

Modelling Land Susceptibility to Wind Erosion in Western Queensland, Australia

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Abstract

Wind erosion is a land degradation process with global consequences. Understanding spatial and temporal patterns in land susceptibility to wind erosion is essential for integrating wind erosion research across scales and enhancing management strategies to control potential land degradation. There are significant gaps in our knowledge of which areas of the Earth's surface are susceptible to wind erosion, and how the erodibility of land changes through time in response to climate variability and land management pressures. This stems from a lack of research into spatial and temporal patterns of land erodibility, particularly at the landscape to regional scales (10^3 to 10^4 km²). This thesis addresses these knowledge gaps by presenting research into the development and application of a model to assess land susceptibility to wind erosion in the rangelands of western Queensland, Australia.

The foundation for a new Australian Land Erodibility Model (AUSLEM) is established through a systems analysis of the factors controlling wind erosion, and a review of approaches for representing land erodibility in wind erosion modelling systems. The thesis explores how meteorological, soil and vegetation conditions affect the susceptibility of land to wind erosion, synthesising the analysis in a conceptual model of the land erodibility continuum. The conceptual model provides the basis for a review of wind erosion models that are applicable from the paddock (10^3 m²) to regional (10^4 km²) and global scales. Current limitations to modelling soil and land erodibility are evaluated and the thesis identifies research priorities for developing new models to predict land susceptibility to wind erosion.

The lack of robust schemes to model temporal changes in soil erodibility adversely affects the performance of wind erosion models. To address this issue, a framework was developed for modelling temporal variations in soil erodibility. The framework is an approach for assessing temporal responses of soils to short-term (event-based) variations in climate and land management conditions. Application of the model was restricted by a lack of quantitative data to parameterise the model functions. To address the issue, the thesis presents an analysis of the current status of soil erodibility research and outlines priorities for future research into soil erodibility dynamics.

AUSLEM was developed as a Geographic Information System tool to assess land susceptibility to wind erosion across western Queensland, Australia. The model operates at a 5 x 5 km spatial resolution on a daily time step, using inputs of grass and tree cover, soil moisture, soil texture and surficial stone cover. Model performance was evaluated by comparing trends in the model output with trends in wind speeds and observational records of dust events recorded at eight meteorological stations in western Queensland between 1980 and 1990. The validation was conducted at four spatial length scales, from 25 to 150 km. Results show that AUSLEM performs well in the arid southern and western regions of the study area. Poor model performance at scales $<10^2$ km² is linked to the lack of a scheme to predict temporal variations in soil erodibility.

The approach for validating AUSLEM was shown to be highly dependent on the types of dust events used for the comparison, calling for additional methods for assessing model accuracy. This was addressed through the development of a method for visually assessing land erodibility at the landscape scale using long distance (10^3 km) vehicle-based transects. Visual assessments of land erodibility in western Queensland were collected and then used in a secondary validation of AUSLEM. The results confirm that the model performs well in the west of the study area in the open rangelands. In regions with heterogeneous woody vegetation cover the model performance decreases, demonstrating the importance of considering the spatial distribution of vegetation when monitoring and modelling land erodibility to wind.

AUSLEM was applied to assess spatial and temporal patterns in land erodibility in western Queensland between 1980 and 2006. Highly erodible land is shown to occur in the Strzelecki Desert dunefields and outer floodplains of the Diamantina, Georgina and Bulloo river systems in the Channel Country and Mulga Lands bioregions. The Mitchell Grass Downs bioregion had consistently low land erodibility for the analysis period. Results show a weak seasonal pattern in land erodibility across the study area, peaking between early spring (SON) and late summer (JFM), and reaching a minimum over winter (JJA). At inter-annual time scales, weak correlations are identified between modelled land erodibility, regional rainfall, the El Niño/Southern Oscillation (ENSO) and the Pacific (inter-) Decadal Oscillation (PDO). The results indicate a complex landscape response to climate variability, with land erodibility dynamics being driven by antecedent

rainfall and vegetation conditions which generate lag responses in land erodibility change.

The thesis makes a significant contribution to our understanding of the extent of erodible land in Australia and the sensitivity of the western Queensland rangelands to climate variability. In an international context, the research provides new methods for assessing spatial and temporal patterns in soil and land erodibility, and demonstrates the role of global climate variability in determining important regional scale changes in land susceptibility to wind erosion. Requirements have been highlighted for pursuing research into methods to monitor and model wind erosion processes at the landscape scale, particularly in light of increasing pressures on dryland environments from agricultural activities and future climate change.

Keywords: wind erosion, erodibility, dust, rangelands, model, assessment, Australia

Declaration by the Author

This thesis is composed of my original work, and contains no material previously published or written by another person except where due reference has been made in the text. I have clearly stated the contribution by others to jointly-authored works that I have included in my thesis.

I have clearly stated the contribution of others to my thesis as a whole, including statistical assistance, survey design, data analysis, significant technical procedures, professional editorial advice, and any other original research work used or reported in my thesis. The content of my thesis is the result of work I have carried out since the commencement of my research higher degree candidature and does not include a substantial part of work that has been submitted to qualify for the award of any other degree or diploma in any university or other tertiary institution. I have clearly stated which parts of my thesis, if any, have been submitted to qualify for another award.

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Nicholas P. Webb

August 2008

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N.P. Webb	Research design, implementation and writing and editing
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* Modifications have been made to the manuscripts for integration into the thesis. This includes minor reformatting, reducing or removing the study area descriptions to eliminate duplication from Chapter 1, and providing links between chapters.

Conference Presentations by the Author Relevant to the Thesis

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Chapter 1

Introduction

1.1 Introduction to Thesis

Approximately 47% of the world's land surface is classified as drylands (Thomas, 2000). This area includes dry sub-humid, semi-arid, arid and hyper-arid regions. The drylands are characterised by low precipitation – high potential evapotranspiration ratios and sparse vegetation cover (UNEP, 2008). These conditions result in a deficiency in soil moisture and large areas of bare ground, making the regions particularly sensitive to climatic changes, anthropogenic disturbance and land degradation.

Wind erosion is a land degradation process that affects dryland environments. It affects approximately 28% of the global land area experiencing land degradation (Oldeman, 1994; Callot et al., 2000; Prospero et al., 2002). While wind erosion is a naturally occurring process in many regions, it has been accelerated by human activities in rangelands and in marginal cultivated lands. Anthropogenic pressures accelerating wind erosion include overgrazing of rangeland pastures and the use of long-fallowing of cultivated lands (Leys, 1999). While in many developed countries cultivation practices have been improved to minimise soil loss due to wind erosion, there remains considerable grazing pressure on rangelands. Arid and semi-arid rangelands cover approximately 45% of the world's land surface (Reid et al., 2008). The effects of grazing on the rangelands are seen through a reduction in vegetation cover (due to consumption), and the disturbance of soil crusts which play a critical role in reducing the erodibility of soils in arid and semi-arid environments (Belnap and Eldridge, 2001).

Wind erosion has numerous on- and off-site effects. These are seen from the individual field scale (10^3 m^2) to regional (10^4 km^2) and global scales. On-site effects of wind erosion relevant to agriculture include (after Leys, 1999; McTainsh and Strong, 2007):

- Loss of nutrient rich topsoil in lands with generally nutrient poor soils;
- Selective removal of fine fraction particulates and organic matter from soils;

- The ability of soils to sustain vegetation and livestock decreases;
- Agricultural and pastoral productivity decreases;
- Nutrient enrichment of streams occurs with the influx of wind blown sediments;
- Spread of herbicides and pesticides off-farm; and
- Damage to property (e.g. fences and roads) and farm infrastructure.

Off-site effects of wind erosion are relevant across scales and relate to the transport and deposition of mineral dust which (after Dentener et al., 1996; Prospero et al., 2002; Jickells et al., 2005 and others):

- Alters the radiation balance in the atmosphere through scattering and absorption of radiation;
- Affects cloud nucleation and optical properties of the atmosphere;
- Acts as a reactive mineral species in the atmosphere;
- Moderates the photochemical oxidant cycle and biogeochemical processes;
- Acts as a source of Fe that may be metabolised by cyanobacteria and may subsequently moderate the nitrogen chemistry of the ocean;
- Acts as a source of Fe that may be a limiting nutrient for phytoplankton; and
- Provides reaction sites for ozone and nitrogen molecules.

Understanding spatial and temporal variations in wind erosion is required to develop methods for managing land degradation at the field (10^3 m^2) to landscape (10^3 km^2) scales, and for understanding the regional to global scale consequences of dust transport. Surprisingly little attention has been given to the development of methods for assessing spatio-temporal patterns in land areas susceptible to wind erosion.

Existing maps of the location and extent of regions prone to wind erosion are based on: field assessments of affected areas (Carter, 1985; Mezösi and Szatmári, 1998); observational data on dust-storm frequencies (Goudie and Middleton, 2006); analyses of satellite imagery (Prospero et al., 2002; Washington et al., 2003); and in a few cases spatial modelling (Böhner et al., 2003; Coen et al., 2004). However, these methods for assessing wind erosion have a number of limitations. These include:

1. Field surveys have historically been conducted as individual studies and have only provided snapshots of the landscape condition relevant to the climatic conditions at the time of survey (Leys, 1999).
2. The analysis of dust-storm frequencies relies on the interpolation of data between distant locations, e.g. meteorological stations in drylands are often >100 km apart, and so is generally unable to resolve dust source areas at scales less than $\sim 10^5 \text{ km}^2$ (McTainsh and Pitblado, 1987).
3. The analysis of aerosol optical depth imagery (Prospero et al., 2002; Washington et al., 2003) and development of dust enhancing indices (Legrand et al., 1994; Miller, 2003) has provided a capability to detect point source and regional dust source areas using satellite imagery. However, these methods cannot provide information on land erodibility unless the surface is eroding at the time of image acquisition.
4. While numerous wind erosion models have been developed, few have been applied specifically to assess land susceptibility to wind erosion. Examples of this application have been restricted to cultivated fields and over small geographic extents (e.g. Coen et al., 2004), or very coarse spatial resolutions (Ginoux et al., 2001).

There remains a great deal of speculation about the location and strengths (erodibility) of wind erosion prone areas (Grini et al., 2005). This is a global issue, relevant to our understanding of the world's major dust producing regions in Africa, the Middle East, Asia, North and South America, and Australia (Hermann et al., 1999). Quantitatively describing spatial and temporal patterns in land erodibility is essential for enhancing land management strategies to combat land degradation in dryland environments. Understanding land erodibility dynamics at the landscape scale (10^3 km^2) is also required to better understand regional scale dust emission and transport processes. Without a detailed knowledge of how the erodibility of drylands responds to climate variability and land management, or of the feedbacks between local wind erosion activity and regional climate, our capacity to mitigate potential impacts of global climate change on this degradation process will be severely challenged.

This thesis develops a model to assess land susceptibility to wind erosion, i.e. land erodibility, in the rangelands of western Queensland, Australia. This chapter provides a background to the research problem and synthesis of the research requirements. This is followed by a statement of the research aims and objectives, and research approach. The

chapter then describes the study area and concludes with an outline of the thesis structure and chapter content.

