scales and is typically associated with a characteristic microtopography. The correspondence of vegetation patches with microtopographic features suggests strong interactions among surface transport, soil moisture, and vegetation patches and thus a close relationship between geomorphic and biological processes in dryland systems.

[43] In grasslands, bare and grassy patches alternate over a few decimeters and, on sloping ground, are often associated with a stepped topography [*Parsons et al.*, 1996; *Dunkerley and Brown*, 1999; *Nash et al.*, 2004]. In shrublands, the spatial scale of patchiness extends to a few meters and the microtopography may comprise bare swales and vegetation

Figure 7. Sandblast injury to cotton plants after exposure to sand abrasion for (left to right) 0, 5, 10, 20, 30, and 40 min (from *Baker* [2007], with permission, copyright American Society of Agronomy).

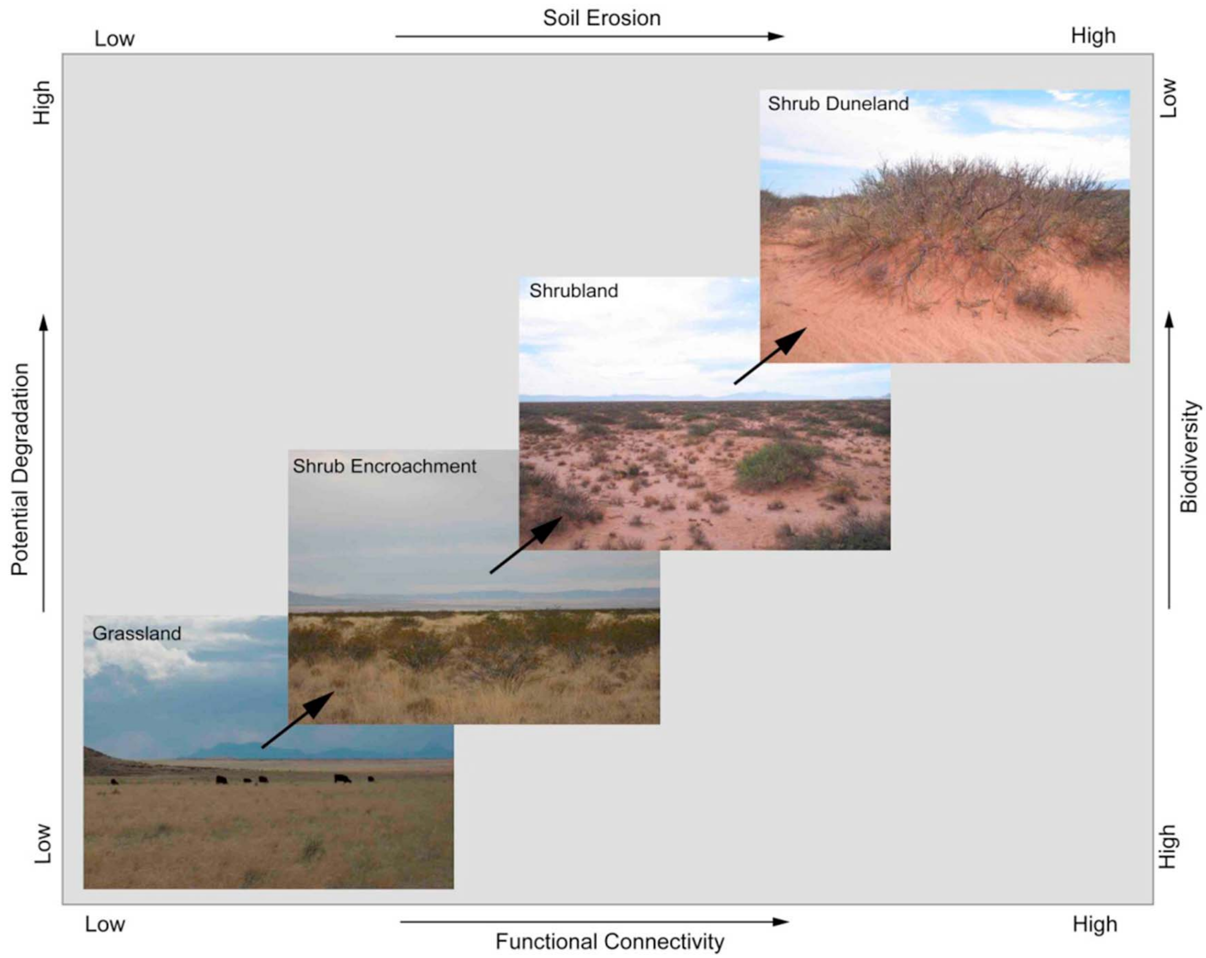


Figure 8. Conceptual diagram showing the stages of land degradation in the Chihuahuan Desert along with changes in functional connectivity, soil erosion rates, and biodiversity [from *Ravi et al.*, 2010].

atop mounds [*Parsons et al.*, 1992; *Rango et al.*, 2000]. In the presence of larger shrubs and trees, the landscape has been shown to consist of alternating concave-upward intergroves and "flatter" (or convex) groves with relatively steeper gradients, each of the scale of tens of meters in the downslope direction [*Berg and Dunkerley*, 2004].

[44] It has been argued that patchiness is an adaptation to resource limitation, specifically of water, that leads to greater biomass than could be maintained where the vegetation is more evenly distributed [*Tongway and Ludwig*, 1990; *Ludwig and Tongway*, 1995; *D'Odorico et al.*, 2006a; *Borgogno et al.*, 2009]. The presence of bare ground patches causes deserts to have a high albedo [*Charney*, 1975], thereby inducing surface cooling, reducing the potential for convection, and decreasing rainfall occurrences with the overall result of promoting the expansion of desert areas [*Taylor et al.*, 2002; *Zeng and Yoon*, 2009]. Moreover, bare soil influences rates of aeolian erosion and dust emission [*Li et al.*, 2007; *Okin*, 2008]; affects rates and patterns of water erosion [*Wainwright et al.*, 2002]; and has consequent significant impacts on the biogeochemistry at scales from individual patches [*Li et al.*, 2007; *Mueller et al.*, 2007] to the entire globe [*Schlesinger et al.*, 1990; *Okin et al.*, 2004].

[45] Patchiness in deserts provides bare ground gaps between vegetation, which are often connected to one another creating a network of conduits for the movement of water and soil resources (including sediment) borne by water or wind. The length of these connected pathways has been suggested by *Okin et al.* [2009] as a key element in the function of desert ecosystems, who argue that increasing length of the pathways (functional connectivity) is linked to desertification (Figure 8). Rapid changes in functional and structural connectivity resulting from rapid vegetation shifts (e.g., annual grass invasions of shrublands) is thought to induce dryland degradation by enhancing wind and water erosion [*Turnbull et al.*, 2008; *Ravi et al.*, 2009a, 2010].

[46] In arid and semiarid regions, aeolian processes redistribute sediments and nutrients with important effects on the soil resources and consequently on the composition and structure of vegetation [*Schlesinger et al.*, 1990]. Despite the relevance of vegetation-wind erosion interactions to the dynamics of arid and semiarid ecosystems, very few studies have addressed these interactions [e.g., *Ravi et al.*, 2007a, 2008, 2010]. For example, in grasslands encroached by shrubs, aeolian processes maintain local heterogeneities in nutrient and vegetation distribution through the removal of

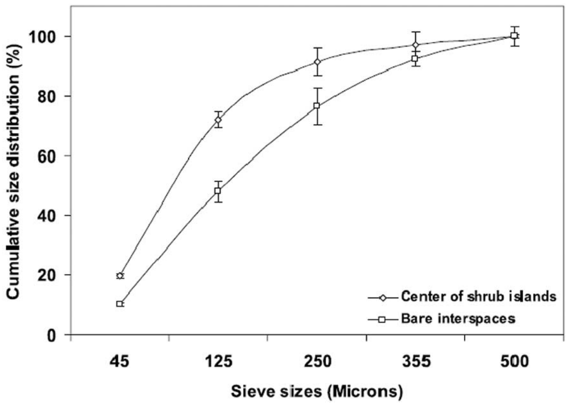


Figure 9. Aeolian deposition of fine particles by (clay, silt, very fine sand, and fine sand) in the center of the mesquite mounds in the Chihuahuan Desert. The soil particle size fractions (dry method, in five size classes) from the center and outer edges of mesquite shrubs mounds were compared using Ro-Tap Test sieve shakers (W. S. Tyler) [from *Ravi et al.*, 2007a].

nutrient-rich soil from intercanopy areas and subsequent deposition onto shrub vegetated patches (Figure 9) [*Schlesinger et al.*, 1990; *Okin and Gillette*, 2001; *Ravi et al.*, 2007a]. Similarly, studies have also shown that aeolian deposition can affect the growth pattern of grasses and shrubs and in some cases lead to plant mortality [*Ravi et al.*, 2007a, 2008]. A typical example is from the short grass ecosystems in the U.S. Great Plains in the 1930s, where soil deposits from dust storms caused mortality of blue grama grass [*Weaver and Albertson*, 1936].

[47] In many arid and semiarid environments, both water and wind erosion are known to be responsible for maintaining the spatial heterogeneity of vegetation cover and soil resource distribution. Water has been traditionally invoked as the main transport agent responsible for nutrient loss and redistribution in drylands [*Parsons et al.*, 1992; *Schlesinger et al.*, 2000; *Augustine and Frank*, 2001]. However, in many desert areas, water-based transport of soil nutrients and particulate matter is limited, especially in closed basins and on flat terrain or on soils with very high infiltration rates [*Gillette and Pitchford*, 2004]. Wind erosion, on the other hand, can remove the fine nutrient-rich particles from the soil surface regardless of the presence of a relief [*Larney et al.*, 1998]. The relative importance of wind and water erosion is further discussed in section 3.6.

[48] Recently, there has been increasing interest in quantifying the role of wind in creating patterned distribution of soil resources, such as the formation soil nutrient "fertile islands" found in desert landscapes around the world [*Schlesinger and Pilmanis*, 1998]. In a long-term study of wind erosion in the Chihuahuan Desert from 1933 to 1978, *Gibbens et al.* [1983] found that wind erosion accounts for large soil loss from a grassland-shrubland ecotone (a transition region between two adjacent but different plant communities). *Schlesinger et al.* [2000] suggested that aeolian processes must be partially responsible for the depletion and redistribution of soil nutrients in the degraded land in the northern Chihuahuan Desert. The results of enhanced aeolian processes on soil nutrient change have been demonstrated by an all-plant removal "scraped site" implemented in 1991 for an experiment aimed at measuring dust flux from loamy sand soils at the Jornada Experimental Range (JER), southern New Mexico [*Okin et al.*, 2006]. After more than a decade, up to 82% of N and 62% of plant available P were depleted from the surface soil of the scraped site [*Okin et al.*, 2001a]. More recently, a unique field-based, multiyear, replicated erosion enhancement experiment was set up at the JER to investigate the effects of wind erosion on soil nutrient depletion and spatial variation in desert grasslands [*Li et al.*, 2007, 2008, 2009a, 2009b]. Enhanced wind erosion was observed with various levels of grass cover reduction. Over three windy seasons, *Li et al.* [2007] found that increased wind erosion removed up to 25% of total organic carbon (TOC) and total nitrogen (TN) from the top 5 cm of soil, and about 60% of TOC and TN loss occurred in the first windy season (Figure 10). Similar but smaller reductions of soil nutrients, e.g., available N, in the surface soils were also observed over a 2 year period in the downwind plots, which were dominated by deposition [*Li et al.*, 2009a].

[49] *Li et al.* [2008] further examined the effects of wind erosion on the spatial heterogeneity of SOC and a variety of soil nutrients at the JER. In the study sites dominated by *Sporobolus* sp.-*Prosopis glandulosa* and *Bouteloua eriopoda*-*Prosopis glandulosa* vegetation, the coefficient of variation of SOC measured in 50 soil samples taken at a 5 × 10 m plot decreased consistently with the continuation of wind erosion. Geostatistical analyses show that aeolian processes appeared to increase the scale of spatial autocorrelation but decrease the scale of spatial dependence of most soil analytes over two to three windy seasons. These authors further observed that the overall consequences of wind on the grass cover reduction plots are the disappearance of small, well-defined fertile islands, which may be related to grasses, and the reinforcement of large fertile islands, which are likely related to mesquite shrubs. The change in the spatial patterns of SOC and soil nutrients induced by enhanced wind erosion may persist and reinforce soil fertility islands associated with shrubs. Nevertheless, this research highlights that biologically essential elements respond differently to enhanced wind. Specifically, results of this study show that soil organic matter related analytes such as SOC, TN, available nitrogen (N_{avail}), and SO_4^{2-} are among the first to be eroded and redistributed; cations such as Ca^{2+} and Mg^{2+} may not be removed and redistributed significantly; and other ions such as K^+, Na^+ and Cl^- showed no discernible pattern of change.