1.2 Wind Erosion in Australia

Australia is the world's driest inhabited continent, with 75% of the land mass being classified as arid and semi-arid (Peel et al., 2007). This means that large areas of Australia are potentially susceptible to wind erosion. Aeolian landforms dominate the arid interior of the continent, and wind erosion processes have long been a natural occurrence in arid and semi-arid Australia (Hesse and McTainsh, 2003).

Early assessments of wind erosion in Australia were conducted during periods of high dust storm activity and drought. Table 1.1 provides a chronology of early studies that sought to identify the extent of wind erosion in Australia.

Table 1.1 Chronology of early surveys of land affected by wind erosion in Australia (1900-1990)

Author	Year	District, Region or State Surveyed
Anon	1901	Western New South Wales (NSW) Royal Commission
MacDonald-Holmes	1936	Australia-wide
Ratcliffe	1937	South Australia
Loewe	1943	Australia-wide
Kaleski	1945	Border region of NSW and South Australia (SA)
Skerman	1947	Channel Country of western Queensland (Qld)
Beadle	1948	Border region of NSW and South Australia (SA)
Taylor	1948	Harden-Young district of central-eastern NSW
Teakle	1957	Australia-wide
Stewart	1968	Central and eastern NSW
Rae	1976	Mallee regions around Mildura, southwest NSW
Woods	1984	Australia-wide
Carter	1985	Eastern Australia
McSwain	1986	Mallee and Wimmera areas of Victoria
Slater	1986	Western Darling Downs-Maranoa areas of southern Qld

Approximately half of the early surveys (Table 1.1) were conducted in central and south-western New South Wales (NSW), a region subject to intensive agriculture, drought, and a history of land degradation associated with a feral rabbit population (McKeon et al., 2004). The erosion surveys provided snapshots of the state of the landscape and did not capture spatial patterns in wind erosion at the continental scale (Leys, 1999). Temporal changes in the

distribution of erodible land between drought and non-drought periods also remained unresolved. A climatology of wind erosion activity across Australia was not established until the mid-1980s.

1.2.1 Spatial Patterns in Wind Erosion

Australia has four regions that experience high annual dust-storm frequencies (>2 days yr^{-1} with visibility <1000 m). Maps of these regions were developed from analyses of observational records of blowing dust by Middleton (1984), McTainsh and Pitblado (1987) and subsequent studies by McTainsh and colleagues (Figure 1.1). While the maps do not explicitly indicate the location of land susceptible to wind erosion, they do provide good continental scale summaries of wind erosion activity.

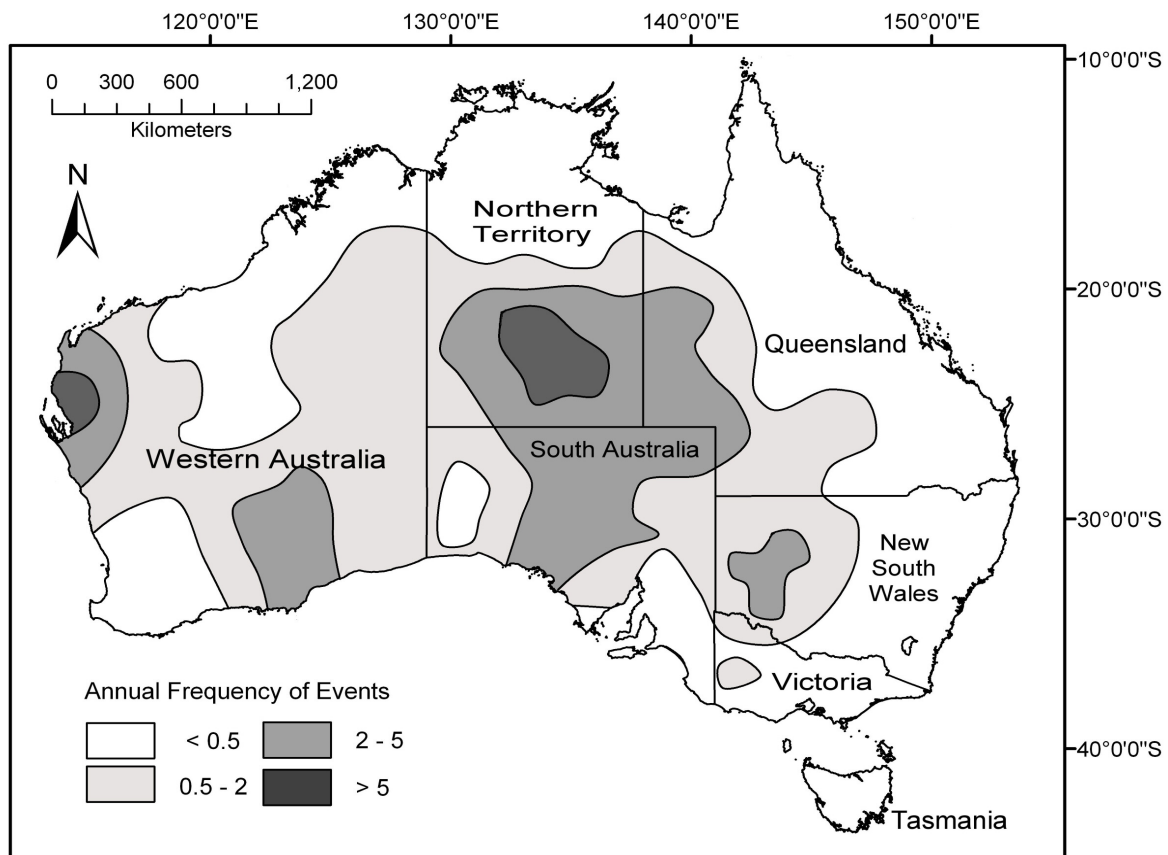


Figure 1.1 Map of mean annual dust-storm frequencies across Australia (1959-1980) (after McTainsh et al., 1989). Dust storms are defined as dust events recorded with visibility <1000 m.

The largest of Australia's dust-storm regions spans the southwest corners of Queensland, the Channel Country and the Simpson Desert – from Cameron Corner to Alice Springs (7.7-10.8

events yr^{-1}), and down to coastal South Australia. This high-frequency region extends east into the Mallee-Riverina and Wimmera regions of western NSW, an area recording ~ 6.3 dust storm events yr^{-1} . The third region is located between Onslow and Carnarvon ($8.1\text{--}5.6$ events yr^{-1}) on the Western Australian (WA) coast. Elevated dust-storm frequencies have been observed in southern WA (Carter, 1986); however, this region has received significantly less attention than wind eroding regions in eastern Australia. The areas of high dust storm activity are located over red siliceous sand dunes and red earthy sands, texture-contrast soils, massive calcareous earths, and cracking clay soils (Middleton, 1984; McTainsh and Leys, 1993). These soils are typically loose and mobile with poor aggregation, or in the case of the cracking clays, have a tendency to self-mulch and disaggregate.

McTainsh and Pitblado (1987) examined the frequency of different types of dust events across Australia. These included (following World Meteorological Organisation SYNOP Codes) dust storms (09), blowing dust (07), dust hazes (06) and dust whirls (08). While dust events in general have been observed throughout much of Australia, dust-event frequencies show a trend of increasing with increasing aridity toward the centre of the continent. This is consistent with global trends in dust source areas in Africa, the Middle East, China and North America (Goudie and Middleton, 2006). Blowing dust events occur less frequently and over a smaller area than dust storms, but in the same general source regions (McTainsh and Pitblado, 1987). The spatial extent of dust hazes and whirls is further restricted within the major regions frequently recording dust storms (Figure 1.1).

Climatic indices to model spatial patterns in dust-storm frequencies across Australia were developed by Burgess et al. (1989), McTainsh et al. (1990) and McTainsh et al. (1998). The models use soil moisture-erodibility (E_m) relationships defined by Chepil (1965), wind erosivity (E_w), and temporal variations between soil moisture, wind erosivity and wind erosion (E_t). The E-indices are based on Thornthwaite's (1931) precipitation-evaporation (P-E) ratio that defines aridity on a broad scale. The E_m index is able to explain 34% of the variance in dust-storm frequencies across Australia (Burgess et al., 1989). Excess wind erosion, defined where observed dust-storm frequencies are greater than those indicated by the models, occurs around Carnarvon (WA), Alice Springs (NT) and in a belt running east from Ceduna (SA) to Mildura (NSW) and up to Charleville in southwest Queensland. McTainsh et al. (1990) improved the performance of the E_m -index by including a factor to account for mean annual wind run (E_w). They found the E_w index explains 66% of the

variance in dust-storm frequencies across eastern Australia. McTainsh et al. (1998) developed a seasonal component in the model (Et) to explain the effect of antecedent soil moisture conditions on wind erosion as they change on a seasonal basis. The index has a strong positive correlation ($r^2 = 0.9257$) with mean annual dust storm-frequencies in Queensland, and in south-eastern Australia ($r^2 = 0.6946$), but by its nature only models dust storm activity and at broad spatial scales (e.g. $>10^5 \text{ km}^2$).

Geographically, the dust-storm regions east of 135° longitude are located in the Lake Eyre and Murray-Darling river basins. The areas receive rainfall $<450 \text{ mm yr}^{-1}$ and have evaporation rates in excess of 2000 mm yr^{-1} (McTainsh et al., 1989). The river basins span arid and semi-arid climate zones and have their headwaters in humid, higher rainfall zones in northern and eastern Australia. There is an important aeolian-fluvial interaction within the basins, with periodic inundation of the expansive river floodplains providing a mechanism for fine erodible sediment recharge (Bullard and McTainsh, 2003). Western Australia has not been identified as a major dust source area. Whereas eastern Australia is considered (wind) energy-limited in terms of potential wind erosion and dust production, it has been suggested that an arid climate and lack of coordinated drainage (river networks) contribute to Western Australia being sediment supply limited (Bullard and McTainsh, 2003). Land areas susceptible to wind erosion outside eastern Australia occur in agricultural lands (the Eyre Peninsular and southern WA) and in dunefields that are disturbed by fire or livestock grazing (Great Sandy, Gibson and Great Victoria Deserts). These are located in areas with sandy soils that are inherently susceptible to mobilisation by wind, for example around Carnarvon (WA).

1.2.2 Temporal Patterns in Wind Erosion

Wind erosion activity in Australia is characterised by significant inter-annual variability (Figure 1.2). This variability has been attributed to multiple factors including: land degradation episodes in the 1960s in central Australia and in the 1970s in south-west Queensland; a spread of woody weeds associated with overgrazing in the 1960s; and a decrease in the feral rabbit population in the 1970s (McTainsh and Leys, 1993; McKeon et al., 2004). Drought is arguably the greatest driver of inter-annual variations in wind erosion in Australia (McTainsh et al., 2005). Increased dust storm activity in the 1960s, early-1980s, mid-1990s and from 2002-2006 has been attributed to drought episodes.

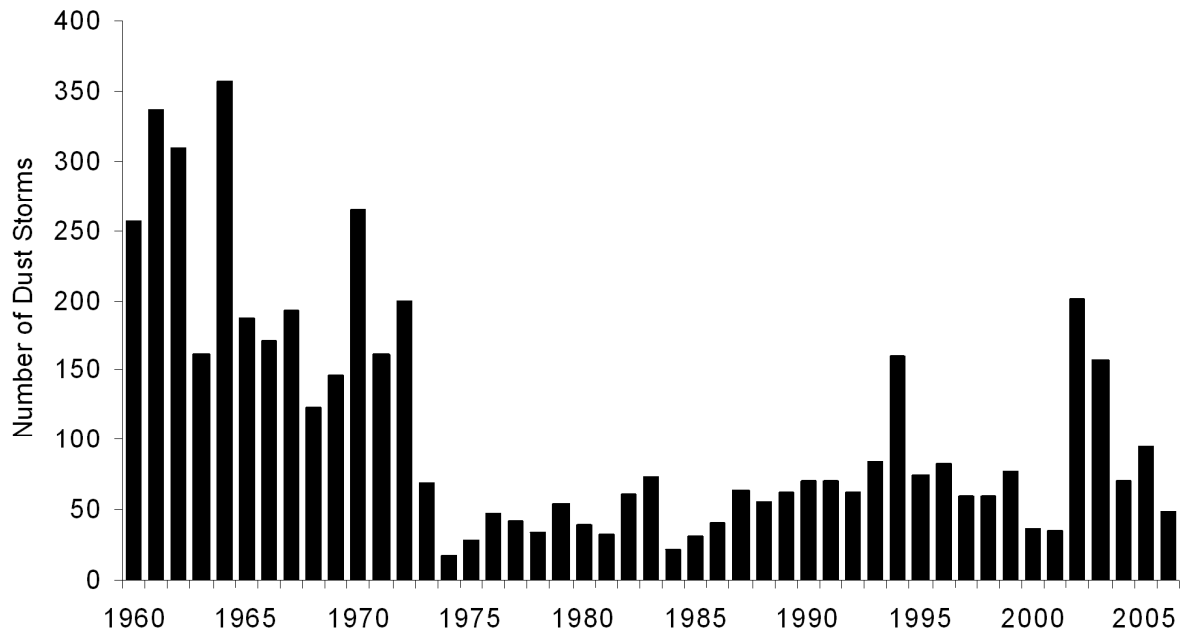


Figure 1.2 The frequency of dust storms across Australia from 1960-2006 (McTainsh et al., 2005; update from BoM). Dust storms are defined as dust events recorded with visibility <1000 m.

McTainsh et al. (1989) reported a large variance in the number of dust storms recorded across Australia's rainfall zones. They found that the number of observed dust events declines in areas receiving rainfall <300 mm yr⁻¹. The relationship provides further evidence of an aeolian-fluvial interaction in the regions receiving higher annual rainfall, with the driest parts of the continent in central Australia not receiving regular fine sediment recharge (as in the more humid Queensland Channel Country). At inter-annual time scales, some consistency has been found between the distribution of dust storms and drought affected areas (McTainsh et al., 1989; McTainsh et al., 2005). However, coastal WA (around Carnarvon) persistently records wind erosion activity irrespective of drought. This highlights the importance of considering factors driving temporal changes in the location of erodible land when assessing spatio-temporal patterns in wind erosion activity.

There is a seasonal pattern in wind erosion activity in Australia (McTainsh et al., 1998). Eastern Australia (east of 138° longitude) can be divided into two wind erosion regions, with the northern region (north of ~34° south) recording a peak in dust storms over spring (SON) and a minimum in autumn and winter (MAM and JJA). The southern region records a peak in dust storms later in summer (DJFM) and a minimum in winter (JJA). This seasonality has

been linked to the timing of rainfall which affects source area erodibility (vegetation cover, soil moisture), and wind erosivity (McTainsh et al., 1998). Both are affected by the west-to-east passage of rain-bearing frontal systems and increased windiness during spring and early summer (Eckström et al., 2004; Leslie and Speer, 2006).

The seasonal and inter-annual variations in wind erosion activity indicate that spatial and temporal changes in land susceptibility to wind erosion are important drivers of variations in wind erosion activity across Australia. The extent to which land erodibility varies in Australia is, however, yet to be quantified. Furthermore, research is yet to quantify the relationship between drivers of rainfall variability like the El Niño/Southern Oscillation (ENSO) and variations in the erodibility of the Australian landscape. Monitoring and modelling studies at spatial resolutions $<10^5 \text{ km}^2$ are required to elucidate the patterns and processes driving the variations in wind erosion activity.

1.2.3 Process Studies and Modelling

Numerous studies have been conducted in the cultivated lands of NSW to establish soil and vegetation relationships with wind erosion. The work includes research to quantify vegetation (Findlater et al., 1990; Leys, 1991a; Raupach et al., 1993), soil texture (Lorimer, 1985; Leys, 1991b; Leys et al., 1996) and soil aggregation and crusting (Leys and Eldridge, 1998; Eldridge and Leys, 2003) effects on wind erosion. This research produced quantitative information on wind erosion controls at the wind tunnel and field scales. However, few studies have examined the effects of these controls on wind erosion in broader spatial and temporal contexts. Research is therefore required to link this process knowledge to the landscape (10^3 km^2) and regional (10^4 km^2) scale wind erosion dynamics.

Since the 1990s a number of experiments have been conducted to assess wind erosion dynamics in the rangelands of western New South Wales, and the Channel Country and Mulga Lands of south-western Queensland. These included studies to assess soil loss and organic matter content in rural dusts (Miles and McTainsh, 1994; Boon et al., 1998), and particle size characteristics of dust emissions (Leys and McTainsh, 1996; Nickling et al., 1999). From 1994-2002 McTainsh and colleagues used wind vane samplers to monitor dust emissions from sand dunes, grassland downs and a claypan (playa) along the Diamantina River (McTainsh et al., 1999; McTainsh et al., 2002). Their results show that the

susceptibility of the rangelands to wind erosion is governed by complex relationships between soil types, vegetation cover and meteorological conditions, in particular rainfall quantities and timing (McTainsh et al., 1999). Subsequent studies sought to quantify changes in the erodibility of the claypan surface using remote sensing techniques (Chappell et al., 2003; Chappell et al., 2006; Chappell et al., 2007), and to quantify the effects of spatial variations in dust source erodibility on emissions (Butler et al., 2005). Importantly, none of these studies sought to map spatial and temporal patterns in land susceptibility to wind erosion at the landscape scale (10^3 km^2).

Little research has been conducted in Australia to model the spatial distribution of wind erosion. Lynch and Edwards (1980) used a pattern analysis approach for delineating wind erosion zones in New South Wales. Their model defined nine zones within the state with varying wind erosion hazards. The zones were aligned with the mean annual rainfall isohyets (increasing wind erosion risk with decreasing rainfall) and the pattern of dust-storm frequencies reported by McTainsh et al. (1998). They could not, however, predict the precise location of areas with a wind erosion risk.

Kalma et al. (1988) mapped potential wind erosion across Australia using an index of wind erosivity (after Fryberger, 1978). They found that stream lines of airflow over Australia follow an anti-cyclonic swirl about the centre of the continent. Variations in the flow occur in accordance with seasonal changes in strength of the Zonal Westerlies and the Trade Winds over northern Australia. Maximum drift potential was found to occur in October, and reach a minimum in April. This result supported later studies which show these times to be roughly coincident with periods of maximum and minimum wind erosion activity in central Australia (Eckström et al., 2004). The relative importance of seasonal variations in land erodibility in driving temporal changes in wind erosion activity remains yet to be considered in detail.

Shao et al. (1994) and Shao et al. (1996) presented the first process-based model to assess wind erosion in the Murray-Darling Basin of south-eastern Australia. The model was subsequently developed into an Integrated Wind Erosion Modelling System (IWEMS) for application on a national basis (Shao and Leslie, 1997; Lu and Shao, 2001). While the model has been used to simulate regional dust emissions, it has not been applied specifically to assess land susceptibility to wind erosion.

Despite a growing body of aeolian research in Australia, we are unable to describe which areas of the country are susceptible to wind erosion at high spatial resolutions (e.g. $<10^2$ km²). While the Lake Eyre Basin is Australia's largest dust source area, we have little knowledge of which parts of this expansive (1.14×10^6 km²) region are susceptible to wind erosion. It has been reported on a number of occasions that wind erosion activity in Australia is highly dependant on the spatial distribution of drought (McTainsh et al. 1989; McTainsh et al., 2005). However, research is yet to quantify the responses of bioregions in the arid and semi-arid zones to climate variability or land management. This means that we still do not know to what extent the susceptibility of the landscape to wind erosion varies at seasonal or inter-annual time scales, or how sensitive the landscape is to these external drivers.

To address this issue, Webb et al. (2006) developed an Australian Land Erodibility Model (AUSLEM) to assess land susceptibility to wind erosion across Australia. The model functions in a Geographic Information System (GIS) environment, using a rule-set to define conditions under which the landscape may become susceptible to wind erosion. The model uses inputs of soil texture (sand, silt and clay content), grass cover, soil moisture and rainfall, and is run at a 5 x 5 km spatial resolution on a monthly time-step. AUSLEM was validated through a comparison of output with spatial patterns in a Dust Storm Index (McTainsh and Tews, 1998). The analysis suggested that the model has potential to be developed into a useful tool for assessing land erodibility in Australia. However, the model threshold-based rule-set biased the output toward certain soil types and did not capture the full range of the soil erodibility continuum. Additional research is required to develop AUSLEM into a functional model to assess land erodibility in Australia. The development will provide the opportunity to assess spatial and temporal patterns in land susceptibility to wind erosion at landscape (10^3 km²) to regional (10^4 km²) scales and quantify the effects of climatic conditions on land erodibility dynamics. That research is presented in this thesis.

1.3 Problem Statement

A number of research problems exist that are relevant to our understanding of the distribution of land susceptible to wind erosion. These relate specifically to the study of spatial and

temporal patterns of potential wind erosion both in Australia and internationally. They include:

- There is a poor knowledge of exactly which areas of Australia are susceptible to wind erosion. This is a significant problem considering that Australia contains the dominant dust source area in the southern hemisphere – the Lake Eyre Basin.
- There is a lack of research into spatial and temporal patterns in land erodibility at the landscape scale. Research into land erodibility at this scale is essential if we are to better link field scale wind erosion processes to regional dust emission and transport processes.
- We have a poor knowledge of how soil and land erodibility respond to climate variability and land management, particularly in rangeland environments which cover ~45% of the global land surface.
- There is a lack of quantitative models to predict temporal changes in soil erodibility to wind. Soil erodibility is a fundamental control on wind erosion and so this issue affects any research that seeks to model wind erosion processes.
- There is a growing requirement to learn more about the sensitivity of rangelands to climate variability and land management pressures in light of uncertain future climate change. Assessing the landscape susceptibility to land degradation processes like wind erosion is an essential component of this research.

1.4 Thesis Aims and Objectives

The thesis has five aims. The aims seek to address the research problems listed in Section 1.3. They are:

1. To develop a framework for modelling temporal changes in soil erodibility in response to climate variability and land management pressures.
2. To develop AUSLEM into a functional model to assess land susceptibility to wind erosion, i.e. land erodibility, across western Queensland, Australia.
3. To validate the performance of the land erodibility model.
4. To map the spatial extent of areas susceptible to wind erosion in western Queensland.
5. To identify the role of climate variability in determining spatial and temporal patterns in land erodibility dynamics in western Queensland.