3.6. Interaction Between Hydrologic and Aeolian Processes

[50] Wind erosion is only one of two major erosional processes, with the other being water erosion. Notably, these two types of erosion should be viewed as interrelated and potentially competing processes [*Breshears et al.*, 2003; *Field et al.*, 2009]. More specifically, in the case of water-

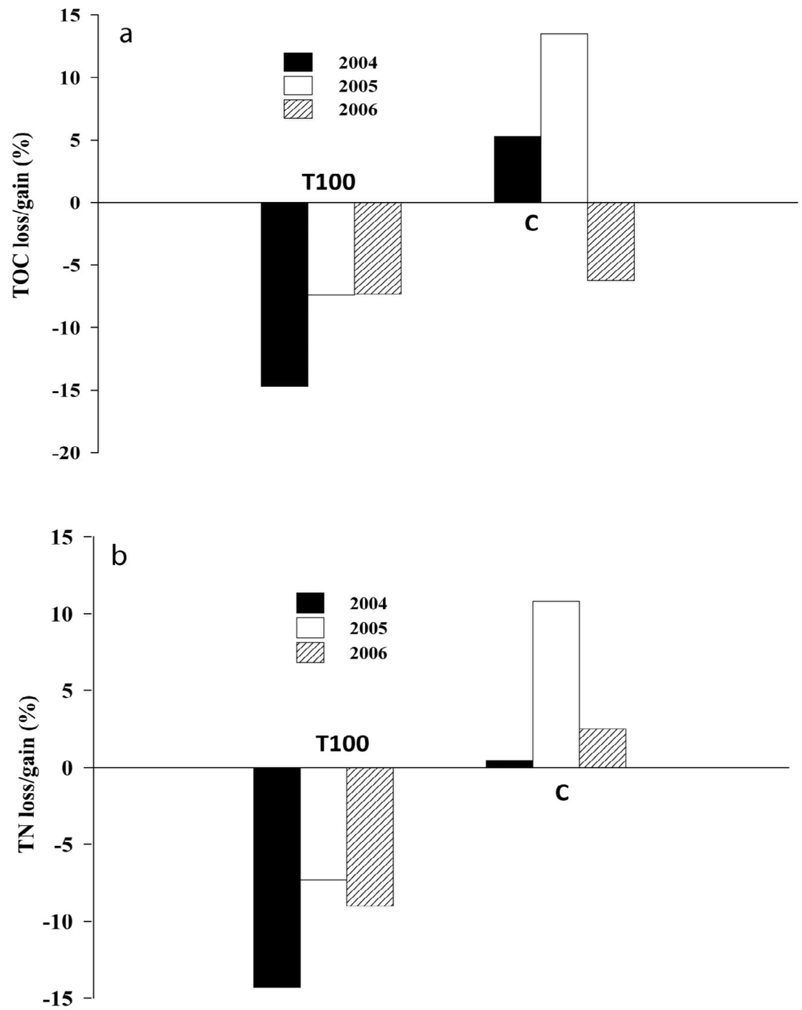


Figure 10. Yearly net loss or gain (%) of (a) TOC and (b) TN on the 100% grass cover reduction plot (T100) and the control plot (C) from 2004 to 2006. Net loss was donated by negative numbers, and net gain was represented by positive numbers [from *Li et al.*, 2007].

limited ecosystems, both wind and water erosion contribute significantly to the redistribution of soil and other resources and can occur almost simultaneously [*Visser et al.*, 2004], yet nearly all field-based measurements and empirical models have implicitly ignored important interactions between wind and water erosional processes. This is perhaps understandable given that they operate at different spatial scales (e.g., fetch length versus contributing area), yet many environmental issues require a site-specific, location-based estimate of erosion, and in many such cases it may be important to distinguish between the wind- and water-driven components. Assessing the relative roles of wind and water erosion, however, is of widespread importance. For example, roughly 80% of the world's arable land is affected by moderate to severe soil degradation [*Lal et al.*, 1989; *Pimentel*, 1993], most of which is attributed to wind and water erosional processes [*Oldeman et al.*, 1990]. Ultimately, the combined effects of wind and water erosion have degraded as much as one third of the world's arable land at rates that undermine long-term productivity [*Brown*, 1981]. Further, recent studies suggest that wind and water erosion are central drivers of desertification of nonarable arid and semiarid environments [*Peters et al.*, 2006; *Okin et al.*, 2009]. These major environmental impacts can translate into substantial economic impacts. For example, in the United States the combined effects of wind and water erosion are estimated to cost nearly $44 billion per year (1992 dollars) due to on-site and off-site agricultural impacts alone [*Pimentel et al.*, 1995].

[51] Despite the growing body of evidence that suggests that wind and water erosion are interrelated and co-occurring processes in arid and semiarid landscapes (Figure 11) [*Breshears et al.*, 2003; *Visser et al.*, 2004; *Ravi et al.*, 2010], most erosion studies measure only the wind- or water-driven component of erosion. Therefore, our current level of understanding about how these processes operate in tandem to redistribute soil and other resources (including key limiting nutrients) across the landscape is limited. In general, the absolute and relative magnitudes of wind and water erosion depend strongly on both climatic factors, such as wind speed

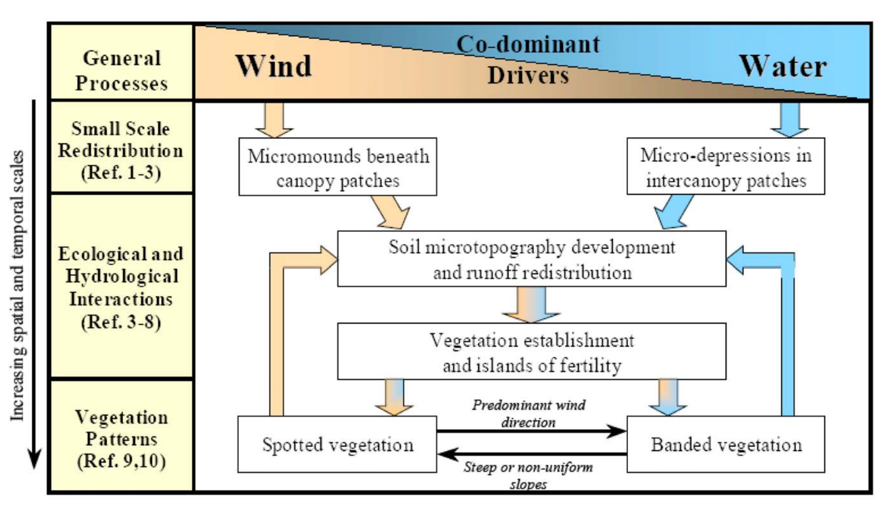


Figure 11. A conceptual framework that highlights the ecohydrological implications of small-scale soil redistribution in arid and semiarid ecosystems with patchy vegetation patterns. Numbers in parentheses represent related references as follows: 1, *Hennessy et al.* [1985]; 2, *Coppinger et al.* [1991]; 3, *Rostagno and del Valle* [1988]; 4, *Schlesinger et al.* [2000]; 5, *Tongway and Hindley* [2000]; 6, *Nash et al.* [2004]; 7, *Abrahams et al.* [1995]; 8, *Bhark and Small* [2003]; 9, *Mauchamp et al.* [1993]; 10, *Aguiar and Sala* [1999].

and rainfall amount, and the physical characteristics of the soil and surrounding vegetation, such as texture, surface crusting, and amount of woody canopy cover [*Visser et al.*, 2004; *Breshears et al.*, 2009; *Field et al.*, 2009].

[52] Although wind and water erosional processes have some aspects in common, several important fundamental differences exist between the two processes (Figures 12a and 12b). For example, certain physical characteristics of the soil and vegetation, such as infiltration rates, saturated conductivity, and percent basal cover, have a disproportionately greater influence on rates of water erosion, whereas other soil and vegetation characteristics, such as shallow (<1 cm) soil moisture content, surface roughness, and vegetation height, have a disproportionately greater influence on rates of wind erosion [*Zobeck et al.*, 2003a]. Perhaps one of the most obvious differences between wind and water erosional processes is the direction and dimensions of transport characteristics of these processes. Wind erosion is a two-dimensional, omnidirectional process that is at least partially reversible in response to changing wind directions, whereas water erosion is mainly a one-dimensional, unidirectional process, with the primary direction of transport being downslope and is largely irreversible [*Breshears et al.*, 2003; *Field et al.*, 2009]. In addition, wind-driven redistribution can occur in both the horizontal and vertical directions as either horizontal mass flux, which is generally thought to contribute to localized redistribution, or as vertical mass flux, which is more characteristic of regional or long-distance redistribution [*Zobeck et al.*, 2003a]. Thus, at large spatial and temporal scales, aeolian transport is expected to be dominant because fluvial transport is confined largely to channels and rivers within watershed boundaries, whereas aeolian transport is not confined to watersheds (Figure 12a). The greatest potential for aeolian-hydrologic interactions occurs at small to intermediate spatial and temporal scales and the degree of interaction is expected to decrease with increasing spatial scale [*Field et al.*, 2009] (Figure 12b).

[53] Important aeolian-hydrologic interactions occur also in time in addition to in space. For example, the alternation of dry and wet epochs in the Earth's history has led to the drying of lakes and rivers and the exposure of fine lake and river sediments to the erosive action of wind. Typical examples include the Aral Sea or the Bodélé Depression discussed in section 2. The mobilization of dunelands resulting from drought conditions may cause sand dune encroachment into dry river beds [e.g., *Bull and Kirkby*, 2002], while the shift to wetter climate conditions may lead to the formation of rivers and lakes in landscapes previously shaped by aeolian processes [e.g., *Ravi et al.*, 2010].

[54] The erosion and deposition of fine sediments by wind and water results in considerable changes in the soil properties creating a heterogeneous landscape with a mosaic of sources and sinks, with bare soil interspaces acting as sources and vegetated patches as sinks of nutrients and sediments [*Dunkerley*, 2002; *Wilcox et al.*, 2003; *Ludwig et al.*, 2005]. This heterogeneity in vegetation and soil resource distribution determines the heterogeneity in the spatial distribution of soil infiltration capacity and runoff and erosion rates, which in turn result in the formation of areas of hydrologically enhanced plant productivity [*Puigdefábregas*, 2005; *Rango et al.*, 2006]. Typical examples for the interactions among hydrologic and aeolian processes and vegetation dynamics are the formation of grass ring patterns and shrub coppice dunes. The growth patterns of desert grasses are strongly affected by the deposition of aeolian sediments

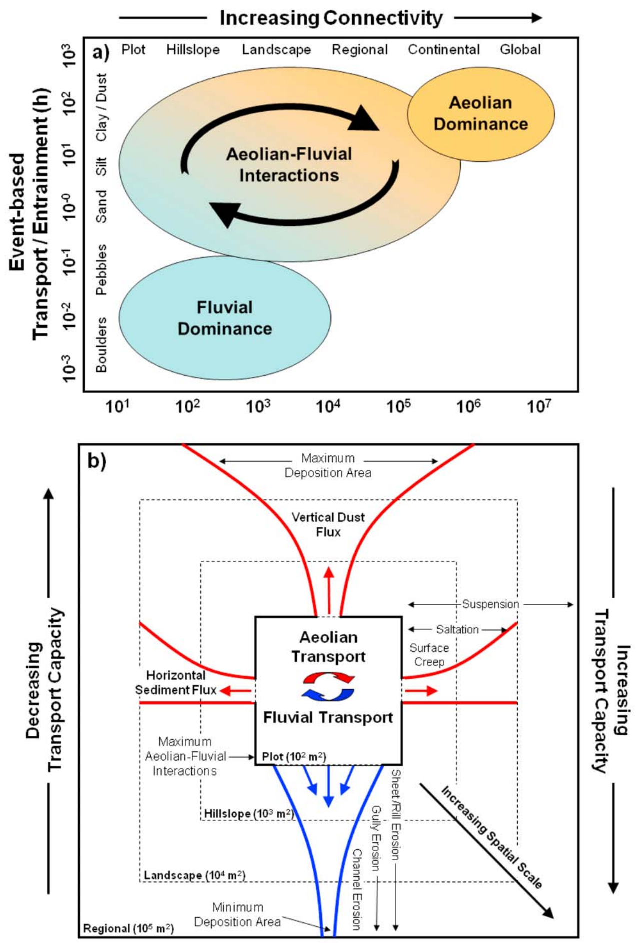


Figure 12. (a) Transport distances and event-based transport and/or entrainment times, highlighting differences between fluvial versus aeolian dominance and scales at which aeolian-fluvial interactions are potentially most important. (b) Scale-dependent interactions between aeolian and fluvial transport, highlighting maximum interactions at plot scale. Width between red or blue lines indicates the maximum depositional area. Note that the potential for fluvial sediment transport capacity increases with increasing scale, but the maximum deposition area simultaneously decreases. Horizontal aeolian sediment flux can move to hillslope and landscape scales, whereas vertical dust flux can extend to regional scales and has the maximum deposition area [from *Field et al.*, 2009].

[*Robertson*, 1939], resulting in a negative soil-plant feedback at the center of the bunch grasses [*Ravi et al.*, 2008]. The considerable changes in the soil texture inside the bunch grass caused by the deposition of fine aeolian sediments alter hydrologic processes like infiltration, soil moisture dynamics, and runoff [*Ravi et al.*, 2008]. Preferential (vegetative) recruitment of grasses occurs on the outer edges, while die off occurs in the center of the bunch grass leading to ring formation [*Ravi et al.*, 2008]. Similar mechanisms can explain the formation of shrub-coppice dunes, where differential rates of soil deposition and removal by aeolian processes result in differential rates of hydrological processes, thereby affecting the formation and expansion of these shrub-dominated coppice dunes [*Fearnehough et al.*, 1998; *Ravi et al.*, 2007a].

[55] Although wind and water erosion and associated transport processes operate at different spatial scales, they nonetheless both impact soil surface stability at any given location, and it is therefore important to investigate their impacts collectively rather than separately [*Breshears et al.*, 2003; *Field et al.*, 2009]. One approach for comparing rates of wind- and water-driven transport is to quantify the amount of transported material of each that crosses per unit length of a line that is oriented perpendicular to the erosion force [*Breshears et al.*, 2003]. For water erosion, the erosional force is parallel to the slope. In contrast, for wind erosion, the erosional force can be omnidirectional and is parallel with the wind direction. Consequently, wind transported material can move in one direction at one time and then subsequently move back in the opposite direction; the degree to which this occurs depends on the degree to which the prevailing wind direction dominates other wind directions. Some initial estimates of both wind and water erosion, based on time series measures of wind transported material and extrapolations from rainfall simulation of water transported material, indicate that wind transport can often dominate water transport in arid and semiarid landscapes. However, because wind transported material is a small fraction of net erosion loss at larger scales, wind and water erosion in different semiarid ecosystems can both be large enough that neither is negligible enough to ignore [*Breshears et al.*, 2003]. Studies are needed that develop methods and techniques to simultaneously quantify colocated rates of both wind- and water-driven transport [*Field et al.*, 2009].