Each of the research aims are addressed through the following objectives:

1. Review the literature describing controls on land susceptibility to wind erosion. This objective will facilitate the development of a conceptual model to describe process-control interactions and provide a foundation for developing soil and land erodibility models (Objectives 3 and 4).
2. Review methods for modelling land erodibility as incorporated within current wind erosion modelling systems. This objective will facilitate an assessment of approaches for modelling land erodibility at spatial scales from the paddock to global scales. The review will identify common approaches for modelling the effects of controls on wind erosion identified in Objective 1, and will identify limitations to the models that will be addressed in Objectives 3 to 8.
3. Develop a conceptual model of the soil erodibility continuum and present a framework for modelling temporal changes in soil erodibility in response to climate variability and land management. This objective seeks to address limitations in land erodibility models identified under Objective 2, and improve the assessment skill of AUSLEM.
4. Develop AUSLEM into a functional land erodibility model that can be applied to assess land susceptibility to wind erosion across western Queensland. The model will be based on process relationships reviewed in Objective 1, and will address deficiencies in the capabilities of wind erosion modelling systems identified in Objective 2.
5. Develop a method for visually assessing land erodibility that can be used to monitor rangeland conditions at the landscape scale and provide data that can be used to validate spatially explicit land erodibility models.
6. Validate the performance of the land erodibility model. This objective will utilise observational records of blowing dust activity and data collected by visual assessments of land erodibility (Objective 5) in a comparison with model predictions.
7. Apply the model developed in Objective 4 to assess spatial and temporal patterns in land erodibility across western Queensland. Spatial patterns in model output showing land areas susceptible to wind erosion will then be related to bio-geomorphic landscape characteristics.
8. Identify processes driving temporal patterns in land erodibility dynamics in western Queensland. Spatial and temporal patterns in model output will be analysed in the context of factors controlling regional scale climate variability over western Queensland.

1.5 Research Approach

This research develops a model that can be used to assess land susceptibility to wind erosion. What sets the research apart from other modelling studies is the focus between the field scale (10^3 m^2) and regional ($>10^4 \text{ km}^2$) to global scale studies. Developing a model to operate between these scales will address a significant gap in our ability to assess and map land areas susceptible to wind erosion. It will also improve our understanding of linkages between field scale processes driving spatio-temporal variations in dust emissions, and regional to global scale dust transport processes. This research is essential for developing an advanced understanding of which areas of the Earth surface are susceptible to wind erosion, and how the erodibility of these areas changes through time in response to climate variability and land management.

Model development has been directed toward assessing land erodibility in rangeland environments. Processes and controls specific to cultivated regions have not been accounted for. This is because wind erosion in cultivated areas, for example in south-eastern Australia, has previously been the focus of much scientific research. Spatial and temporal patterns in potential wind erosion in rangeland environments have received considerably less attention. Australia's rangelands cover the largest dust source in the southern hemisphere, the Lake Eyre Basin. Monitoring and modelling land erodibility in this environment will make a significant contribution to our understanding of a major global dust source area. The research also provides a modelling framework that can potentially be adopted for application in other dryland environments.

The research does not seek to develop a model that can necessarily be integrated into existing modelling systems. The model was developed so that it can be applied with readily available input data, is robust, conceptually easy to understand, and is capable of capturing the nature of relationships between land surface conditions and land erodibility. This addresses data requirements and application limitations of current wind erosion models. These aspects are critical for the application of a model at the landscape (10^3 km^2) and regional scales (10^4 km^2), and in particular for the potential uptake of the research in land management and policy development contexts. Current wind erosion models are limited by their inherent complexity, and data and computing power requirements (Raupach and Lu, 2004). These attributes make

them inaccessible to those without appropriate resources or an understanding of the micro-physical processes on which they are based.

It is acknowledged that the modelling approach adopted in this research has a number of limitations, some of which are common to existing wind erosion models. These relate to challenges in accounting for the heterogeneous distribution of vegetation cover in arid and semi-arid environments, and an inability to account for temporal changes in soil erodibility. Developing methods to account for the spatial distribution of vegetation cover in wind erosion models was beyond the scope of this research; however, the issue of soil erodibility modelling is addressed in this thesis.

Finally, the development of models to assess complex dynamic systems is an iterative process. The research presented in this thesis reflects this characteristic. There is, however, a focus on presenting the research successes. Early into the research a field monitoring program was established in an attempt to obtain quantitative data on the interactions between meteorological and land surface conditions controlling land erodibility in the western Queensland rangelands. Considerable resources were directed toward the field study, including substantial support from volunteers. Ultimately, a reliance on coarse sampling resolutions meant that the data collected was insufficient to resolve process interactions suitable for incorporation into the model. For this reason the data have not been included in this thesis. The experience I gained from spending time working at the field sites was, however, invaluable in developing an understanding of the rangeland system.

1.6 Study Area

The study area is the arid and semi-arid rangelands of western Queensland, Australia (Figure 1.3). The region forms the north-eastern half of the Lake Eyre Basin, the dominant dust source area in Australia and in the southern hemisphere (Goudie and Middleton, 2006). The study area is ~672 000 km² in size and can be divided into four biogeographical regions. These include: the Mulga Lands, Mitchell Grass Downs, Channel Country, and Simpson-Strzelecki Dunefields. Each of the regions can be described by characteristic landforms and vegetation types (DEWHA, 2007).

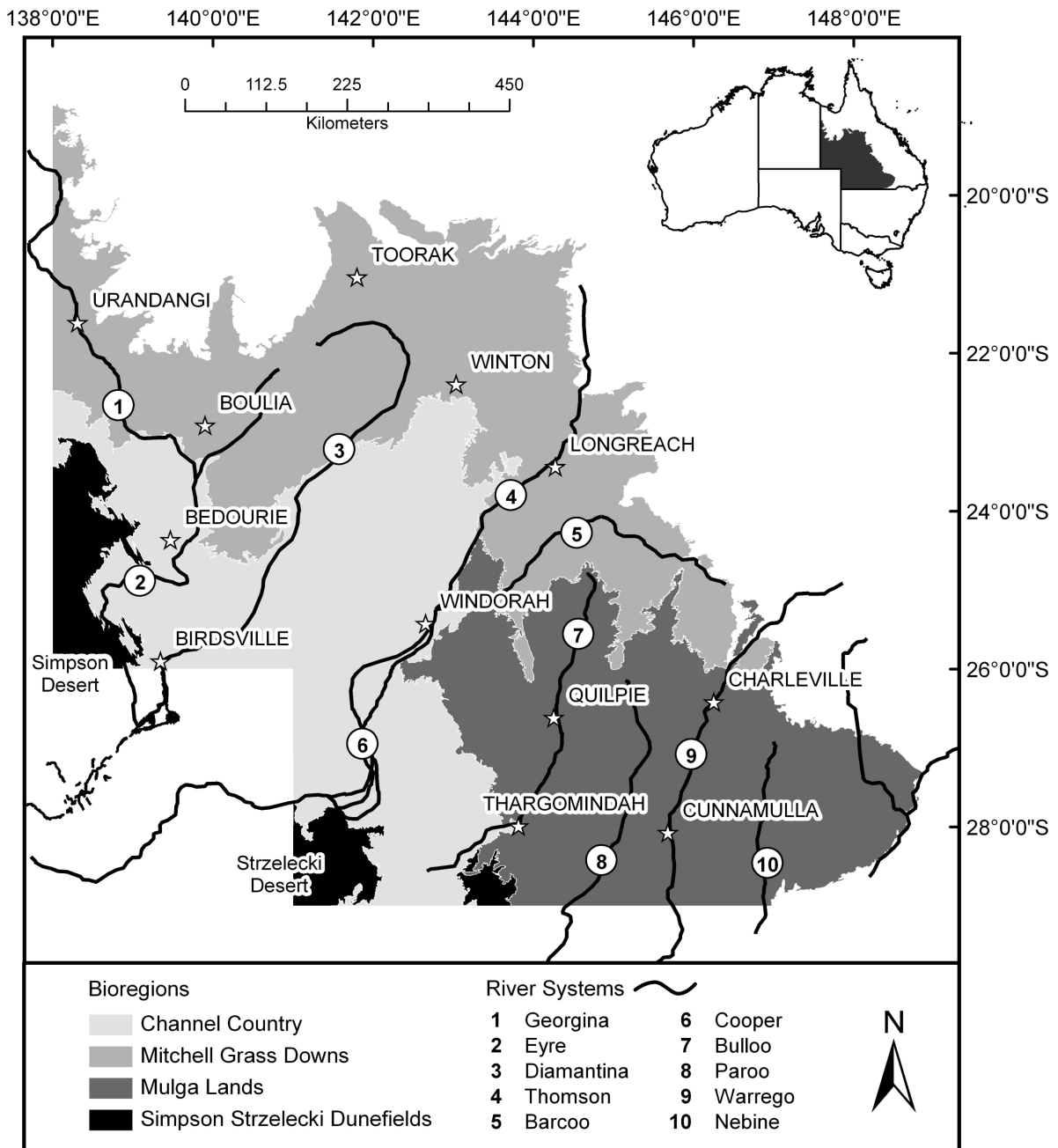


Figure 1.3 Map showing study area locations, the four bioregions, and the Simpson and Strzelecki Deserts. Major river systems are labelled. These are ephemeral systems that drain inland to the south of the study area. The rivers contain some permanent water holes.

Channel Country (195 825 km²):

The Channel Country bioregion is characterised by braided, flood and alluvial plains of the Georgina, Diamantina and Cooper River systems. Gibber plains, dunefields and low ranges surround the river floodplains. Soils on the floodplains of the river systems are grey and brown cracking clays (vertisols). Kandosols, rudosols and sodosols lie beneath the gibber plains and on the dissected ranges separating the Georgina, Diamantina and Cooper river catchments. The south-eastern border of the Channel Country is characterised by low tablelands and undulating plains that run into the Mulga Lands. Vegetation in the north of the Channel Country is characterised by *Acacia* spp. shrublands and hummock grasslands (*Astrebla* spp.). The floodplains in the central and southern areas of the bioregion are covered with open herbfields and grassland downs (Figure 1.4). Coolibah (*Eucalyptus coolabah*) woodlands fringe the river channels dissecting the bioregion. Spinifex grasses (*Tiodia* spp.) cover the sandplains to the west of the bioregion.

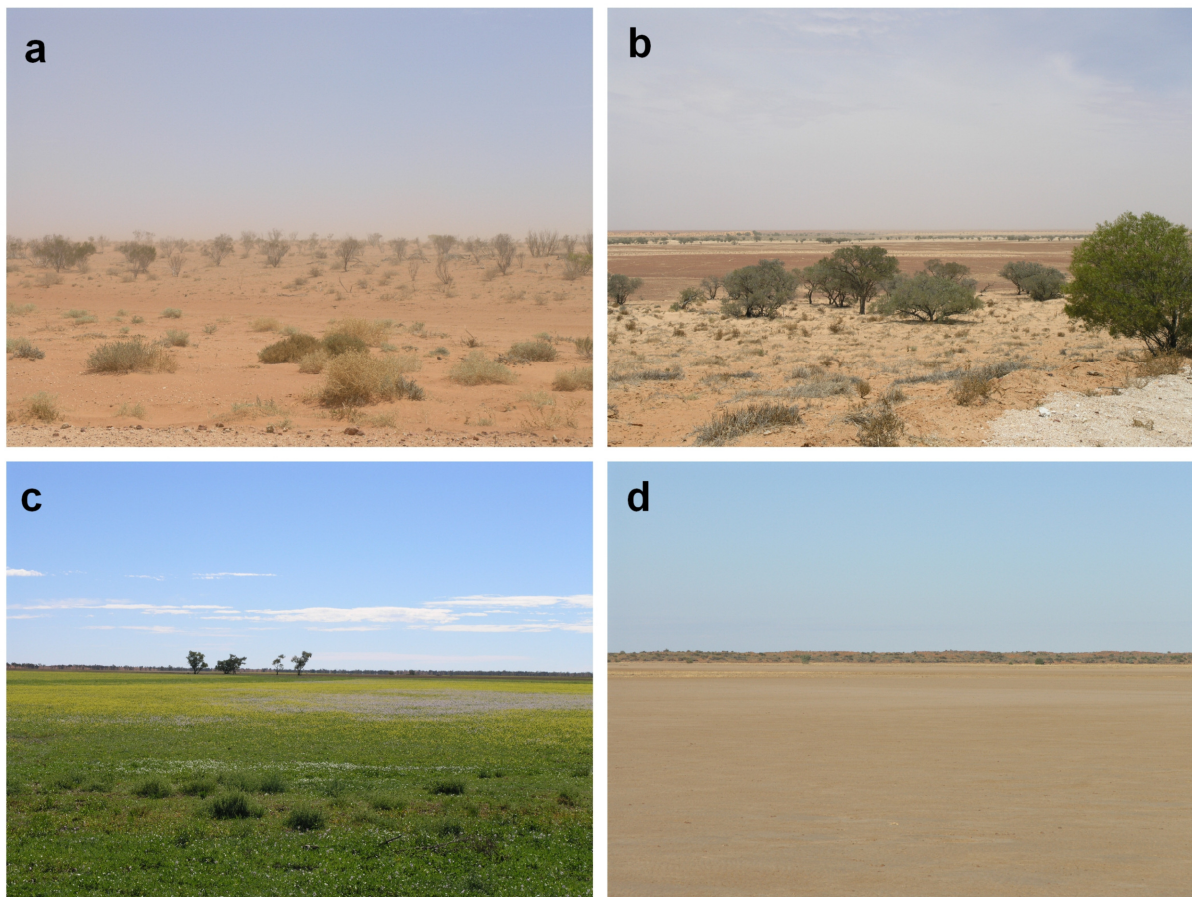


Figure 1.4 Images of the Channel Country showing (a) an eroding sandplain on the eastern side of the Simpson Desert, (b) low dunes and gibber plains along the Eyre Creek, (c) lush pasture following flooding of Coopers Creek near Windorah, and (d) an expansive claypan on the Diamantina River

Mitchell Grass Downs (245 635 km²):

The Mitchell Grass Downs bioregion is characterised by undulating grassland downs. The bioregion lies across the headwaters of the river systems flowing through the Channel Country. Soils are predominantly grey and brown cracking clays (vertisols) that are self-mulching during wet-dry cycles. This bioregion supplies the fine silts and clays that are deposited on the lower river floodplains following periodic flood events. To the west of the bioregion rudosols and chromosols are found on the low ranges separating the Georgina and Diamantina river catchments. The dominant vegetation in the bioregion includes the Mitchell grasses (*Astrebla* spp.), which cover the expansive open downs (Figure 1.5). Tree cover is sparse or absent from much of the bioregion, however stands of Gidyea (*Acacia cambagei*) and Boree (*Acacia tephрина*) are found in the eastern parts of the bioregion, particularly on dissected ridgelines bordering the Channel Country.

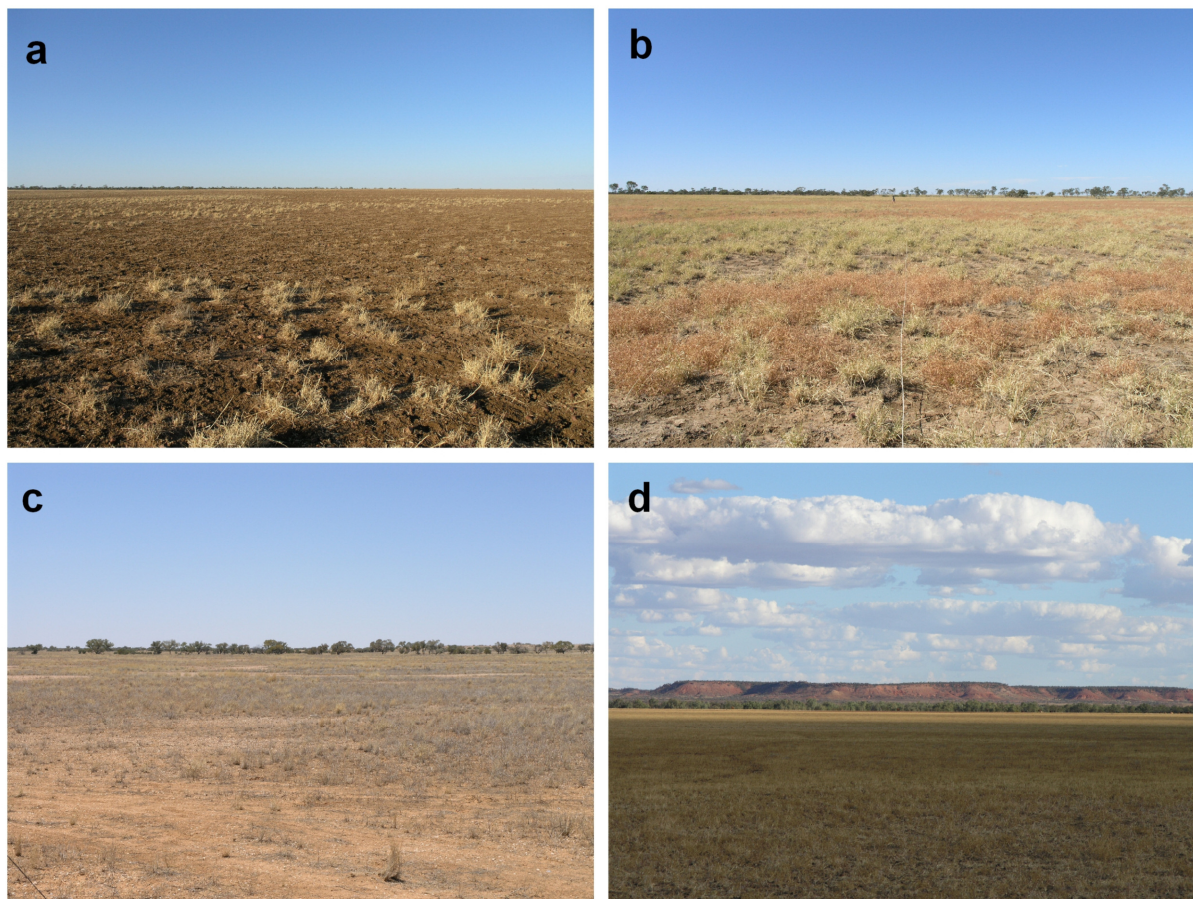


Figure 1.5 Images of the Mitchell Grass Downs showing (a) grassland pasture during a winter drought, (b) the same location following 100 mm rainfall in one month, (c) degraded pasture west of the Diamantina River, and (d) pastures and mesas to the northeast of the Diamantina River. Stands of Gidyea trees can be seen in (b), (c) and (d)

Mulga Lands (192 469 km²):

The Mulga Lands are characterised by undulating plains and low hills on red earth soils (kandosols). Grey clays (vertisols) are common on the floodplains of the major river systems. The eastern half of the Mulga Lands lies in the Murray-Darling Basin, with the Nebine, Warrego and Paroo Rivers draining into the Darling River. The western side of the Mulga Lands bioregion is covered by small internal drainage depressions, e.g. Lake Bindegolly, and the Bulloo River Basin. The Bulloo River flows from the north-east of Quilpie, terminating in a large floodout south-west of Thargomindah. Vegetation in the Mulga Lands is characterised by shrublands and low woodlands (*Acacia aneura*). Tree density and herbaceous vegetation cover decrease from east to west across the bioregion with increasing aridity. Expansive open grassy areas, referred to as “black soil plains” are located through the central region of the Mulga Lands, between Quilpie, Cunnamulla and Thargomindah. These plains have no tree cover and overgrazing can result in a complete reduction in herbaceous cover, leaving bare surfaces that are susceptible to erosion (Figure 1.6).

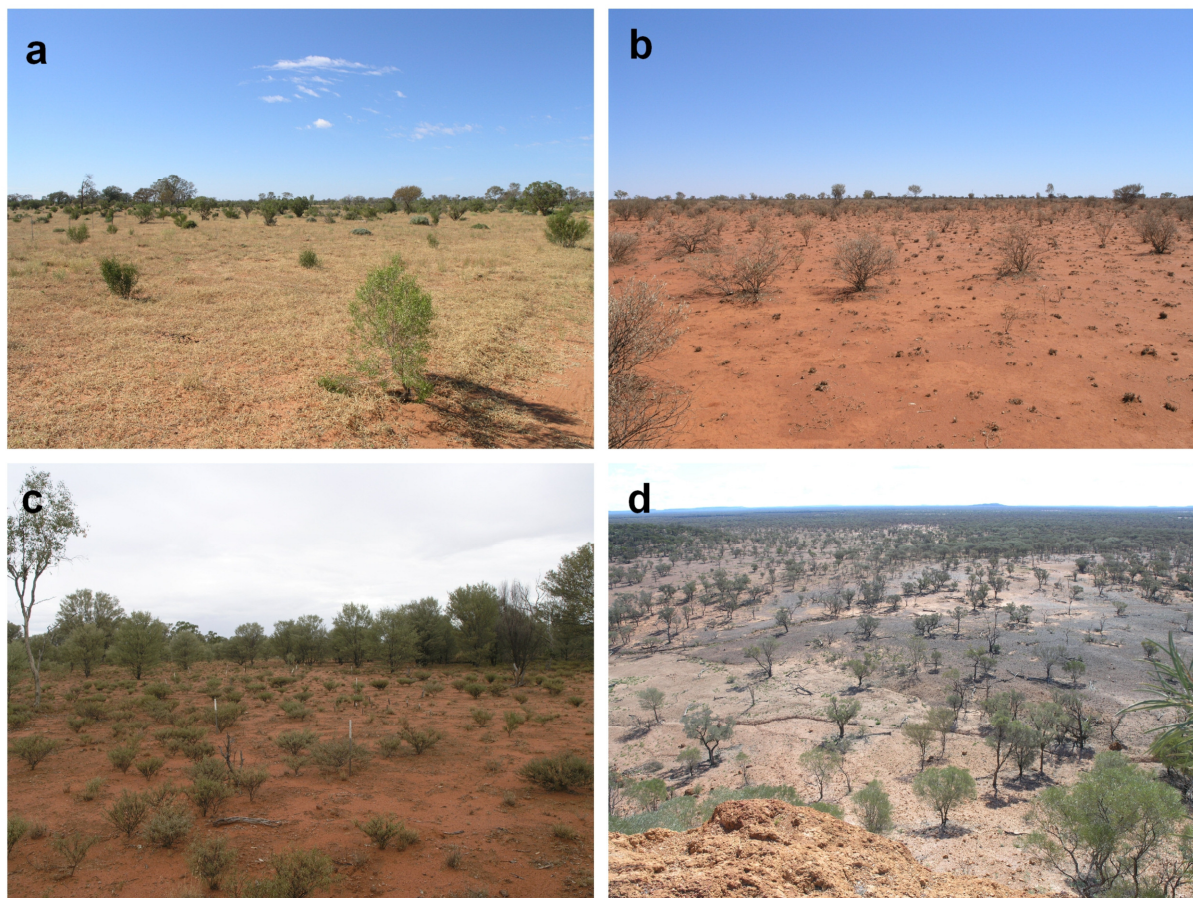


Figure 1.6 Images of the Mulga Lands showing (a) and (b) high and low cover on a mulga sandplain near Thargomindah, (c) woody weed incursion near Charleville, and (d) “hard mulga” lands.