[56] A possible approach for comparing rates of wind- and water-driven transport is to quantify the amount of transported material of each that crosses per unit length of a line that is oriented perpendicular to the erosion force (Figure 13) [*Breshears et al.*, 2003]. One of the only studies to date to use field-based measurements to explicitly evaluate both wind and water erosion in water-limited ecosystems using this approach, which included semiarid shrubland, grassland, and forest sites from the southwestern United States, found that horizontal wind-driven transport was greater than water-driven transport for all three systems (Figure 14) [*Breshears et al.*, 2003]. The shrublands are generally associated with 14%–40% cover, which results in wake interference airflow, maximizing the potential for wind erosion [*Wolfe and Nickling*, 1993]. More specifically, in the study by *Breshears et al.* [2003] rates of horizontal wind-driven transport were remarkably greater than rates of water-driven transport by up to a factor of 2200 at the shrubland, 4 at the grassland, and 2 at the forest. In addition, wind erosion, estimated as vertical mass flux, exceeded water erosion by 33 times at the shrubland and by 5 times at the forest; however, water erosion exceeded wind erosion by 3 times at the grassland site where soils had a higher clay

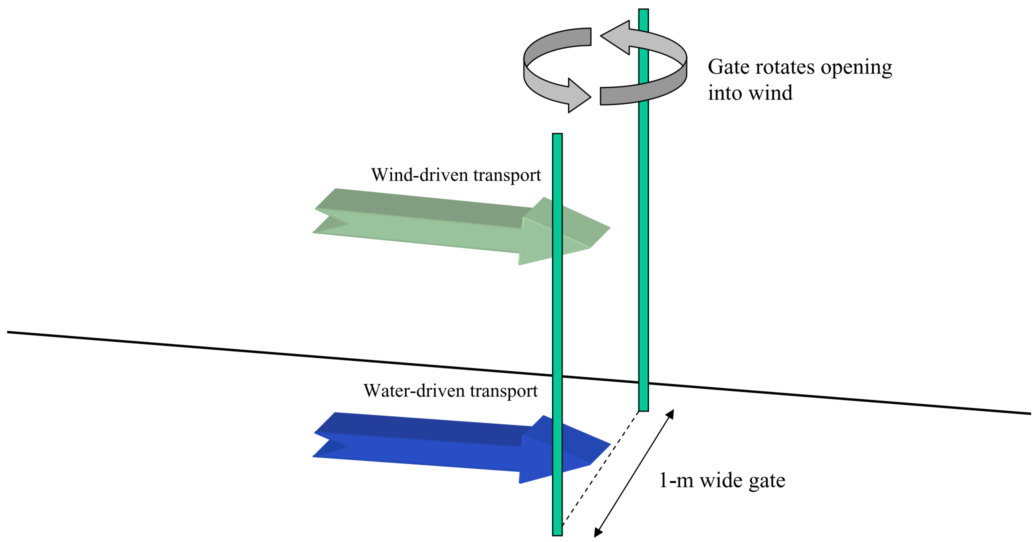


Figure 13. Conceptual comparison of horizontal wind- and water-driven sediment transport through a 1 m "gate." For water-driven transport, the gate has a fixed orientation that is perpendicular to the slope. For wind-driven transport, the gate rotates to orient perpendicular to the wind direction [from *Breshears et al.*, 2003].

content [*Breshears et al.*, 2003]. Figure 15 shows the hypothesized relationships for erosion and transport for shrubland, grassland, and forest ecosystems [*Breshears et al.*, 2003]. In addition, rates of horizontal wind-driven transport at another grassland that was a sandy loam exceeded rates of water-driven horizontal transport by factors of 3 to more than 40, depending on climate variability [*Field et al.*, 2011]. Such high annual rates of horizontal-driven dust flux can, in at least some cases, result largely from the accumulation of small but persistent weekly rates rather than predominantly from big wind events [*Field et al.*, 2011].

[57] Both wind and water erosion can have significant adverse impacts on soil productivity by reducing nutrient concentrations, organic matter content, soil biota, water holding capacity, infiltration rates, and soil depth [*Troech et al.*, 1991; *Abrahams et al.*, 1995]. Wind and water erosion can cause shortages of essential plant nutrients by selectively removing fine particles and organic debris, leaving behind coarse particles with minimal nutrient storage capacity [*Pimentel et al.*, 1995]. In systems with patchy vegetation cover, soil and nutrients can be transported offsite or redistributed from the source areas to the sink areas through interactions between wind and water erosional processes. Over time, the redistribution of soil from source to sink areas may reinforce patterns of wind and water erosion, leading to further degradation of source areas and additional enrichment of sink areas. Arid regions are particularly susceptible to increases in rates of wind and water erosion after disturbance such as fire or livestock grazing [*Whicker et al.*, 2002; *Ludwig et al.*, 2005; *Field et al.*, 2011] (see also section 4.5). For example, proxy records of dust deposition from high-elevation lakes in the southwestern United States indicate that dust load levels have increased by 500% above the late Holocene average, likely due to land use change and disturbance associated with the expansion of livestock grazing in the early twentieth century [*Neff et al.*, 2008]. Soil erosion and dust emission rates in the southwestern United States and other arid regions will likely continue to increase in coming decades due to projected climate change [*U.S. Climate Change Science Program* (*USCCSP*), 2008]. The high probability of increased aridity across many water-limited regions in conjunction with widespread anticipated increases in wind speed, temperature, and drought frequency suggests that wind erosion and dust emissions from arid lands will become increasingly important in the coming decades, likely causing substantial continental-scale impacts on downwind ecosystems, air quality, and populations [*USCCSP*, 2008].

4. DRIVERS OF DUST EMISSIONS AND THEIR CONTROLS

[58] The physical processes conducive to wind erosion are complex in nature. Three distinct mechanisms of soil particle transport by wind were described by *Bagnold* [1941] as suspension, saltation and soil creep (Figure 16). These major

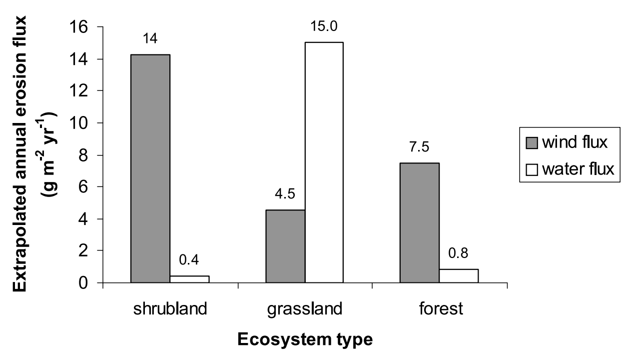
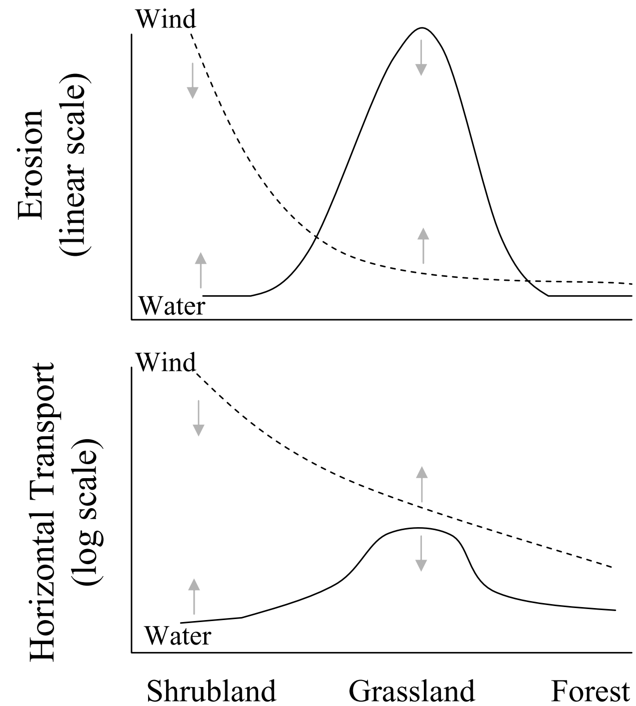


Figure 14. Annual erosion rates of wind and water among three ecosystem types [from *Breshears et al.*, 2003].



	Shrubland	Grassland	Forest
Precipitation	Low	Medium	High
Near-surface wind velocity	Medium	High	Low
Ground cover	Low	Medium	High
Mean bare patch size	High	Medium	Low
Soil texture (clay content)	Sand (Low)	Clay (High)	Silt Loam (Medium)

Figure 15. Hypothesized relationships for erosion and transport for shrubland, grassland, and forest ecosystems. Factors related to the hypothesized trends are listed below each ecosystem. The hypotheses are specific to the soil textures listed. Arrows indicate the expected direction in which the hypothesized curves would shift if all soil were adjusted to intermediate texture [from *Breshears et al.*, 2003].

types of soil grain motion were classified based on the particle size. Sand-size particles are transported by saltation (60–2000 μm) and soil creep (>2000 μm), while smaller-size particles like clay and silt are transported to larger distances by the process of suspension (<60 μm). Dust emissions are due either to the suspension of fine-size particles (i.e., silt and clay) present in the soil or to the production (and subsequent suspension and entrainment) of dust-sized particles through abrasion of saltating mineral particles [*Kuenen*, 1960; *Smith et al.*, 1991; *Shao et al.*, 1993; *Wright*, 2001; *Bullard et al.*, 2004; *Hagen*, 2001; *Bullard and White*, 2005; *Mackie et al.*, 2006; *Bullard et al.*, 2007].

[59] The entrainment of particles occurs when the wind shear at the soil surface exceeds the shear strength of the aggregates and their resistance to detachment and removal. The wind shear velocity needs to exceed a certain minimum value, the "threshold shear velocity," for soil erosion to occur [e.g., *Shao*, 2008]. Wind speed controls the erosive action, while field surface conditions, size and shape of the aggregates, and clay content as well as near-surface soil water content affect the ability of soils to be eroded, i.e., the values of the threshold shear velocity [*Chepil*, 1953, 1958; *Belly*, 1964; *Gregory and Darwish*, 1990; *Fecan et al.*, 1999]. Both theoretical [*Bagnold*, 1941; *McKenna Neuman and Nickling*, 1989; *Fecan et al.*, 1999; *Shao and Lu*, 2000] and empirical [*Chepil*, 1956; *Belly*, 1964; *Bisal and Hsieh*, 1966; *Saleh and Fryrear*, 1995] methods have been suggested in the past to express the threshold friction velocity as a function of these factors. The threshold shear velocity is also controlled by other factors in addition to grain size and moisture content, including surface soil compaction, presence of soil crusts, and vegetation cover [*Chepil*, 1958; *Hagen*, 2001].

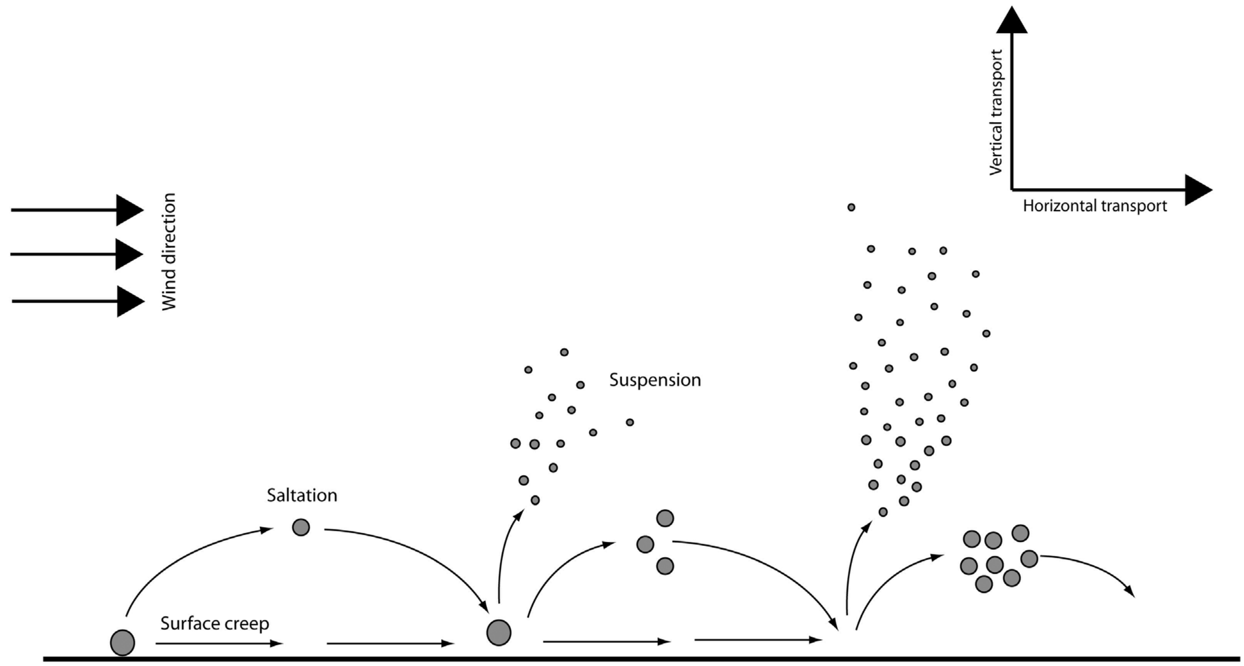


Figure 16. Three distinct phases of motion of the soil particles in wind erosion: suspension, saltation, and soil creep.

[60] A soil particle at the surface experiences several forces under the influence of an air stream (Figure 17), namely the aerodynamic forces (aerodynamic drag (F_d) and the aerodynamic lift (F_l)) and the stabilizing forces (the gravity force (F_g) and the interparticle cohesive force (F_i)) [*Shao*, 2008]. The classical theoretical approach to wind erosion studies involves deriving an equation for threshold friction velocity (u_t^*) from the balance of forces experienced by a soil particle at the point of threshold or initiation of particle motion [e.g., *Bagnold*, 1941; *McKenna Neuman and Nickling*, 1989; *Shao*, 2008; *Ravi et al.*, 2006a]. The early theoretical models referred to soil particles with spherical geometry: as noted, *Bagnold* [1941] derived an expression for threshold friction velocity (u_t^*) of dry soils affected only by aerodynamic drag and gravity:

$$u_t^* = \mathrm{A}\sqrt{\frac{(\rho_s - \rho_a)}{\rho_a} gd}, \tag{1}$$

where d is the particle diameter, ρ_a is the air density, ρ_s is the grain density, g is the gravitational acceleration, and A is a dimensionless threshold parameter. Drivers of dust emissions that merit specific consideration include particle size, soil moisture, vegetation and roughness, biological and physical crusts, and disturbances.

4.1. The Effect of Particle Size

[61] *Bagnold* [1941] observed that threshold friction velocity (u^*) depends on particle diameter and density as well on the density of air. For friction Reynolds numbers greater than 3.5, the value of threshold velocity varied as the square root of grain diameter (equation (1)). Wind tunnel experiments [*Bagnold*, 1941; *Chepil*, 1945] showed that this relationship was not valid when the grain size decreases below a critical value (60 μm). It was observed that there is an increase in threshold friction velocities with decreasing grain size for particles smaller than this critical value due to increased interparticle cohesion forces [*Iversen and White*,

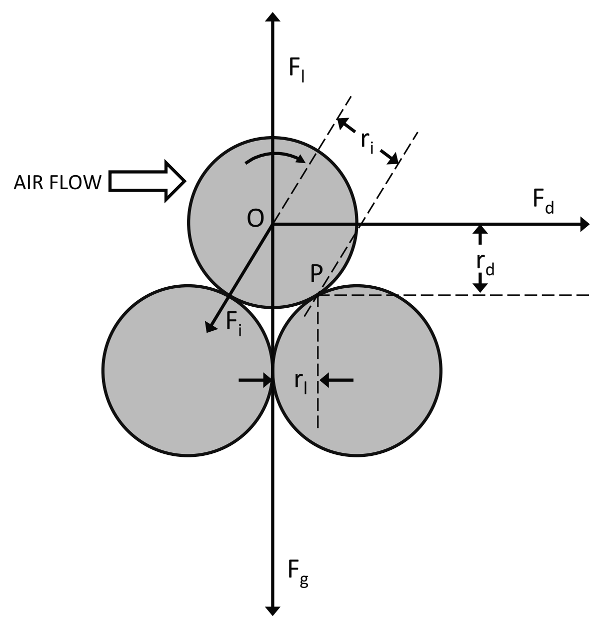


Figure 17. Forces acting on a particle at the threshold of motion (modified from *Shao and Lu* [2000]). O is center of gravity of the particle, and P is the pivot point for particle entrainment. The aerodynamic forces are the aerodynamic drag (F_d) and lift (F_l); the stabilizing forces are the gravity force (F_g) and the interparticle cohesive force (F_i).

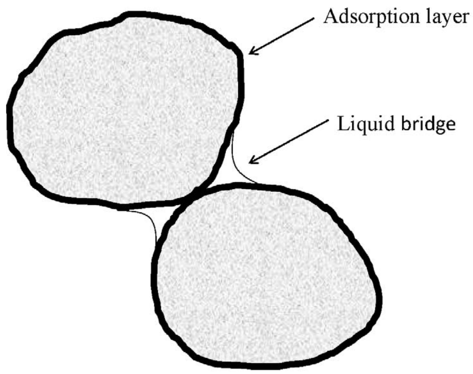


Figure 18. Schematic illustrating soil water held in liquid bridges and in the adsorption layer. In air-dry soils (RH < 65%), the adsorptive component dominates the wet bonding forces because the soils are too dry for the liquid bridge bond to exist. In higher-humidity conditions (RH > 65%) water condenses into liquid bridges between the soil grains, and then the liquid bridge bonding dominates the wet bonding forces (modified from *Ravi et al.* [2006a]).

1982; *Marticorena and Bergametti*, 1995]. These forces become increasingly important as the particle size becomes smaller.

4.2. The Effect of Soil Moisture

[62] Soil moisture can be the most important factor controlling the changes in soil erodibility at short time scales (e.g., diurnal). This fact calls for an accurate understanding of the dependence of wind erosion on both wind speed and near-surface soil moisture. Both theoretical [*Bagnold*, 1941; *McKenna Neuman and Nickling*, 1989; *Fecan et al.*, 1999; *Cornelis and Gabriels*, 2003; *Cornelis et al.*, 2004a] and empirical studies [*Chepil*, 1956; *Belly*, 1964; *Bisal and Hsieh*, 1966; *Saleh and Fryrear*, 1995] have investigated the dependence of threshold friction velocity on soil moisture. *Chepil* [1956] showed that interparticle cohesion forces due to soil water retention can explain the influence of moisture on the threshold velocity and that these effects depend on soil texture. *Belly* [1964] experimentally demonstrated that as the moisture content of a sandy soil increases the threshold shear velocity increases.

[63] Early theoretical expressions for thresholds, such as the Bagnold's equation, does not include interparticle forces between soil particles, which become increasingly important as the particle size becomes smaller and smaller [*Shao*, 2008]. The interparticle forces that bind soil particles to one another are electrostatic forces, van der Waals forces, and forces due to presence of moisture in the area of contact between the adjacent grains [*Cornelis et al.*, 2004b].

[64] Soil moisture is either adsorbed on the grain surface or present as "liquid bridges" on the wedges or spaces between interparticle contact areas (Figure 18). In both cases moisture directly contributes to the interparticle forces and significantly affects the entrainment and supply of grains to the air stream [*Belly*, 1964; *Ravi et al.*, 2006a]. For sandy soils the interparticle forces contributed by this adsorbed layer can be considered negligible when the soil is relatively wet, while adsorbed water may have a significant influence in clayey soils. In air-dry soils where adsorption forces (adsorption of moisture as a film over the soil grains) dominate, the effect of capillarity is negligible, and thus the effect of surface tension of the air-water interface is not accounted for in the expression of interparticle bonding forces [*Haines*, 1925]. At higher moisture levels, capillary forces contribute to wet bonding because water condenses to form liquid bridges between soil grains. In this scenario, the effect of surface tension was added to interparticle force equation for spherical soil grains by *Fisher* [1926]. *McKenna Neuman and Nickling* [1989] considered a more general geometry of soil particles and calculated the interparticle capillary forces acting through dissymmetric, conical contact areas. *Fecan et al.* [1999] generalized *McKenna Neuman and Nickling*'s [1989] equation to the case of clayey soils. This model was tested using wind tunnel studies, and the results showed that most sands are exceedingly resistant to wind erosion at gravimetric moisture contents above 0.2%. However, these authors did not explicitly account for interparticle adhesion forces. These forces were included by *Cornelis et al.* [2004b], who accounted for electrostatic, van der Waals, and wet-bonding forces, including both adhesion and capillarity. *Ravi et al.* [2006b] accounted for the dependence of threshold shear velocity on soil water repellency, through its effect on the contact angle between soil grains and the air-water interface. *Gregory and Darwish*'s [1990] model investigated the effect of water adhesion in air-dry soils. These authors noted that atmospheric variables such as temperature and specific humidity could be better predictors of soil erodibility than surface soil moisture, due to the difficulties commonly experienced in the accurate measurement of surface soil water content. This approach was recently adopted by *McKenna Neuman* [2003] and *Ravi et al.* [2004, 2006a], who investigated the effect of temperature on soil moisture and threshold shear velocity. This atmospheric humidity dependence is expected to be significant (Figure 19), especially in arid environments and during the dry seasons, when moisture at the ground surface is neither supplied by capillary rise from a water table nor by precipitation [*Ravi et al.*, 2004; *Ravi and D'Odorico*, 2005].

4.3. The Effect of Vegetation and Surface Roughness

[65] Vegetation provides a sheltering effect to the soil surface in that it absorbs a fraction of the wind momentum flux. This effect increases with increasing surface roughness [*Stockton and Gillette*, 1990]. The effectiveness of vegetation cover in protecting the soil surface from erosion depends upon the vegetation type and orientation. It has been shown that the soil erodibility depends strongly on soil texture and ground surface characteristics [*Gillette*, 1979; *Gillette and Stockton*, 1989] like vegetation cover and the presence of clods, rocks, and crop residues. A major factor affecting the threshold velocity in natural situations is the presence of nonerodible roughness elements. Apart from

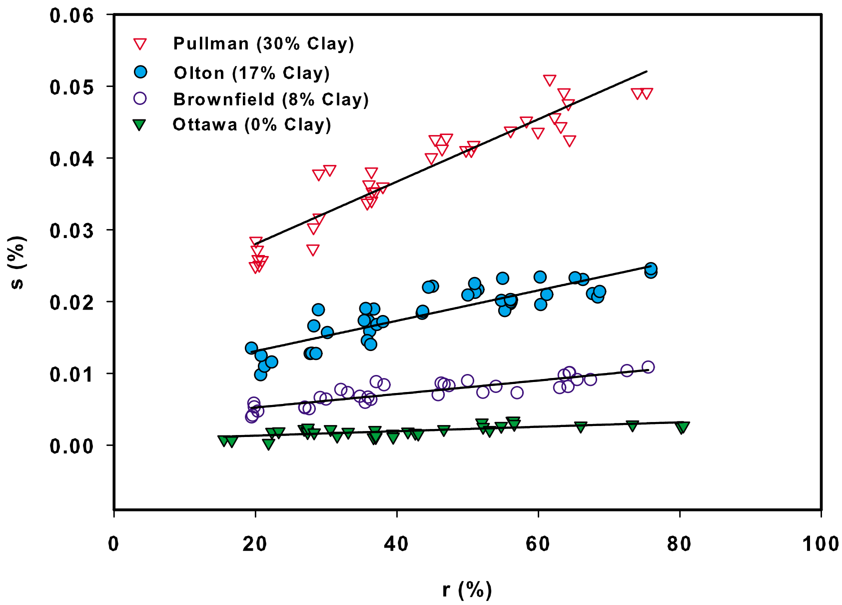


Figure 19. Relation between surface (top 2 mm) soil moisture (s) and near-surface relative humidity (r) for four soil types (modified from *Ravi et al.* [2004]).

providing a protective cover over the soil surface, the roughness elements also reduce the transfer of wind energy to the erodible surface. The roughness elements decrease the wind stress on erodible surface by absorbing a significant fraction of the downward momentum flux from the airflow above [*Raupach*, 1992; *Raupach et al.*, 1993]. Experimental studies have shown that the wind erosion thresholds observed on rough surfaces are significantly higher than those observed on smooth surfaces [*Musick and Gillette*, 1990; *Gillette and Stockton*, 1989]. Soil roughness can be contributed by plants, gravel, or soil aggregates. The amount, stability, and placement of soil aggregates on the soil surface are major factors affecting the susceptibility of the soil to wind erosion.

[66] The wind shear stress or drag (τ) exerted on the soil surface is related to the shear velocity (u*) as $\tau = \rho_a(u^*)^2$. The overall shear stress (τ) exerted by the wind over a rough surface is partitioned between stress on the roughness elements and stress on the soil surface [*Marshall*, 1971; *Raupach et al.*, 1993], $\tau = \tau_r + \tau_s$, where τ_r and τ_s are the shear stress acting on the roughness elements and uncovered soil surface, respectively. *Raupach et al.* [1993] showed that this drag partitioning between the roughness elements and substrate surface is controlled by the frontal area of the protruding roughness elements, which is given by the lateral cover (λ, also known as roughness density)

$$\lambda = \frac{nbh}{s}, \qquad (2)$$

where n is the number of roughness elements, b is their mean width, s is the area of the ground with n roughness elements and h is their mean height; λ is dimensionless.

[67] An alternative but equivalent formulation for lateral cover was given by *Okin* [2008] as $l = NA_p$, where N is the number density of plants (or other nonerodible roughness elements), and A_p is the average profile area of the roughness elements. When nonerodible roughness elements are present the shear stress, τ_s, acting on the erodible soil surface is less than the shear stress corresponding to the threshold friction velocity without nonerodible roughness elements. Thus, the threshold velocities observed when roughness elements are present are higher (i.e., the soil is less erodible) than in unprotected soils.

[68] *Raupach* [1992] developed a theory to express the ratio of overall shear stress (τ) to shear stress on the uncovered surface (τ_s) as a function of roughness density:

$$\frac{\tau}{\tau_s} = \frac{1}{(1 - m\sigma\lambda)}\frac{1}{(1 + m\beta\lambda)}, \qquad (3)$$

where σ is the basal to frontal area ratio of the roughness elements, $\beta = \frac{C_R}{C_S}$ and C_R and C_S are the drag coefficients of an individual roughness element and of a smooth surface, respectively, while m is a dimensionless empirical parameter that characterizes the difference between the average substrate surface stress and the maximum stress at any point. The ratio between the threshold friction velocity of an erodible surface without roughness to that of a surface with nonerodible roughness present can be calculated as $R_t = (\tau/\tau_s)^{1/2}$. To overcome the difficulties in determining the frontal area in field situations where there is a random distribution of both nonerodible aggregates and surface heights, *Marticorena and Bergametti* [1995] developed a parameterization of the threshold friction velocity in a rough situation as a function of diameter of erodible particles and aerodynamic surface roughness (roughness length).

[69] More recently, *Okin* [2008] developed an alternative model to describe the surface shear stress partitioning on vegetated landscapes. Arguing that lateral cover was both scale-dependent and difficult to measure in the field, this model calculates the horizontal aeolian flux for every point

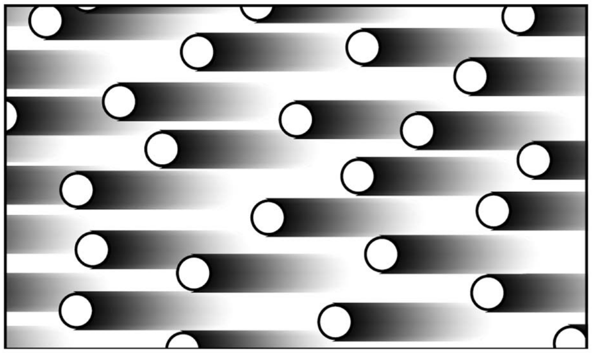


Figure 20. Schematic showing reduction of shear velocity (dark areas) downwind of plants (circles) in the *Okin* [2008] model of wind erosion in vegetated areas.

at regular distances downwind of vegetation (Figure 20). It then uses the probability distribution of those distances, which can be derived from a simple gap intercept method [*Herrick et al.*, 2005], to determine the flux from the landscape as a whole. This method focuses on the size distribution of unvegetated gaps to characterize nonerodible elements, rather than lateral cover, and reproduces both field and laboratory measurements of shear stress ratio. It also explains observed flux measurements at relatively high levels of vegetation cover and explains the wide range of values for the *m* parameter that appears in *Raupach et al.*'s [1993] model.

4.4. The Effect of Soil Physical and Biological Crusts

[70] Soil physical and biological crusts also affect aeolian processes. The erodibility of a soil surface is a difficult property to quantify [*Geeves et al.*, 2000]. It depends on a variety of interrelated soil textural, mineralogical, chemical, hydrological, and biological characteristics that are constantly varying in space and time. *Shao et al.* [1996] suggest that it is the inability to quantify the evolution of surface soil conditions during erosion events that constrains the effectiveness of contemporary wind erosion models. It follows that better understanding of the evolution of soil cohesion and roughness will lead to improvements in our ability to predict aeolian sediment transport [*Sokolik and Toon*, 1996; *Shao and Leslie*, 1997; *Chappell et al.*, 2003]. In drylands this will require erosion models to incorporate cycles of the development and degradation of physical and biological soil crusts.

[71] Soils with crusted surfaces without mobile aggregates are generally stable and have lower wind erodibilities than similar uncrusted soils, except under extreme winds [*Marticorena et al.*, 1997]. Crust formation reduces the availability of mobile soil on the surface, thereby reducing the erodibility of crusted soils. The formation of crusts can be due to structural properties of the soil [*Chepil*, 1951], to salt concentration [*Nickling*, 1984], or to microbial activity [*Belnap and Lange*, 2003]. The formation and stability of physical crusts and clods are related to the clay content [*Skidmore and Layton*, 1992]. *Chepil* [1953] observed that increasing the soil clay content leads to the formation of bigger and more stable soil aggregates, which reduce the susceptibility of the soil to wind erosion. When the strength of the crust is augmented by an increase in the proportion of fine materials (clay), the amount of abrasion by saltating grains decreases [*Rice and McEwan*, 2001]. In the case of clayey soils after a rainfall and subsequent drying, crust formation with high cohesion between soil particles can protect the soil surface from the erosive action of the wind [*Chepil*, 1953]. However, *Zobeck and Onstad* [1987] observed that the reduction in roughness caused by rainfall or irrigation on sandy soils often leaves them more erodible than in their aggregated prerain conditions. Wind tunnel tests have shown that presence on the surface of soluble salts like $MgCl_2$ and $CaCl_2$, even in relatively low concentrations, can significantly increase the threshold friction velocity [*Nickling*, 1984]. These soluble salts increase the threshold velocity by cementation of the soil particles near the surface to form a resistant surface crust.