Simpson-Strzelecki Dunefields (38 079 km²):

The Simpson-Strzelecki Desert bioregion is characterised by arid dunefields and sandplains, extensive saltpans and dry riverbeds and floodplains. The dunefield soils are rudosols, with grey clay vertosols fringing the dunes on the floodplains of the Georgina River and Eyre Creek to the east of the bioregion. The dunefields consist of longitudinal dunes with crests generally 5-30 m high and crest spacing from 200-500 m (Mabbutt, 1977). The dunes vary in length and may be up to a few hundred kilometres long. Stony gibber plains surround the eastern margins of the dunefields. Rivers dissect the dunefields but flows are intermittent, with large flood events filling the inter-dune areas. The longitudinal dunes are generally stabilised with hummocky Spinifex (*Triodia* spp.) and cane grasses (*Zygochloa paradoxa*) which grow on the dune flanks and inter-dune areas (Figure 1.7). River channels following the inter-dunes are fringed with narrow woodlands of Coolibah (*Eucalyptus coolabah*) and *Acacia* species. Dune crests may be devoid of vegetation and become highly erodible following fire (McGowan and Clark, 2008).

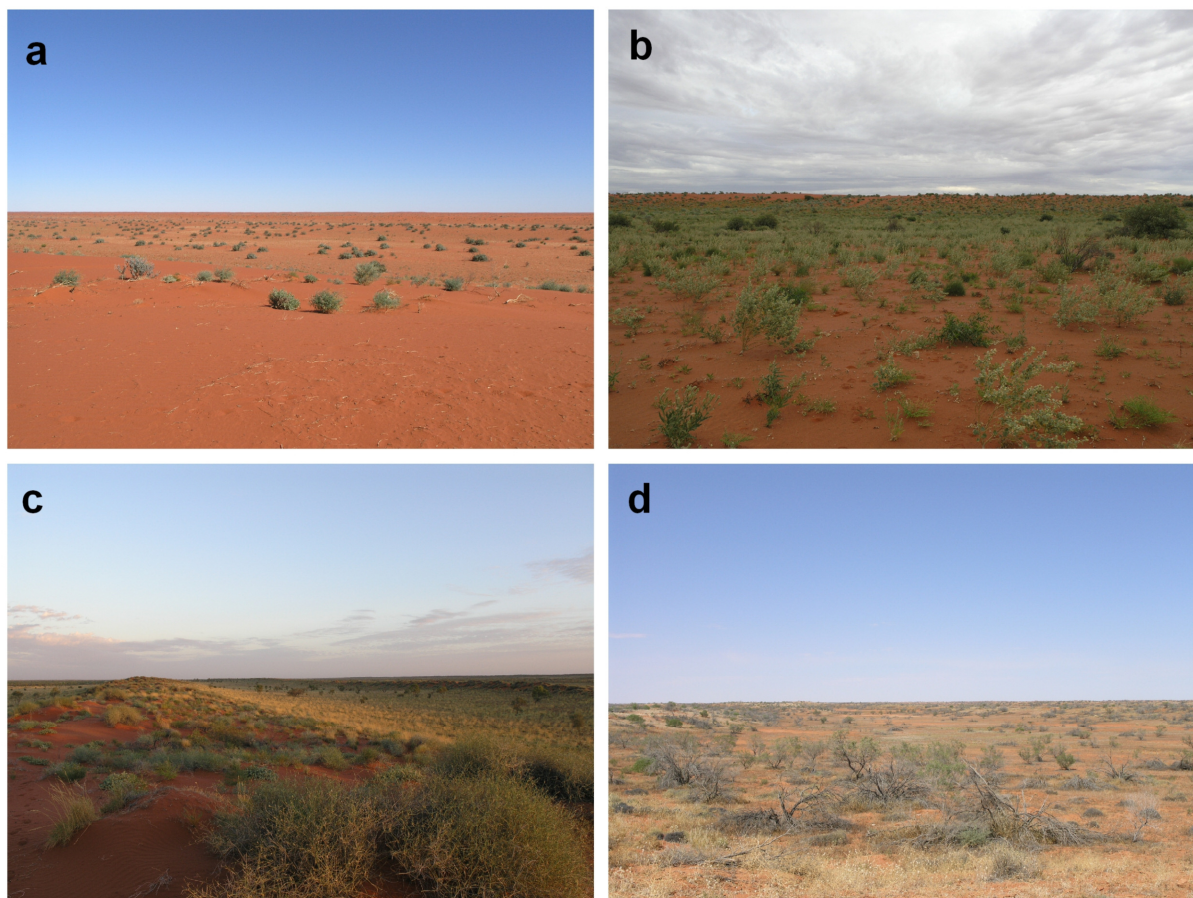


Figure 1.7 Images of the Simpson and Strzelecki Deserts showing (a) a fire scar in 2006 following burning in 2002, (b) the same site in 2007 following flooding of the site, (c) vegetated dune crest and swale typical of un-burnt areas, and (d) sand dunes in the Strzelecki Desert

The study area has an arid to semi-arid climate (<200-500 mm annual rainfall) with seasonal rainfall in summer (DJF). Maximum air temperatures range from over 40°C in summer to the high-teens and mid-twenties in winter (JJA). Mean annual rainfall ranges from 450 mm to the east of the study area to <200 mm in the south-west corners of the state (Figure 1.8).

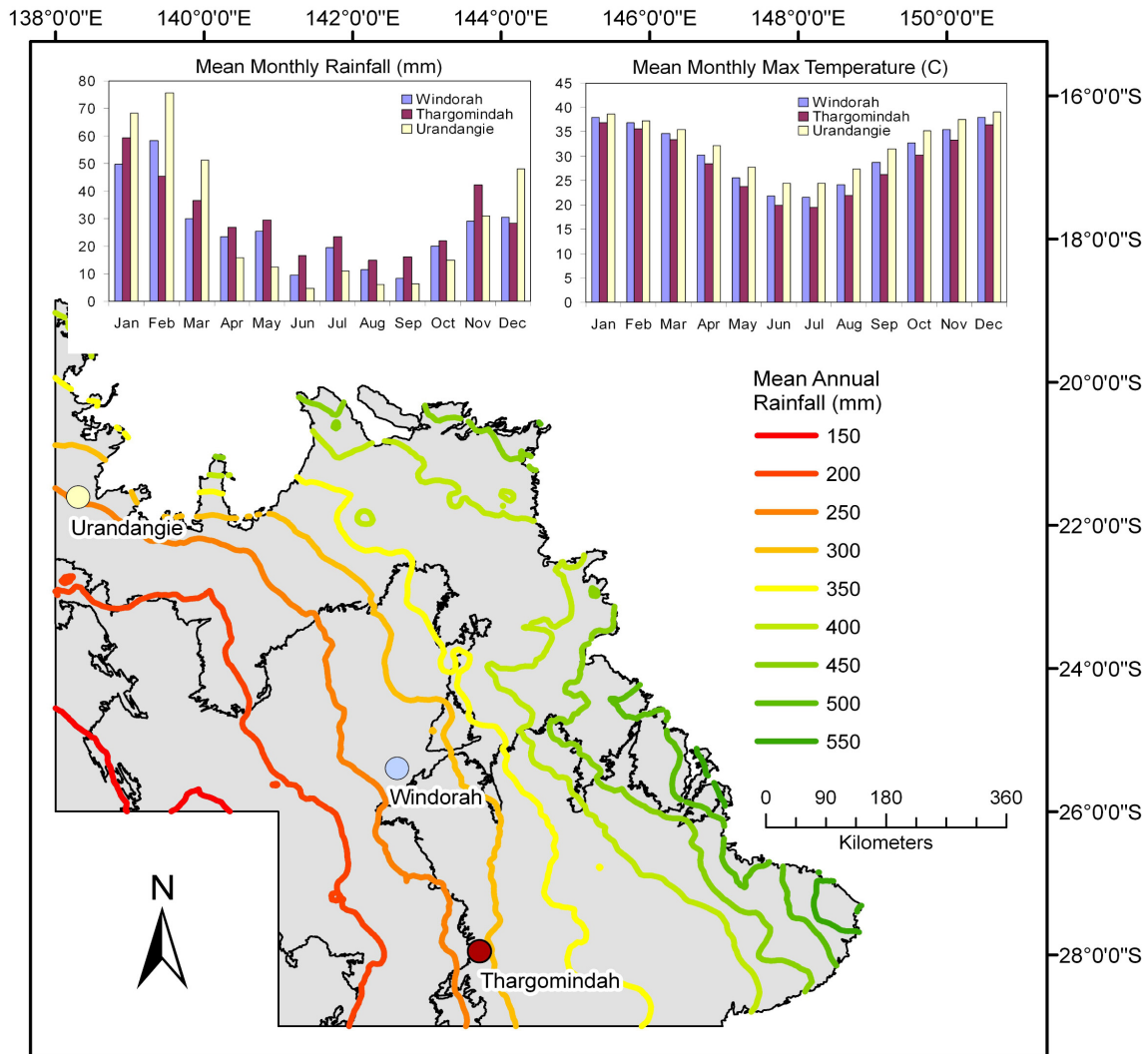


Figure 1.8 Map of study area showing mean annual rainfall isohyets and mean monthly rainfall and maximum (max) temperature variations at Thargomindah, Windorah and Urandangie (data source BoM, 2007a).

Annual evaporation varies from 2400 mm along the eastern boundary of the study area to >3600 mm in the south-west. Concurrent to these trends, aridity increases from east to west across the study area, and this is matched by a similar trend in increasing dust-storm frequencies (McTainsh et al., 1990). The boundary of the semi-arid zone (along the 450-500 mm rainfall isohyet) borders the eastern boundaries of the Mulga Lands and Mitchell Grass

Downs. Wind erosion is infrequently observed outside this zone due to higher annual rainfall and vegetation cover, so that area is not considered in this study.

There is a seasonal pattern in wind speeds across the study area. The highest wind speeds are associated with southerly to south-easterly air flow (BoM, 2008). This flow is largely driven by the movement of a baric ridge across central Australia during the winter months and anti-cyclonic flow around the high pressure system. Dust entrainment is associated with the passage of cold fronts and trough lines across the study area (McGowan et al., 2000). The frontal systems originate in the Southern Ocean to the south-southwest of the Australian continent and travel north over the study area through the cols that develop between subsequent eastward tracking anticyclones. This synoptic setup generates strong pre-frontal northerly winds through central Australia. Behind the fronts and/or trough lines, anticyclonic ridging may generate moderate to strong south to south-easterly winds over the study area which are also responsible for dust entrainment (Sturman and Tapper, 2001).

Rainfall seasonality varies from north to south across the study area (Figure 1.8). The northern regions are influenced by the Australian Summer Monsoon, thunderstorm activity and the incursion of rainfall depressions from the east coast during the summer months. The southern regions of the study area have a less pronounced summer rainfall peak and a secondary peak during winter. However, winter is generally the driest time of the year across the study area.

At inter-annual time scales rainfall variability across the study area is associated with the El Niño/Southern Oscillation (ENSO) (Whetton, 1997). Pittock (1975) reported a positive correlation ($r^2 = 0.4$) between rainfall and the Southern Oscillation Index (SOI), an indicator of the intensity of positive (La Niña) and negative (El Niño) ENSO phases, and rainfall in eastern Australia. The correlation between rainfall and the SOI varies seasonally, and is influenced by phase interactions of ENSO (3-7 year cycle) with the Pacific (inter-) Decadal Oscillation (PDO, a 15-30 year cycle) (Crimp and Day, 2003). Consequently, episodes of pasture degradation and recovery in the study area are linked to variations in Australian rainfall driven by ENSO-PDO interactions (McKeon et al., 2004).

Land use in the study area is dominated by pastoral activity with sheep and cattle grazing of rangelands. Sheep grazing is dominant to the east the Mitchell Grass Downs and Mulga

Lands. Cattle are typically grazed in the drier western Mulga Lands, Channel Country, Mitchell Grass Downs and Simpson-Strzelecki Desert bioregions. Property sizes increase from east to west across the study area, consistent with increasing aridity and decreasing stocking rates.

1.7 Thesis Structure

The thesis contains six chapters which address the research objectives. These include two review chapters followed by four chapters presenting research into model development, validation and application. This is followed by a synthesis of the research outcomes. Figure 1.9 is a flow chart outlining the structure of the thesis and links the thesis chapters to the research objectives.

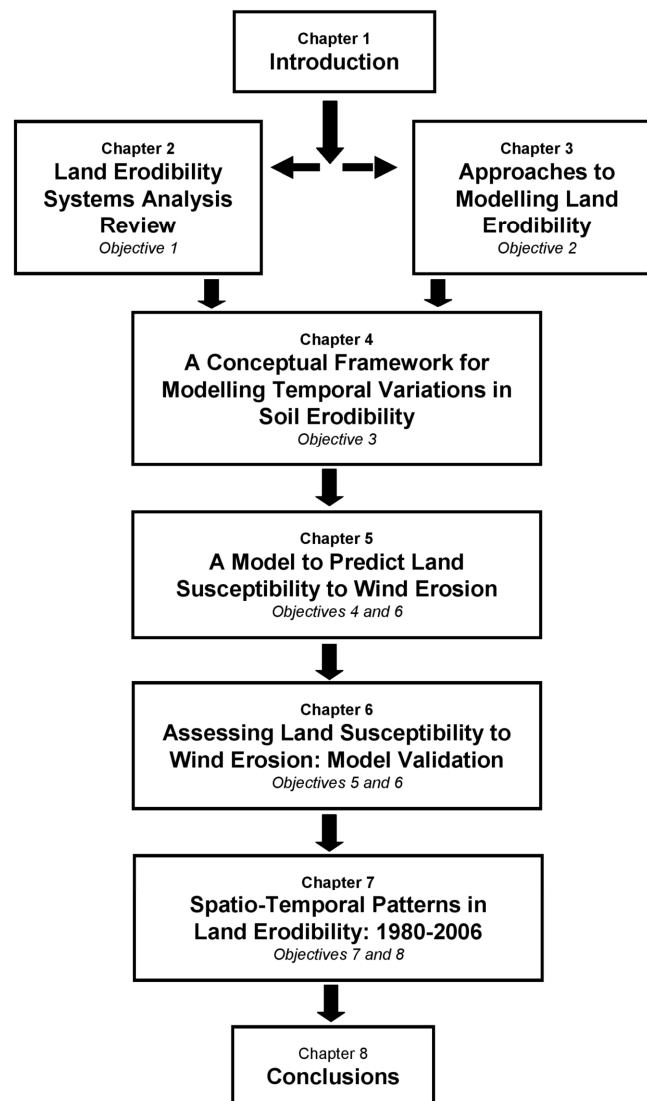


Figure 1.9 Flow chart of the thesis structure, linking the thesis chapters to the research objectives.

Chapter 2 provides a systems analysis of the controls on wind erosion, discusses anthropogenic influences on wind erosion, and presents a conceptual framework for modelling land erodibility. Chapter 3 presents a review of approaches to modelling land erodibility to wind, and discusses requirements for future research into model application to address a deficiency in land erodibility mapping. Chapter 4 examines a significant limitation to wind erosion modelling, presenting a framework for modelling temporal changes in soil erodibility in rangeland environments. The framework is based on a soil crust formation-disturbance temporal model. Chapters 5, 6 and 7 constitute three stand-alone publications. Minor alterations have been made to these publications to integrate the research with previous chapters in the thesis and to remove repetition in the study area descriptions. Chapter 5 reports on the development and validation of a model (Australian Land Erodibility Model, AUSLEM) to predict landscape to regional scale patterns in land erodibility across western Queensland, Australia. Chapter 6 presents a field method for monitoring land susceptibility to wind erosion at the landscape scale and examines application of the data for validating AUSLEM. Chapter 7 reports on the application of AUSLEM to assess land erodibility dynamics in western Queensland between 1980 and 2006. Spatial and temporal patterns in land erodibility dynamics are then examined in the context of regional to global scale climate variability. Finally, Chapter 8 synthesises the outputs and outcomes of the research. The thesis concludes by identifying the priorities for future research.

Chapter 2

Land Erodibility to Wind: Systems Analysis

This chapter addresses Objective 1 by presenting a review of the concepts of soil and land erodibility, and a systems analysis of the factors controlling wind erosion. The systems analysis addresses how meteorological, soil and vegetation conditions affect land susceptibility to wind erosion. The review is synthesised in a conceptual model of the land erodibility continuum. The conceptual model can be used to understand approaches for modelling wind erosion reviewed in Chapter 3, and provides the foundation for developing new soil and land erodibility models in Chapters 4 and 5.

2.1 Erodibility Concepts and Rankings

The term “erodibility”, denotes susceptibility to erosion. Definitions of the term are often scale and process dependent. This means that in any study of erosion the term must be clearly defined to avoid confusion. Bryan et al. (1989) review the use of “erodibility” in fluvial research. They identify implicit assumptions about which the term has been applied. These include (after Bryan et al, 1989: 393):

- A soil erodibility ranking can be defined which is valid for all erosional processes;
- A soil erodibility ranking can be uniquely defined by measurement of a few, usually physical, soil properties; and
- A relative erodibility ranking is not affected by short-term changes, particularly in soil moisture status.

The assumptions provide a good starting point for a discussion about the application of the term in the field of aeolian research. In this field the term has been used to describe the susceptibility of soils and land areas to wind erosion. Houghton and Charman (1986) provide definitions of the terms erodibility, erosion hazard, and erosion risk in the context of wind erosion:

Erodibility	• Susceptibility of a soil to detachment and transportation by an erosive agent. • A composite expression of soil properties that affect the mechanical, chemical and physical properties of a soil. Is independent of topography, land use, rainfall intensity and plant cover, but may be changed by management.
Erosion Hazard	• Susceptibility of a parcel of land to erosion. • Dependent on a combination of climate, landform, soil, land use and management factors.
Erosion Risk	• Intrinsic susceptibility of a parcel of land to erosion. • Dependent on climate, landforms and soil factors, but independent of land management.

Erodibility and erosion hazard/risk work opposite to “erosivity”. Erosivity is the potential or capacity of the wind to mobilize particles on the soil surface (Lal and Elliot, 1994). Susceptibility rankings are based on the premise that a location having a high erodibility is more susceptible to erosion than a location with low erodibility, and that a highly erosive wind has more energy than a wind with low erosivity. Conceptually, erodibility is independent of wind erosivity. However, if erodibility is to be measured to develop a ranking system, wind erosivity must be considered. This necessity is reflected in even the earliest studies of wind erodibility. These studies applied wind tunnel experimentation and field monitoring to determine the susceptibility of soils to wind erosion in cultivated fields.

Chepil (1953) used constant wind erosivity conditions to determine soil and wind tunnel erodibility factors for the Wind Erosion Equation (WEQ). More recent studies have defined erodibility rankings under a range of wind erosivities. Studies in the late 1950s to 1960s provided a range of approaches for developing erodibility factors (e.g. Chepil and Woodruff, 1963; Woodruff and Siddoway, 1965). The factors can be characterised by differences in the spatial scales at which they were derived (Table 2.1).

Table 2.1 Summary of selected erodibility factors used in the Wind Erosion Equation (WEQ) model.

Erodibility Index	Definition	Spatial Scale	Time Scale	Measurement	Reference
Soil Erodibility (<i>I</i>)	- Potential annual wind erosion for a given soil under a given set of field conditions - Expressed as the average annual soil loss in tons per acre per year for a field that is isolated, unsheltered, bare, smooth, level, loose, non-crusts, and at a location where the climatic factor (<i>C</i>) for the WEQ is 100	Plot and Field	Minutes to Months	Wind Tunnel Dust Traps	Chepil and Woodruff (1963) Woodruff and Siddoway (1965)
Knoll Erodibility (<i>I_s</i>)	- Accounts for the increased wind erosion potential on topographic features (i.e. short slopes, knolls) - Varies for slopes with gradient > 3% and < 150 m long - An adjustment to <i>I</i> given the increase in wind velocity on the windward slope	Field	Minutes to Months	Field	Woodruff and Siddoway (1965)
Wind Tunnel Erodibility (<i>i</i>)	- Index of relative erodibility under wind tunnel with constant conditions of friction velocity of 0.61 ms⁻¹ for five minutes over soil in trays with dimensions 5 ft long and 8 inches wide - Expressed in tons per acre - Due to fetch and tray length restrictions in wind tunnel, <i>i</i> does not account for abrasion experienced under field conditions	Plot	Minutes	Wind Tunnel	Chepil (1953)
Field Erodibility (<i>E</i>)	- Determined by application of the Wind Erosion Equation to determine potential soil loss in tons per acre per year - Accounts for soil (<i>I</i>) and knoll (<i>I_s</i>) erodibility, local wind erosion climatic factor, soil ridge roughness, field length and vegetation cover.	Field to Region	Annual	Modelled	Woodruff and Siddoway (1965)

There is a tendency for erodibility factors to deviate from the assumptions and formal definitions of erodibility presented by Houghton and Charman (1986) and Bryan et al. (1989). This has arisen from authors creating new definitions of the term to suit particular studies. For example, while the soil erodibility factor (I) holds to the definition of erodibility (Houghton and Charman, 1986), the field erodibility factor (E) could be defined as a measure of wind erosion hazard. While Geeves et al. (2000) note that the factor is a contradiction to erodibility definitions, nearly all other ‘erodibility’ indices developed since could also be considered to be so.

The United States Department of Agriculture (USDA) developed the Wind Erodibility Groups (WEGs) to provide a ranking of the erodibility of soils. The scheme was based on wind tunnel experiments conducted by Chepil (1953) and Zingg (1951). The WEGs rank the erodibility of soils based on surface textural characteristics and an assigned wind erodibility index (Table 2.2).