[72] Biological soil crusts are a common feature of many dryland soils. They form from the association of soil particles and organic matter with varying proportions of cyanobacteria, algae, lichens, and mosses. Discontinuous vascular plant cover in dryland environments means that bare soil patches form an important component of the landscape. Consequently, soil properties are just as, if not more, important than vegetation cover in determining the likelihood of sediment transport. Numerous studies demonstrate the role of biological soil crusts in providing stability to soil surfaces in drylands [e.g., *Tsoar and Møller*, 1986; *Belnap and Gillette*, 1997, 1998; *Leys and Eldridge*, 1998; *Belnap*, 2003]. Such is the importance of crust organisms to soil stability that *Viles* [2008], in a recent review of the role of crusts in dryland landscapes, describes them as "ecosystem engineers" because of their ability to improve their own habitat by modifying the rate of physical processes [*Jones et al.*, 1994].

[73] There are two principal ways in which biological soil crusts improve soil stability. First, extracellular bacterial excretions (EPS), such as mucilage and polysaccharides, act as cementing agents and create microaggregates from individual grains (Figure 21a). Second, bacterial filaments and fungal hyphae entangle individual grains and microaggregates to create a cohesive crust (Figure 21b) [*Belnap and Gillette*, 1997, 1998; *Langston and McKenna Neuman*, 2005; *Xie et al.*, 2007]. *Bowker et al.* [2008] provide further details on the crusting process, describing how cyanobacteria aggregate soil grains both chemically and physically. Extracellular secretions have charged surfaces that bond with clay minerals, making it particularly effective at forming microaggregates, whereas bacterial filaments create macroaggregates through physical entanglement. The surface area of sheaths and the total length of the filaments are therefore key properties affecting crusting. As neither is directly related to an assay of total EPS content, *Bowker et al.* [2008] found this to be a poor predictor of soil stability compared to chlorophyll a.

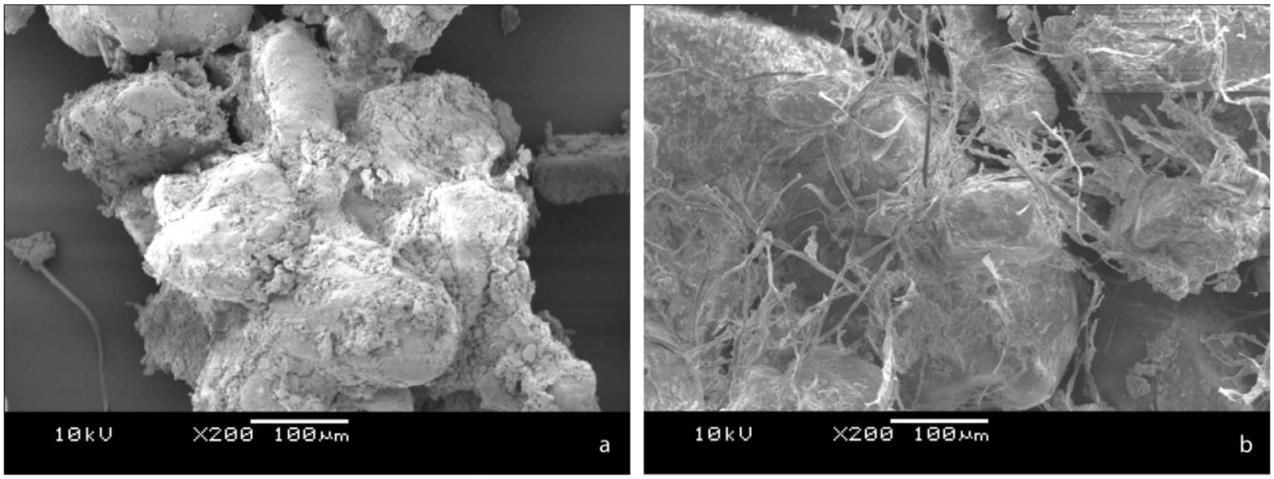


Figure 21. (a) Amorphous mucilage from a biological crust on Kalahari Sands forms microaggregates of individual sand grains [from *Thomas and Dougill*, 2007]. (b) Cyanobacterial filaments from a biological crust on Kalahari Sands entangle sand grains and microaggregates.

[74] The ability of biological crusts to decrease the erodibility of soil surfaces has been shown to vary with species composition as well as the amount and type of extracellular secretions [*McKenna Neuman et al.*, 1996]. For example, crusts containing lichens and mosses were found to be better than cyanobacteria crusts at reducing erosion in Utah [*Belnap and Gillette*, 1997]. *Thomas and Dougill* [2007] also found the compressive strength of crusts in the Kalahari to vary with crust type, with better-developed (and older) crusts having significantly higher strengths than less well-developed (younger) crusts.

[75] The degree to which crusts consolidate soil surfaces affects how much they increase threshold shear velocities (u_t^*) of wind needed to initiate sediment transport [*Belnap*, 2003]. *Eldridge and Leys* [2003] found that biological crusts in Australia significantly reduced the occurrence of wind erosion events when vegetation cover declined during drought years. They found a strong log linear relationship between biological crust cover and sediment transport with crusts, explaining 66% of the variation in sediment transport. Similarly, *Belnap and Gillette* [1998] found the threshold friction velocity for sediment transport was related to biological crust development at the Jornada Experimental Range in New Mexico. Data from a portable wind tunnel show that the u_t^* of unconsolidated soils was significantly lower (30 cm s^{-1}) than soils with a thin cyanobacterial crust cover (40–82 cm s^{-1}). Sandy soils with thicker cyanobacterial crusts had average u_t^* of 260 cm s^{-1}. The highest threshold friction velocities were found on well-developed lichen crusts on silt and sand soils with u_t^* from 323–471 cm s^{-1} [*Belnap and Gillette*, 1998].

[76] Consolidation of the soil surface by biological crusts can however be ephemeral as the active biomass is concentrated in a relatively thin and fragile surface layer, making it vulnerable to a range of disturbances. This is particularly marked when crusts are dry as they become brittle and easily broken by compressional impacts by humans, vehicles, and animals [*Belnap and Gillette*, 1998]. Disturbance leads to a loss of species diversity and biomass and a reduction in metabolic activity and surface cover, which can last for several years [*Lalley and Viles*, 2006]. *Belnap* [2003] reports similar concerns and shows how crust damage adversely affects the photosynthetic activity and N fixation capability of crust organisms. Damage significantly reduces C and N inputs, potentially leading to long-term reductions in productivity and degradation.

[77] Even in undisturbed areas crusted surfaces can be degraded during wind erosion events by saltating particles acting as abraders [*McKenna Neuman and Maxwell*, 1999; *Rice et al.*, 1997]. *McKenna Neuman and Maxwell* [2002] examined the breakdown of crusts subject to multiple grain impacts and wind shear stresses. They found that high grain impact velocities were not necessary for crust breakdown but that the location of the grain impacts was more important. If fractures occur at key locations where filaments are holding aggregates together, crust disintegration can be rapid. *Rice et al.* [1996, 1997] modeled crust rupture by particle impact, finding it to be a function of the probability distributions of the kinetic energy of grain impacts and the energy needed to breakdown the crust surface. Erosion could then be predicted from these two distributions. In earlier laboratory-based experiments, *McKenna Neuman and Maxwell* [1999] found that resistance to saltator damage varies with crust type. Fungal crusts had significantly greater resistance to disintegration under grain impact and wind stress than cyanobacterial crusts, largely due to the greater thickness of the former. They found that once pits or grooves form in the crust surface, underlying loose sediment is released and perpetuates the disintegration of the surface.

[78] The maintenance of a stable crusted soil surface, even in undisturbed environments, is therefore a delicate balance between the insidious process of abrasion during sporadic wind events [*McKenna Neuman and Maxwell*, 2002] and

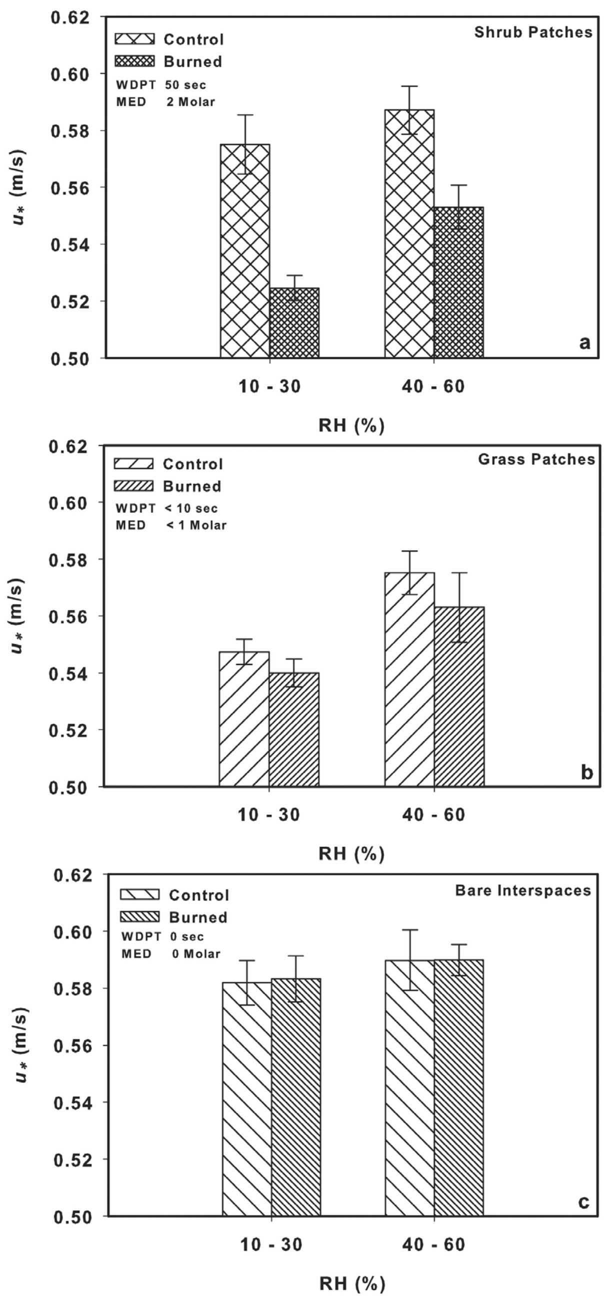


Figure 22. Decrease in threshold shear velocity after fires in soils from (a) shrub patches, (b) grass patches, and (c) bare interspaces [from *Ravi et al.*, 2007b]. These wind tunnel experiments were conducted at two different ranges of air humidity to account for the dependence of threshold velocity of air-dry soils on humidity. The effect of fire on soil erodibility was stronger in areas affected by the burning of shrub biomass, where the emergence of soil water repellency was found to be stronger.

the synthesis of new extracellular material and microbial biomass that occurs during windows of optimal light, moisture, and temperature. Although many studies report the rapid recovery of crusts following disturbance [e.g., *Thomas and Dougill*, 2007], the full recovery of a "mature," highly wind erosion resistant crust with lichens and mosses can take decades [*Belnap*, 1995]. At any rate, recovery (measured as the metabolic activity of crust organisms) is restricted by hydration status [*Lange et al.*, 1998; *Zaady et al.*, 2000; *Conant et al.*, 2004; *Thomas et al.*, 2008], and therefore even relatively small changes to the climate in drylands could tip the balance toward sustained growth or complete destruction of the crust, with significant implications for wind erosion and dust production.

4.5. The Effect of Disturbances

[79] Disturbance is a key driver of aeolian erosion and subsequent dust emissions. Arid and semiarid regions are affected by disturbances like fires and grazing, which may render soils in these regions more susceptible to aeolian erosion with important impacts on crop productivity and regional dust emissions [*Whicker et al.*, 2002; *Ravi et al.*, 2009b; *Neff et al.*, 2008; *Sankey et al.*, 2009a; *Field et al.*, 2011]. Fires are known to affect wind erosion and subsequent dust emission from these landscapes [*Whicker et al.*, 2002; *Breshears et al.*, 2009; *Ravi et al.*, 2009b; *Sankey et al.*, 2009b]. Fires increase the susceptibility to erosive action of wind by destruction of surface vegetation cover and microbial soil crusts. Moreover soil-water repellency induced by fires decreases the threshold velocity needed for occurrence of aeolian erosion [*Ravi et al.*, 2006b, 2009b]. In fact, the burning vegetation releases volatile organic compounds, which induce different levels of water repellency in the surface soil, depending on the vegetation type, fire temperature and soil properties [*Doerr et al.*, 2000]. The hydrophobicity decreases the threshold shear velocity (Figure 22), by decreasing the strength of interparticle wet bonding forces in the soil [*Ravi et al.*, 2006b, 2007b].

[80] Fires have a two-way relationship to erosion processes in that fire frequency and intensity influence the type and distribution of vegetation in a landscape [e.g., *Du Toit et al.*, 2003; *D'Odorico et al.*, 2006b]. Vegetation cover, in turn, affects erosion both directly, by limiting the exposure of the soil surface to the erosive action of winds, and indirectly, through its control on the fire regime, in that both fire intensity and frequency depend on the relative abundance of trees and grasses [*Anderies et al.*, 2002; *van Wilgen et al.*, 2003]. Fire in turn, affects erosion processes by vegetation removal and altering soil properties. This fact is especially important in the case of land degradation caused by the encroachment of shrubs in grassland ecosystems at the desert margins. The encroachment of shrubs into former grasslands is promoted by disruption of the fire cycle [*Van Auken*, 2000] by grazing. Grazing reduces grass cover, and therefore fuel load and connectivity, which reduces the occurrence of fires, which are crucial in killing shrub seedlings. Once fires are reduced in a landscape, shrubs are able to become established and concentrate soil resources beneath their canopies in "resource islands" [*Schlesinger et al.*, 1990]. Vegetation cover is much less, and bare gaps much larger, in these shrublands than in the native grassland, leading to significantly greater wind erosion and dust emission from shrub-encroached former grasslands

with wind-erodible soils [*Gillette and Pitchford*, 2004]. However, at the early stages of this encroachment process, there exists enough grass cover to carry fires from one shrub patch to another. Thus, fire erosion feedbacks during this phase can favor the redistribution of soil resources from the resource islands beneath the shrub canopies to the nutrient-depleted interspaces, thereby promoting the reconversion of the landscape into a state with more uniform distribution of vegetation (grass) and soil properties [*Ravi et al.*, 2009c; *Ravi and D'Odorico*, 2009].

[81] Grazing can enhance aeolian erosion processes in two ways: by removing vegetation and by physical disturbance of the soil surface. The soil microbial crust, which often protects the soil surface in arid and semiarid landscapes, is susceptible to disturbance by grazing and trampling [*Nash et al.*, 2004; *Belnap*, 1995]. Intense grazing is also known to affect the soil microtopography, i.e., variations in surface soil elevation at the scale of 1–2 m, and to affect the erosive action of wind and vegetation patterns [*Nash et al.*, 2004]. Indirectly, grazing impacts wind erosion and dust emission through disturbance of the fire regime and promoting shrub invasion [*Van Auken*, 2000].

[82] Human activities have a profound influence on the activation and enhancement of wind erosion and dust emissions [*Reynolds and Stafford Smith*, 2002; *Neff et al.*, 2008]. Around 70% of the world's drylands are used as grazing lands [*Millennium Ecosystem Assessment* (*MEA*), 2005]. Their overexploitation due to overgrazing and conversion to croplands contributes to the anthropogenic activation of wind erosion and dust emissions in these marginal landscapes. For instance, in Mongolia and China (Inner Mongolia province), the most important anthropogenic factor contributing to land degradation is animal husbandry [*Batjargal*, 1992; *Zhao et al.*, 2005]. The carrying capacities of these grazing systems are increasingly exceeded, resulting in the degradation of vegetation and enhanced soil erosion. Even in the case of monsoon deserts like the Thar in India, which turns lush green following precipitation events, the overexploitation of fodder and fuel wood has caused the ecological destruction of the desert ecosystem resulting in slow rates of natural regeneration of vegetation following precipitation [*Sinha et al.*, 1999; *Ravi and Huxman*, 2009]. Dryland degradation is further enhanced by climatic changes, urbanization, and poor land management as these landscapes are very sensitive to climate change and disturbances.

5. AEOLIAN RESEARCH METHODS

[83] Accurate and reliable methods to quantify wind erosion and the factors that affect wind erosion through time are needed to develop models to predict the impacts on agricultural production, air quality and dust emissions [*Zobeck et al.*, 2003a]. The methods typically used in aeolian and wind erosion studies include laboratory-scale, plot-scale, field-scale, and regional-scale techniques. We summarize approaches relative to laboratory and field techniques, remote sensing, and modeling.

5.1. Laboratory and Field Techniques

[84] Laboratory-scale methods include the use of wind tunnels (recirculating and nonrecirculating types) to estimate threshold shear velocity for wind erosion, to assess the relation between sediment flux and soil surface conditions, to investigate the aerodynamics of winds close to the soil surface, to evaluate the impacts of surface roughness elements, and for many other applications [*Leys et al.*, 1999; *Ravi et al.*, 2006a; *McKenna Neuman and Nickling*, 1989]. A field wind tunnel is often utilized for erosion studies in small plots under field conditions (Figure 23a) [*Gillette*, 1978; *Leys and Raupach*, 1991]. However, there are limitations to the use of wind tunnels, and hence results derived from wind tunnel experiments may differ considerably from field experiments. A wind tunnel may not represent real field conditions because of issues of scale and the inability to accurately reproduce the aerodynamic forces acting on the soil surface.

[85] Other laboratory techniques in wind erosion studies include quantifying physical and chemical soil properties that influence erosion [*Zobeck*, 1991a, 1991b; *Hagen et al.*, 1999a; *Zobeck et al.*, 2003b; *Ravi et al.*, 2006a]. Examples of such properties are soil texture, aggregate stability, the characterization of clay minerals, and the assessment of the presence of hydrophobic compounds.

[86] Laboratory instruments have been proposed to generate dust and evaluate dust emissions, composition, and health impacts from geologic materials. A review of laboratory dust generators has been provided by *Gill et al.* [2006]. For example, the Lubbock dust generation, sampling, and analysis system (LDGASS) was designed to simulate wind erosion of soils by agitating a small sample of soil in a rotating tube and collecting the resulting dust in a settling chamber (Figure 23b). The dust falling in the chamber is collected on filters for further processing and monitored in situ with various real-time dust particle size monitoring systems. The dust generator has been used to relate soil properties to dust emissions in west Texas [*Amante-Orozco and Zobeck*, 2002] and other areas in the United States, to measure particulate matter of size 10 μm or less (PM_{10}) and of size 2.5 μm or less ($PM_{2.5}$) emission potential on sediments from the Aral Sea in Uzbekistan [*Singer et al.*, 2003], to measure and relate the microbial properties of dust and source soil [*Acosta-Martínez and Zobeck*, 2004], and to determine the mineral composition of dust of Bodélé Depression samples from Chad (A. L. O'Donoghue, personal communication, 2009).

[87] Although field and laboratory wind tunnels can provide valuable information on wind erosion, they require considerable resources to build and maintain, are time consuming, and can be expensive to operate. As a result, simpler field and laboratory instruments have been proposed to generate dust and evaluate dust emissions. A new portable field device has been developed to evaluate potential for wind erosion and dust emissions. Known as the Portable In-Situ Wind Erosion Laboratory (PI-SWERL) [*Etyemezian et al.*, 2007], this device uses a 39 cm diameter annular ring

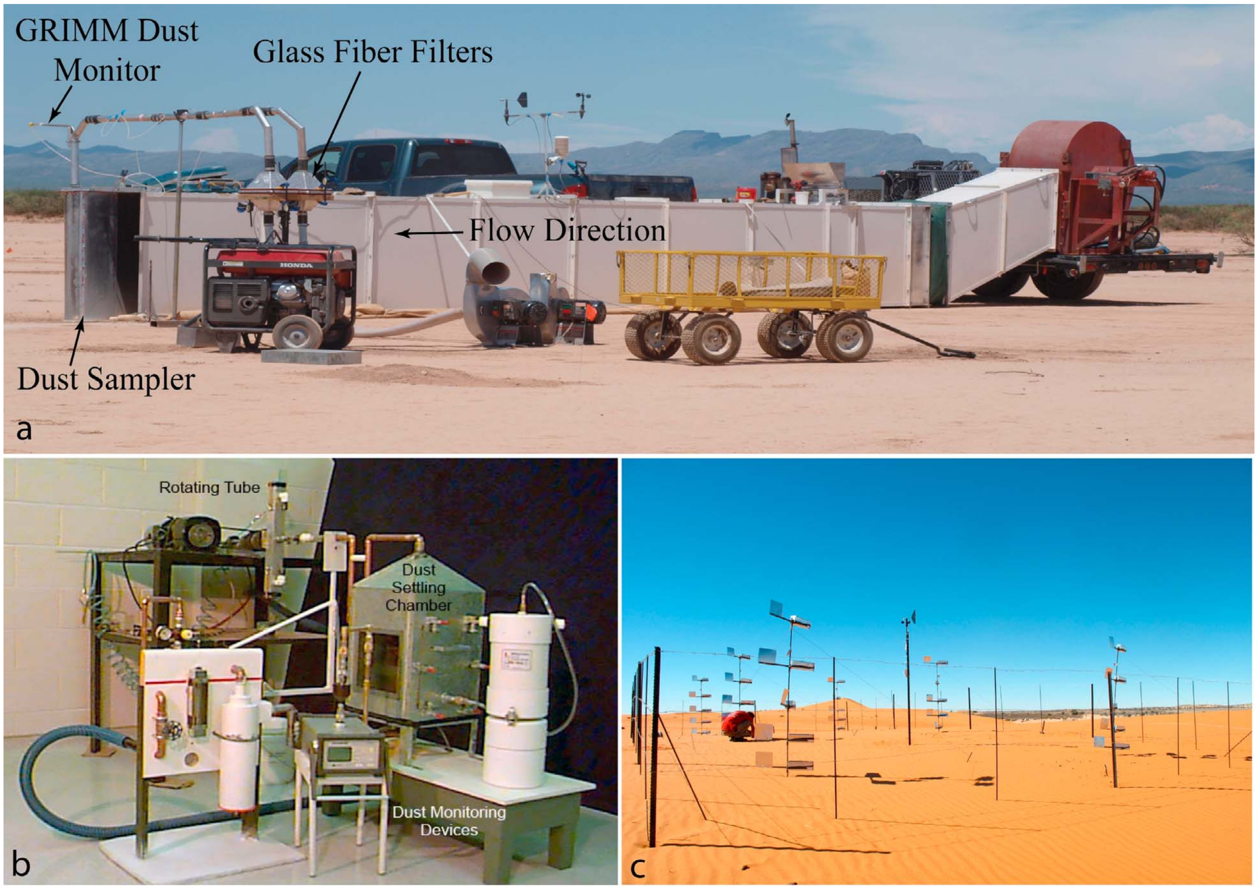


Figure 23. (a) A field wind tunnel, (b) a laboratory dust generator, LDGASS in USDA–Agriculture Research Service Wind Erosion and Water Conservation Research Unit in Lubbock, Texas, and (c) a wind erosion monitoring site (using BSNE dust samplers) in Botswana (courtesy of Abinash Bhattachan).

that rotates 6 cm above the soil surface inside a larger 51 cm diameter ring. Dust and sand are mobilized by the rotating action, and airborne samples are measured and collected within the rings. While the PI-SWERL does not realistically simulate natural wind erosion processes that are often driven by saltation, the measurements are believed to provide a robust index of wind erosion and dust emission potential [*Etyemezian et al.*, 2007].

[88] Field methods include estimating wind erosion as a function of soil eroded from the plots and quantifying the factors that affect wind erosion under field conditions [*Li et al.*, 2007]. The stream wise saltation flux (horizontal dust flux) can be calculated using active or passive dust sampler measurements at different heights. Windblown sediments (sand drift) are collected using sediment traps, such as the commonly used Big Spring Number Eight (BSNE) isokinetic dust samplers [*Fryrear*, 1986b] (Figure 23c). However, at the field scale the majority (more than 90%) of the mass flux occurs very close to the soil surface, and hence samplers rarely need to exceed 1 m in height to determine the horizontal flux. Using a profile of dust concentration, the vertical flux can be derived [*Shao*, 2008]. The saltation activity (i.e., soil movement close to the surface) and the kinetic energy of the saltating soil grains can be measured using wind eroding mass sensors (e.g., sensit, or safire, or saltiphone) [*Stout and Zobeck*, 1997; *Van Pelt et al.*, 2009]. Sediment tracers (e.g., rare earth element tracers, isotopic tracers) can be used for tracking the aeolian sediment sources and sinks in the landscape [*Sutherland et al.*, 1991; *Ping et al.*, 2001; *Li et al.*, 2005]. The climatic factors that affect aeolian processes can be monitored using standard micrometeorological instrumentation typically deployed on meteorological towers. The meteorological data coupled with data from saltation sensors can be used for determining threshold shear velocity for wind erosion [*Stout and Zobeck*, 1997; *Stout*, 2004]. More recently, fast-response instruments to measure wind characteristics (e.g., sonic anemometers) are widely used to measure threshold shear velocity [*van Boxel et al.*, 2004]. These instruments can be deployed in the field along with active or passive dust aerosol samplers to investigate dust movement in the entrainment, transport, and deposition phases over large landscapes [*Whicker et al.*, 2002; *Breshears et al.*, 2003; *Mikami et al.*, 2006].

5.2. Remote Sensing of Dust

[89] Important research questions on the sources, trajectories, and regional and global impacts of atmospheric dust

on climate [*Kaufman et al.*, 2002], human health [*Pope et al.*, 1995; *Griffin et al.*, 2001], and biogeochemical cycles [*Duce and Tindale*, 1991; *Swap et al.*, 1992; *Okin et al.*, 2004] can be addressed with remote sensing data. Whether natural processes alone are responsible for increases in atmospheric dust concentrations or whether land use also has an impact remains unclear. In combination with other data sets, remote sensing data will help address this question.

[90] The remote sensing techniques used to observe atmospheric dust aerosols can be divided into (1) those which work over oceans but not land and (2) those which work over both oceans and land. Multispectral sensors that function in the visible are, for the most part, only valuable for mineral aerosol remote sensing over the oceans. Two VNIR sensors, advanced very high resolution radiometer (AVHRR) and the Sea-viewing Wide Field-of-view Sensor are commonly used to provide information about AOD or AOT (a dimensionless measure of the degree to which aerosols prevent the transmission of light, primarily due to the process of scattering and absorption) over oceans. These instruments can be used to map atmospheric dust distributions because dust, when illuminated by the Sun, scatters a fraction of the solar radiation back to space [*Husar et al.*, 1997]. Over the continents, the methods developed for these instruments do not work because the radiation scattered by dust is mixed with that reflected from the surface. Nonetheless, the combination of radiative transfer models and visible and near-infrared data does allow retrieval of aerosol information over the continents for NASA's MODIS instruments [*Kaufman et al.*, 1997; *Tanré et al.*, 1997; *Remer et al.*, 2005] as well as NASA's Multiangle Imaging Spectroradiometer instrument [*Kahn et al.*, 2005].

[91] Over the oceans, the problem of surface reflectance is minimized and multispectral instruments can be used to retrieve aerosol information. The algorithm used to retrieve dust distributions from AVHRR data uses the backscattered radiance in the 0.63 μm band of the instrument and a radiative transfer model that includes idealized mineral aerosols [*Husar et al.*, 1997]. This work has characterized tropospheric dust over the oceans and has revealed many spatially coherent plumes that can be interpreted in terms of reasonable sources, with the dust plumes coming from the deserts of North Africa and northwestern China being prominent features.

[92] Two methods have been developed to measure atmospheric aerosols using non-VNIR methods. The TOMS instrument has a UV spectrometer designed to provide accurate global estimates of total column ozone. Recent developments have shown that it is also capable of estimating both absorbing and nonabsorbing aerosols. *Herman et al.* [1997] have developed an AI derived from TOMS, which is defined as

$$\mathrm{AI} = -100\left\{\log_{10}\left[(I_{340}/I_{380})_{meas}\right] - \log_{10}\left[(I_{340}/I_{380})_{calc}\right]\right\}, \quad (4)$$

where I_{340} is the radiance 340 nm, I_{380} is the radiance 380 nm, *meas* denotes the measured radiance using the TOMS instrument, and *calc* denotes the radiance calculated using a radiative transfer model that is constructed to give nearly zero AI in the presence of clouds.

[93] *Prospero et al.* [2002] used the TOMS AI index to identify areas that are sources of atmospheric dust on a global scale. By looking at the number of days where the AI value was above a predetermined threshold, *Prospero et al.* [2002] were able to improve our understanding of where, within large desert areas, dust tends to be generated. They suggest that dust emission is a spatially varying process that tends to be concentrated in large basins where there are ample fine-grained sediments to be eroded.

[94] The Infrared Difference Dust Index (IDDI) is derived from images obtained from the Meteosat 10.5 μm to 12.5 μm thermal infrared (TIR) channel [*Chomette et al.*, 1999]. The IDDI is sensitive to the decrease of the thermal infrared radiance due to the presence of dust in the atmosphere during daytime. To compute the IDDI, a time series of geometrically and radiometrically calibrated images are used to create a reference image representing approximately clear and dust-free conditions. Clouds and dust are separated from the surface information by subtracting the calibrated images from the reference image, and cloudy pixels are masked out. The resulting images provide a time series of IDDI values related to atmospheric optical depth. In an application of the IDDI to understand the distribution of dust emission and potential land degradation in North Africa, *Chomette et al.* [1999] combined IDDI values for North Africa and wind speed at 10 m height to determine threshold wind speed (the wind speed above which wind erosion can occur). This parameter is sensitive to both surface texture and vegetation cover [*Marticorena et al.*, 1997]. *Gillette et al.* [1980] have also shown that threshold wind speed is sensitive to disturbance through land use.

[95] A method that uses the same logic as IDDI, namely that dust over continental areas will have similar reflectance but lower temperature than the underlying soil surface, has been developed by *Miller* [2003]. Miller's method provides visual enhancement of atmospheric dust from NASA's MODIS instruments, which have spectral bands in both the VNIR and TIR, although the technique can be applied, in principle, to any system that provides radiometric data in the appropriate spectral bands. The visual enhancement of airborne dust has allowed identification of dust sources in addition to imaging of large dust storms [e.g., *Liu et al.*, 2007; *McGowan and Clark*, 2008; *Miller et al.*, 2006].

5.3. Modeling

[96] Data from wind tunnel experiments, field monitoring, and remote sensing can be inserted in geographical information systems or used in modeling studies to predict wind erosion at various spatial and temporal scales using empirical or process-based approaches [*Leys et al.*, 1999; *Shao and Leslie*, 1997]. This approach enables prediction of erosion rates and identification of geographical areas and time periods that are highly susceptible to aeolian erosion [*Shao and Leslie*, 1997; *Shao et al.*, 2003]. More recently,

an integrated wind erosion modeling scheme for simulation and prediction of all aspects of wind erosion form entrainment, transport and deposition has been proposed [*Shao*, 2008]. This modeling scheme enables quantitative assessment and prediction of wind erosion and dust emissions at local to global scales by coupling atmospheric and land surface models [*Gillette and Hanson*, 1989; *Shao and Leslie*, 1997].

[97] In an effort to help agricultural producers make better and more cost effective soil management decisions, some predictive models for wind erosion in agricultural land have been developed. The first of these models, the Wind Erosion Equation (WEQ) [*Woodruff and Siddoway*, 1965], is still widely used by NRCS to assess producer compliance. WEQ is a highly empirical model that uses assigned values or nomographs (graphs with linear or logarithmic scales depicting a family of curves) of the controlling factors of soil erodibility, soil surface roughness, climate, field length, and vegetative cover to predict periodic soil loss. WEQ has been found to be accurate in the Great Plains of western Kansas and eastern Colorado where it was developed, but tends to underestimate wind erosion in the southern High Plains of Texas and overestimate wind erosion in the northern Great Plains [*Van Pelt and Zobeck*, 2004]. Other empirical and semiempirical models are used in other countries and have been proposed for use in the United States. In a validation exercise comparing the output from two of these models against field measured erosion, the Wind Erosion Stochastic Simulator (WESS) was found to have event magnitude biases in which small events were overestimated and large magnitude events tended to be underestimated, and the Revised Wind Erosion Equation (RWEQ) was found to underestimate soil loss and maximum transport capacity by the wind [*Van Pelt et al.*, 2004].

[98] A more mechanistic approach to wind erosion modeling is the Wind Erosion Prediction System (WEPS) [*Hagen*, 1991]. WEPS can be used to predict wind erosion for discrete periods or for single events in agricultural lands. It is a daily time step, process-based model that predicts wind erosion by simulating and integrating the fundamental factors involved, including wind speeds, soil surface conditions, crop growth, and residue degradation. Using the same data set used to evaluate the performance of WEQ, WESS, and RWEQ, WEPS estimated the event-wise soil loss with reasonable agreement (R^2 = 0.71) for 46 wind events at 6 North American locations [*Hagen*, 2004]. A comparison of measured soil loss in Germany with WEPS output showed excellent agreement (R^2 > 0.9) between measured and predicted soil loss [*Funk et al.*, 2004]. Like WEQ, WEPS was developed primarily for use with cropped ground and has been found to be sensitive to cropping system related factors such as soil surface wetness, dry aggregate stability, oriented roughness, random roughness, and residue management [*Hagen et al.*, 1999b; *Feng and Sharratt*, 2005]. The sensitivity of predictive models to management related parameters is indicative of the profound effects that land management has on wind erosion. Despite all of their successes in agricultural lands, it is critical to understand that these models were not designed and are not appropriate for use in rangeland or nonagricultural lands that produce dust.

6. MANAGEMENT OF WIND EROSION

[99] In an effort to mitigate the numerous negative effects of wind erosion, agricultural management practices have been developed. Management methods that are used to control wind erosion on cropland include planting windbreaks to alter wind flow patterns, retaining plant residue after harvest, stabilizing the surfaces using water or applied chemicals, and tilling the field to bury erodible particles and increase the roughness by increasing the percentage of nonerodible aggregates on the surface and creating bed patterns perpendicular to the predominant winds [*Nordstrom and Hotta*, 2004].

[100] Windbreaks and shelterbelts have been used to decrease the erosive force of the wind in many local settings. They are typically rows of trees and shrubs planted along the margins of the field or farmstead they are intended to protect, but may also be composed of fences, rock walls, or earth berms. Such barriers effectively decrease the wind speed for a distance of about 10–15 times their height downwind and about 3 times their height upwind [*Oke*, 1978]. Because of the limitations of tree growth in many regions, especially semiarid regions, this distance rarely exceeds a few hundred meters downwind [*Vigiak et al.*, 2003]. Windbreaks and shelter belts are not as common as they once were. Although there were approximately 65,000 km of them planted in the Great Plains of North America by the 1960s [*Griffith*, 1976], by the 1970s many were dying or were being removed [*Sorenson and Marotz*, 1977]. Several factors are attributed to the removal of some Great Plains shelter belts, such as maturity or overmaturity of trees, poor matching of tree species with soil and climatic conditions, tree mortality due to insect, disease and drought, and poor management by land owners [*Fewin and Helwig*, 1988].

[101] Maintaining crop residues on the cropped ground is perhaps the most effective management solution for controlling wind erosion. The value of crop residues for controlling wind erosion has been recognized for at least six decades [*Chepil*, 1944]. Residue protects the ground by offering elements that prevent saltating particles from cascading and by increasing the roughness height. WEQ and other predictive models treat all crop residues, whether standing or flat on the ground, as equivalent protection of flat small grain residues [*Woodruff and Siddoway*, 1965; *Bilbro and Fryrear*, 1985]. Standing residues and growing crops provide greater protection than flat residues because they absorb much of the shear stress in the boundary layer [*Skidmore*, 1994]. This displaces the effective roughness height by a zero-plane displacement height, which depends on the height, density, and stiffness of the vegetation [*Oke*, 1978]. The displacement height modifies the wind profile and the shear stress at the ground surface.

[102] The effects of vegetation on soil loss is estimated using the soil loss ratio, an index calculated by dividing the amount of soil loss from a residue-protected soil surface by

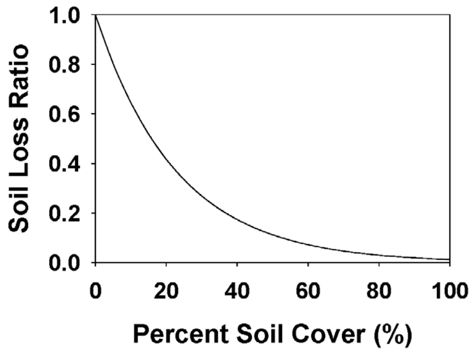


Figure 24. Soil loss ratio as a function of soil cover by nonerodible elements (modified from *Bilbro and Fryrear* [1994]).

the loss from a similar bare surface (Figure 24). The soil loss ratio decreases rapidly from 1.0, for a bare unprotected surface, to a value of approximately 0.2, with 40% soil cover [*Fryrear*, 1985].

[103] Another description of plant canopy or residue used by predictive models for standing vegetation is the plant silhouette through which the wind must pass. *Bilbro and Fryrear* [1994] observed a strong relationship between plant silhouette and the soil loss ratio. However, very sparse residue or other roughness element cover may actually increase soil loss by compressing airflow and creating localized high wind velocities that exceed threshold [*Sterk*, 2000].

[104] Roughening of the soil surface with tillage is an effective way of preventing the cascade of saltation that occurs over flat surfaces [*Fryrear*, 1984]. Roughness may be oriented such as raised beds or random as large clods and other soil aggregates. Oriented roughness is most effective when oriented perpendicular to the prevailing wind (Figure 25) [*Hagen and Armbrust*, 1992; *Saleh*, 1994]. Oriented roughness has little effect when the wind is blowing parallel to the direction of tillage, and random roughness then becomes the dominant protection of the surface. Roughness helps prevent wind erosion by increasing the aerodynamic roughness length parameter [*Saleh et al.*, 1997] and by the increasing cumulative shelter angle distribution [*Potter and Zobeck*, 1990], which has been shown to be sensitive to tillage tools, wind direction, and rainfall.

[105] The ability of the soil to maintain a roughened condition depends on several factors, including the presence of nonerodible soil aggregates [*Soil Survey Division Staff*, 1993]. Soil clods are similar to aggregates, but soil forming processes have exerted little or no control on their size. They are produced by external mechanical forces and may be composed of many natural aggregates. Following tillage, the soil surface is typically composed of clods and aggregates with a wide range of diameters. Aggregates > 0.84 mm are generally considered nonerodible [*Chepil*, 1958]. The stability of these aggregates in the dry condition is known as dry aggregate stability and is a measure of the resistance to breakdown from physical forces [*Skidmore and Powers*, 1982]. The physical forces leading to aggregate breakdown may come from tillage or from other mechanical impact such as traffic from vehicles or animals and from the forces of saltating sand grains. The dry aggregate stability of tilled agricultural soils is typically determined by repeated sieving using a rotary sieve [*Chepil*, 1962]. Aggregate clay content and water content at the permanent wilting point (−1.5 MPa matric potential) have been found to be good predictors of dry aggregate stability [*Skidmore and Layton*, 1992].

[106] The resistance to breakdown by the forces of water is termed wet aggregate stability. Natural organic structures including particulate organic matter [*Gale et al.*, 2000] and organic compounds including glomalin [*Wright and Upadhayaya*, 1998; *Wright et al.*, 1999] in the soil are important in wet aggregate stability. Tillage of the soil leads to decreases of these important stabilizing agents in soils [*Fenton et al.*, 1999; *Six et al.*, 1999; *Bronson et al.*, 2004]. The impact force of raindrops, particularly during intense convective rainstorms, is sufficient to disperse the aggregates in soils with poor wet aggregate stability. When this happens, soil roughness is diminished, and a crust may form. The strength of the crust and its ability to withstand abrading particles is related to the soil properties and the rainfall rate and impact energy that created the crust [*Zobeck*, 1991a]. A strong crust will armor the surface and help protect it from wind erosion [*Zobeck*, 1991b], but in sandy soils, loose, erodible material (LEM) may be deposited on the crust. This LEM is highly erodible and, if present, may initiate saltation and erode the crust during wind events. The amount of LEM left on the surface is a function of soil texture, microtopography of the surface, and rainfall [*Potter*, 1990]. Postrainfall tillage is often used to bury the LEM and reroughen the surface.

[107] The best management practice (BMP) to prevent erosion is to limit the contact of wind with the soil surface by maintaining an effective cover of residue such as a cover crop or carefully managed stubble. The emergence of no-till and conservation tillage practices have resulted in more effective postharvest standing and flat residue over cropped ground. Advances in harvest equipment such as finger headers on small grain combines have also led to improvements in the postharvest heights of standing residue. In semiarid regions that represent marginal dryland farming

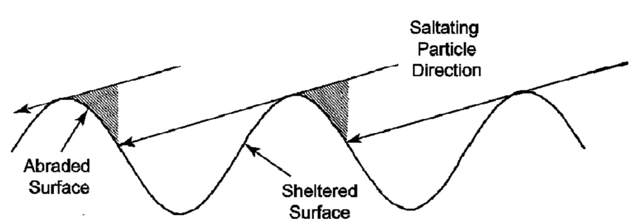


Figure 25. Schematic representation of a ridged field with windblown saltating particles traveling at a nearly perpendicular direction. The shaded areas are susceptible to erosion, while the much greater unshaded surface is not [from *Zobeck*, 1991a].

regions and with certain locally important crops such as cotton or sunflowers, insufficient silhouette or flat residue may fail to protect the soil.

[108] On bare soils or soils with limited crop residues, the tillage technique remains the predominant BMP to prevent erosion. Raising beds perpendicular to the prevailing wind direction increases the aerodynamic roughness and provides regularly spaced roughness elements offering sheltered areas to prevent cascading saltation. By creating a surface dominated by nonerodible aggregates, a random roughness is formed that offers the same protective sheltered areas to prevent cascading saltation. In fragile soils with low dry aggregate stability, erosion may start in localized areas of the field or at the downwind end of a long, frequently traveled, unpaved road. In such locations, it may be necessary to use a snow fence or other barrier to encourage deposition and discourage saltation. Intense rainfall on soils with low wet aggregate stability often results in a smooth crusted soil surface with loose sand-sized material on the surface. The use of crust breaking and clod forming tillage implements such as a rotary hoe or a sand fighter is often used after spring thunderstorms to create random roughness to the field surface. Once the crop is established and the canopy covers a significant portion of the soil, tillage is only used to control weeds. In management of arid and semiarid rangelands, monitoring of vegetation canopy-based indicators (e.g., bare patch index) and soil surface characteristics (e.g., soil texture, strength of physical and biological crusts) can be used as indicators of wind erosion susceptibility and early warning signs of land degradation [*Whitford et al.*, 1998; *Herrick*, 2000; *Herrick et al.*, 2005].

7. CONCLUDING INSIGHTS AND RESEARCH PRIORITIES

[109] This review highlights the many ways in which aeolian processes play a major role in the biosphere. There is a growing interest in the scientific community to understand and to model these processes, as indicated by the increasing trend in aeolian publications in the past few decades [*Stout et al.*, 2009]. Aeolian research has shifted from small-scale investigations on sand movement to also include the analysis of continental-scale transport of dust and its impact on biogeochemical cycles, climate, and human health; it has also seen the introduction of novel field and computational research methods and the continuing emergence of interdisciplinary approaches for understanding aeolian processes [*Shao*, 2008].

[110] What are some the key insights to take away from this review? Regarding the geography of dust emissions (section 2), our review has highlighted that dust emissions generate a substantial amount of connectivity across the globe, with areas of hyperaridity being important sources and areas of the Sahara, Middle East, and China being hot spots of particular concern.

[111] Regarding impacts, interactions, and feedbacks for aeolian processes (section 3), there are six key insights, corresponding to the review subsections. First, the global connectivity in dust emissions, identified regarding the geography of dust emissions, has direct implications for connectivity of global biogeochemical cycles related to aerosol impacts and source deposition patterns. Second, there are important climate-mediated interactions between dust and ecosystems stemming from the effects of dust on light absorption and scattering and on cloud condensation nuclei, including a hypothesized feedback linking dry soils, increased dust emissions, and reduced precipitation that may be applicable to the West African Sahel region and the Dust Bowl event. Third, dust emissions directly affects human health via risk from dust inhalation associated with pathogenic microorganisms, toxic chemicals, and radionuclides; disturbances to the land surface that increase these types of dust emissions also increase risks to human health. Fourth, aeolian effects on agriculture from reductions in soil cover and from altering the soil surface and are due to either changes in the soil resource, through soil loss or redistribution, or by mechanical injury (sandblasting). Fifth, dryland systems inherently have patchy vegetation, and redistribution of sediment by aeolian processes is not only affected by this vegetation patchiness and the associated connectivity of bare areas, but it also influences vegetation dynamics. Sixth, aeolian processes interact with and can be competing with fluvial processes, although most research has focused on only one of these two major drivers but not both.

[112] Regarding drivers of dust emissions and their controls (section 4), processes are complex and require consideration of fine to coarse particle sizes and transport and erosion via suspension, saltation, and creep. The threshold friction velocity that is driving aeolian processes is itself determined by five other factors in addition to climate. First, with respect to particle size, threshold friction velocity depends on particle diameter and density as well as the density of the air. Second, soil moisture is a critical factor controlling dust emissions over shorter time scales (e.g., days) due to its effects on interparticle forces, whether through moisture on particle surfaces or "bridging" between them; atmospheric humidity is also an important determinant of emissions, particularly under dry conditions. Third, vegetation and surface roughness affect emissions by sheltering the surface and absorbing a fraction of the wind momentum flux. Shear stress, related to shear velocity, is split between stress on the soil surface and stress on roughness elements, the later of which depends on lateral cover (frontal area) and/or the size distribution of unvegetated gaps. Fourth, soil physical and biological crusts are prevalent in many dryland ecosystems and lower wind erodibility (except for under extreme winds). Biological soil crusts, comprising soil particles, organic matter, cyanobacteria, algae, lichen, and mosses, act as cementing agents and can entangle particles; the effects of these crusts can be ephemeral and are sensitive to disturbance. Fifth, disturbances to the soil surface generally increase dust emissions, with grazing and fire being two important types of disturbance for dryland ecosystems: grazing reduces vegetation cover and disturbs the soil surface, while fire additionally generates

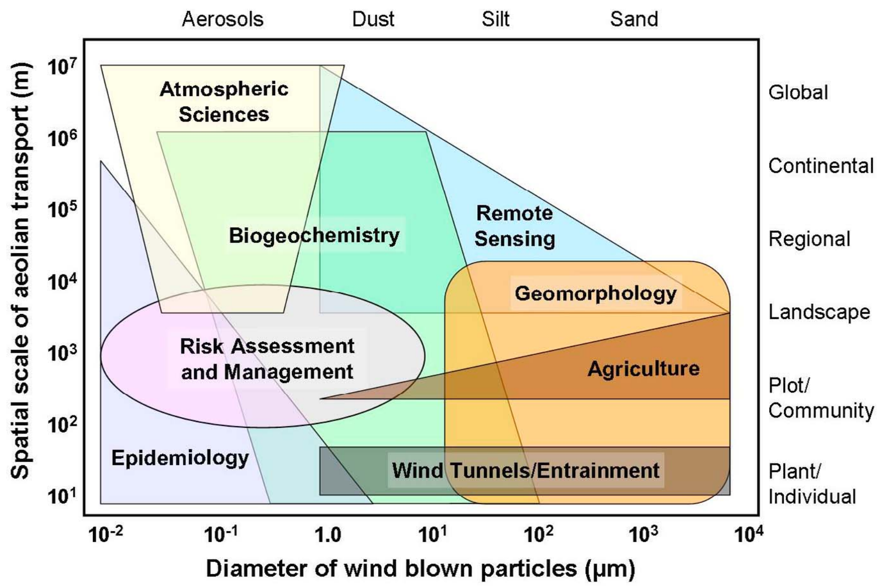


Figure 26. Conceptual representation of the span of aeolian research in various disciplinary areas differentiated with respect to spatial scale and portion of the particle size spectrum.

feedbacks between aeolian processes and vegetation. In addition to the land surface-generated, human-driven disturbances, climate extremes, especially severe drought, is also an important disturbance.

[113] Regarding aeolian research methods (section 5), approaches include laboratory and field methods, remote sensing, and modeling. Laboratory and field methods employ wind tunnels and dust generators, which provide controlled conditions but are limited in scale, in vegetation complexity, and/or with respect to aerodynamic forces. Field studies focus on wind erosion and associated transport as a function of soil eroded from plots and factors affecting that erosion; measurements of various soil properties are also highly relevant for studying aeolian processes. Remote sensing is particularly useful for assessing dust emissions sources and trajectories at regional to global scales; approaches can be differentiated between those for over ocean only (VNIR) and those for over both ocean and land (TOMS, IDDI). Models of aeolian processes have been developed for a variety of spatial and temporal scales, using both empirical and process-based approaches; the most widely used models have been focused on agricultural fields (WEQ, WEPS).

[114] Regarding management of wind erosion (section 6), for croplands approaches include windbreaks, retaining plant residue after harvest, stabilizing surfaces with water or chemicals, and tilling a field to bury erodible particles and to increase the roughness. For arid and semiarid rangelands, monitoring focused on vegetation-based indicators and soil surface characteristics provides indications of wind erosion susceptibility and early warning signs of land degradation.

[115] Collectively, aeolian research relevant to the biosphere conducted to date can be viewed as spanning several disciplinary areas that can largely, but not wholly, be differentiated with respect to spatial scale and portion of the particle size spectrum emphasized (Figure 26; the indicated boundaries are only approximate and numerous exceptions exist; typical temporal scales could also be associated with each of the boundaries). One key insight that emerges for this review is that the existing research to date within different disciplinary boundaries collectively fills most of the possible areas on spatial scale by particle size distribution parameter space (Figure 26; the void in upper right corner of the parameter space for spatial scale by particle size distribution combinations are not relevant, e.g., sand is not generally redistributed at global scales). This perspective highlights that in addition to individual research priorities, each disciplinary area should consider opportunities to glean relevant information from other related disciplines.

[116] Many key research and management challenges emerge regarding aeolian processes in the biosphere that lie within or cut across the disciplinary boundaries. Among these challenges are (1) lack of reliable, long-term erosion-dust emission data sets from different ecosystems; (2) difficulty in accurately determining some key parameters affecting wind erosion, particularly soil moisture and roughness elements; (3) lack of understanding and information about interactions between aeolian and fluvial processes; (4) lack of accurate, reliable, and direct measurement of the hydroclimatic factors affecting wind erosion and dust emissions that are needed to both validate erosion models and to assess the intensity of aeolian processes in dryland environments; (5) limitations associated with parameterization of complex aeolian process models and methods to determine the uncertainty associated with model predictions; (6) development of dust emission models that can be effectively used to support land management decisions and incorporated in regional or global climate models; (7) impacts of climatic changes and disturbances on the

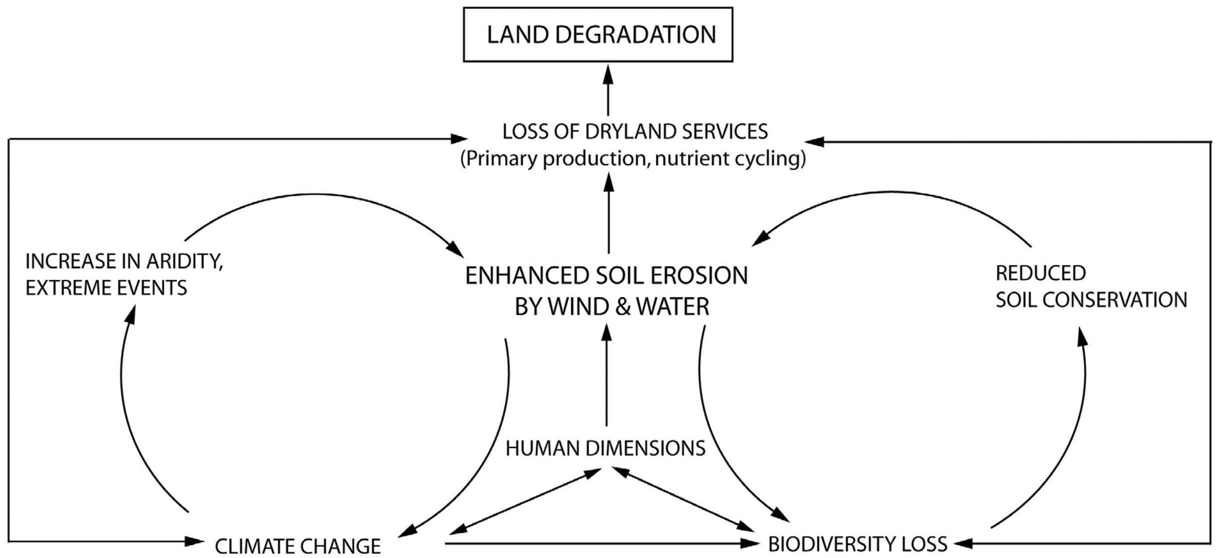


Figure 27. Conceptual diagram showing the interrelations among soil erosion, land degradation, climate change, and biodiversity loss (modified from *MEA* [2005] and *Ravi et al.* [2010]).

susceptibility of the landscape to aeolian erosion, which will require obtaining more accurate and reliable measurements of aeolian erosion and deposition across a wide range of spatial scales from individual plants up to regional and global scales; (8) uncertainty regarding the global distribution of sediment and dust emission and the concentration and mineralogy of atmospheric aerosols as well as their impacts on human health and climate; (9) uncertainty regarding how disturbances such as fire and grazing affect interactions and feedback loops between aeolian processes and the biosphere; and (10) needed methodologies and models to address the impacts of wind- and water-driven erosion collectively and of a more integrated perspective of aeolian-hydrologic dynamics [*Field et al.*, 2009; *Ravi et al.*, 2010]. (We note that research investments made in aeolian research, in general, have been considerably smaller than those for fluvial research, at least apparently in the United States, and therefore perhaps have not been in proportion to the relative importance of aeolian transport processes [*Field et al.*, 2009].)

[117] Overarching these research needs and management challenges is the need to address how changing climate will compound already pressing issues related to land use. Global climate models have predicted an increase in aridity in dryland systems around the world [*Burke et al.*, 2006; *Seager et al.*, 2007], which could increase the dominance of abiotic controls of land degradation, including the susceptibility of the landscape to aeolian erosion. Increase in aeolian erosion rates due to climatic changes (increase in aridity) results in enhanced loss of soil resources and subsequent loss of vegetation cover. The loss of vegetation cover results in the loss of vital ecosystem services, which may include primary production and carbon sequestration [e.g., *Chapin et al.*, 1997], with impacts on regional climate and desertification (Figure 27). Climatic changes and disturbances, on the other hand, can alter vegetation patterns with impacts on aeolian transport processes. How is vegetation cover expected to vary under regional and global environmental change scenarios? How is the rate of soil loss going to be altered by land cover and land use change? Will new major sources of atmospheric dust be activated as it recently happened in the case of the Aral Sea? What are the local and large-scale impacts of these new dust emissions? These are all important questions in the environmental change debate that require a better understanding of the interaction of aeolian processes with vegetation dynamics.

[118] As evident from the sections above, wind erosion and transport and associated dust emission processes affect almost all aspects of the biosphere, including global biogeochemical cycles, climate, human health, agricultural production, and land degradation. Aeolian processes have important consequences for land degradation in arid and semiarid regions, which is one of the major global issues of the 21st century because of its impact on world food security and environmental quality. Increased wind erosion activity in drylands is considered to be a major factor contributing to land degradation, having socioeconomic and political implications. According to the United Nations Convention to Combat Desertification around 3600 million ha, or 70%, of the world's dry lands are degraded. Wind erosion is often considered both as a cause and effect of desertification, and in combination with water erosion it is thought be responsible for more than 80% of the degraded land [*Middleton and Thomas*, 1997; *Lal*, 2001]. Climate change, grazing pressure, and the lack of adequate soil conservation practices may render dryland regions (around 40% of Earth's terrestrial surface with more than 2 billion inhabitants) more susceptible to wind erosion [*Nicholson*, 2000]. Even though areas affected by desertification range from one third to one half of the world's land surface and affects one sixth to one

third of the world's population, the problem of desertification remains poorly understood [*Reynolds and Stafford Smith*, 2002; *MEA*, 2005]. Thus, the understanding of the biophysical drivers and biogeochemical implications of aeolian processes is motivated by the increasing need to estimate long-term changes in soil productivity, assessing the economic cost of soil erosion and desertification, evaluating land conservation and reclamation programs, analyzing the effect of climate change scenarios on soil erosion and dust cycles, and assessing the rate of suspension of dust entrainment into the atmosphere, its contribution to tropospheric aerosols, and deposition to downwind ecosystems.

[119] In conclusion, the view emerging from this review is that the impacts of aeolian processes on the biosphere are more significant than previously thought, with many recent studies focusing on the role of aeolian processes and dust on ecosystem processes. Multidisciplinary approaches and perspectives are urgently needed to advance the science in order to address complex environmental issues such as climate change and desertification, which will be fundamentally impacted by aeolian processes.

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