Table 2.2 Wind Erodibility Groups and Wind Erodibility Index for soils in the United States (after Skidmore et al., 1994).

WEG	Soil Description	Wind Erodibility Index (t/ha)
1	Very fine sand, fine sand or coarse sand	659
2	Loamy very fine sand, loamy fine sand, loamy sand, loamy coarse sand or sapric soil materials	300
3	Very fine sandy loam, fine sandy loam, sandy loam or coarse sandy loam	193
4	Clay, silty clay, non-calcareous clay loam or silty clay loam with more than 35% clay content	193
4L	Calcareous clay loam and silt loam or calcareous clay loam and silty clay loam	193
5	Non-calcareous loam and silt loam with less than 20% clay content or sandy clay loam, sandy clay and hemic organic soils	126
6	Non-calcareous loam and silt loam with more than 20% clay content or non-calcareous clay loam with less than 35% clay content	108
7	Silt, non-calcareous silty clay loam with less than 35% clay content and fibric organic soil material	85
8	Soils not susceptible to wind erosion	0

The wind erodibility index (Table 2.2) is a measure of the mass of sediment eroded from soil containing more than 60% dry aggregates (diameter >0.84 mm) relative to the soil containing other portions of aggregates under the same conditions. A similar classification was developed by Leys (1991b) for Australian soils in western New South Wales. The WEGs are faithful to the definition of erodibility given by Houghton and Charman (1986).

Two additional erodibility ranking systems have been proposed that are of interest here: the ‘Lorikey’ and the Land Erodibility Index (LEI). These provide contrasting approaches for assessing and ranking the susceptibility of soil and land areas to wind erosion.

Lorimer (1985) developed the ‘Lorikey’ to provide a method for assessing the susceptibility of soils and land to wind erosion. The key extended early soil erodibility ranking systems (Marshall, 1973; Kimberlain et al., 1977) and the WEGs. These systems were combined with factors quantifying wind strength, site exposure (topography) and the frequency of erodible winds, as well as soil surface condition, organic matter and texture. The Lorikey does not have a measurable unit like the WEGs or the soil or field erodibility factors (Table 2.1). Like the field erodibility index (K), the Lorikey allows for the assessment of land susceptibility to wind erosion. However, the Lorikey does not include vegetation effects on wind erosion. Application of the Lorikey is therefore restricted to bare (cultivated) agricultural regions.

McTainsh et al. (1999) developed a Land Erodibility Index (LEI). The LEI provides an indication of the relative susceptibility of land types (e.g. downs, dunes and playa) to wind erosion, as well as temporal changes to these in response to climate variability. The LEI is calculated as the dust flux (sediment flux at 0.5 to 2 m) divided by the cube of the mean wind speed above 6 ms^{-1} for the sampling period (monthly or annual). Unlike the soil erodibility factor, WEGs and the Lorikey, the LEI represents an historic account of wind erosion processes and is responsive to changes in soil surface condition, moisture, vegetation cover and windiness.

2.1.1 Temporal Changes in Soil Erodibility

Soil erodibility is not constant but varies through time. Factors controlling soil erodibility include texture, soil moisture and binding agents (both mineral and organic). Temporal variations in soil erodibility are influenced by soil aggregation and crusting which affect the

surface roughness and availability of loose erodible sediment (Zobeck, 1991). Overarching controls on temporal changes in soil erodibility are factors that affect moisture content, aggregation and crusting, and are attributable to climate, land use and land management practices. These are described in detail in Section 2.2. The conceptual basis for soil erodibility to wind therefore differs from the third assumption noted by Bryan et al. (1989), with soil properties governing erodibility in practice being highly responsive to environmental change.

Soil erodibility rankings like the WEGs are static. The position of a soil within a ranking system must be flexible to account for dynamic changes in soil properties. Static soil erodibility classifications like the WEGs are therefore indicative of long-term average, maximum or minimum erodibility conditions. They should not be used to infer the erodibility of a particular soil unless its condition (in terms of crusting and aggregate size distribution) matches that of the same soil group in the WEG classification.

Soil erodibility varies at multiple time-scales as a function of its controlling factors (Geeves et al., 2000). While soil texture is an underlying factor controlling soil erodibility (by its influence on grain binding potential) and varies at very long time scales, soil moisture, surface crusting and aggregation may vary at time scales of minutes to years depending on climate and management conditions. While variations in soil erodibility due to changes in soil surface conditions were acknowledged in early research (e.g. Chepil, 1953), classifications like the WEGs do not reflect the presence of an erodibility continuum.

2.1.2 Assessing Erodibility

There are three general approaches that may be used to assess erodibility. The first approach ranks soil erodibility by measurement of soil physical properties that control the availability of loose erodible material. These properties include the aggregate size distribution and the percentage of aggregates <0.84 mm in diameter (e.g. Chepil, 1954; Leys et al., 1996). The second approach ranks erodibility based on sediment transport rates (mass unit area⁻¹ unit time⁻¹) from a land surface area. These may be measured using sediment traps or by wind tunnel experimentation (e.g. Chepil and Woodruff, 1963; Woodruff and Siddoway, 1965; Leys, 1991b). The latter method relies on the measurement of erosion rates so it is dependent on or relative to a reference wind erosivity. The third approach is based on the physics of

particle entrainment and quantifies susceptibility to wind erosion through the threshold friction velocity (u_{*t}) at which particle entrainment will occur. This method is frequently used in process-based wind erosion modelling (e.g. Marticorena and Bergametti, 1995; Shao, 2000). Computation of u_{*t} to assess erodibility provides a measure that is independent of, but accounts for the dependence of wind erosion on wind erosivity.

Aeolian abrasion processes can affect assessments of soil and land erodibility. This is relevant to assessing erodibility through soil erosion rates, but has not always been considered in these studies. Aeolian abrasion (described in Section 2.2.7) is the process by which fine clay-sized particles are emitted into the atmosphere. The process is a response to saltation (Section 2.2.1), whereby bouncing (saltating) particles and/or particles on the soil surface break down or become detached due to particle impact forces. Soils with high sand content are typically mobile and are considered highly erodible (Table 2.2). However, they do not necessarily produce large quantities of dust. This may be the case if the relative proportion of fine (clay) particles in the soil is low (Pye, 1987). Soils with high clay content may have a low erodibility due to strong inter-particle binding or surface crusting. However, with the introduction of some abrasion mechanism to release the fine particles, clay soils can be significant dust emitters. The presence of abrasion material is governed by soil surface conditions, and the availability of sand size particles, for example from a neighbouring dune source. When assessing erodibility it is important to recognize and separate potential soil mobility measured by erosion rates as distinct from potential dust production induced by aeolian abrasion (e.g. McTainsh et al., 1999 – Land Erodibility Index).

2.1.3 Definitions of Erodibility

The terms ‘soil erodibility’ and ‘land erodibility’ will be used in this thesis to describe the separate but related conditions of soil and land susceptibility to wind erosion. The terms are defined as follows:

Soil Erodibility:

The susceptibility of a soil to mobilisation by wind. Soil erodibility is spatially variable and temporally dynamic, and is related to the availability of loose erodible material (< 0.84 mm) on the soil surface as determined by aggregation (aggregate size distribution and aggregate stability), surface crusting and soil moisture content. Factors controlling soil erodibility are

soil texture (particle size distribution), mineral content, organic/biological content, climate, and land management.

Land Erodibility:

The susceptibility of a land area to erosion by wind. Land erodibility is spatially variable and temporally dynamic. The land area may vary in size from a field or paddock (10^2 to 10^3 m²) to regional scales ($> 10^4$ km²). Land erodibility is a function of soil erodibility with the added effects of non-erodible surface roughness elements (rocks, vegetation, landforms) that influence wind erosivity. Factors controlling land erodibility include those affecting soil erodibility, land type characteristics (vegetation and geomorphology), climate and management. Where non-erodible roughness elements are absent, land erodibility is controlled by soil erodibility.

2.2 Controls on Soil and Land Erodibility

The following sections present a systems analysis, describing the process of wind erosion and the effects of key environmental controls.

2.2.1 Physics of Wind Erosion and Modes of Sediment Transport

Airflow over natural landscapes is generally turbulent. Wind velocity measurements that describe airflow represent time-averaged conditions (Lancaster, 1995). If wind velocity is measured at various heights above a surface a logarithmic profile results, with wind velocity increasing with height away from the surface. Turbulence caused by surface heating and airflow over topographic obstacles may disrupt this profile (Livingstone and Warren, 1996; Sturman and Tapper, 2001). Frictional effects, enhanced by roughness elements on the land surface, reduce the wind velocity close to the surface.

For bare soil surfaces, a number of forces can be described that act on individual soil particles. These forces determine whether or not particle mobilisation and entrainment can occur:

- Drag, lift and the particle moment;
- Gravitational acceleration;

- Particle weight; and
- Inter-particle cohesion.

Figure 2.1 illustrates the effect of these forces on a particle. Airflow around the particles and over the soil surface results in the lift and drag forces. The particle moment represents a combined rotational effect resulting from these forces. A decrease in fluid static pressure (Bernoulli Effect) combined with a steep velocity gradient over the surface determines the magnitude of the lift forces (Lancaster, 1995). The lift, drag and moment forces act in favour of grain entrainment. Opposing these forces are the effects of particle weight and inter-particle cohesion. A number of environmental conditions affect the strength of these factors. These include: soil texture and grain size distribution; grain weight; particle packing density; soil moisture content; soil chemistry; and soil organic matter content. These factors control the susceptibility of a soil to entrainment – the soil erodibility.

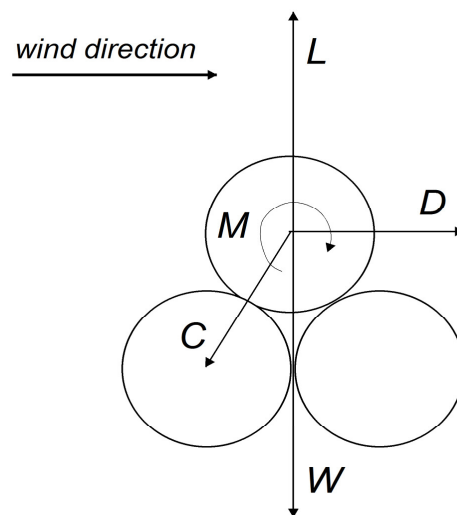


Figure 2.1 Forces acting on soil particles exposed to the air-stream, including lift (L), drag (D), inter-particle cohesion (C), particle weight (W) and the particle moment (M) (after Bagnold, 1941).

In order for wind erosion to occur the forces of lift and drag incident on a particle must exceed the opposing forces of particle weight, inter-particle cohesion and surface friction (Pye, 1987). While turbulent airflow characterises the wind profile, flow close to the surface is also non-linear. Soil particles that protrude into the flow create a layer of zero velocity

called the roughness height (z_0). For bare sandy surfaces this effect can be attributed to the mean particle dimensions:

$$z_0 = \frac{1}{30} D \quad (2.1)$$

where D is the mean diameter of particles on the surface. The distance between particles and other roughness elements will also affect the roughness length (Lancaster, 1995). The shear velocity (u_*) describes the total drag force imparted by the airflow on the roughness elements and surface. The shear velocity is related to wind speed and roughness length by:

$$u_* = \frac{kU}{\ln(z/z_0)} \quad (2.2)$$

where k is the von Karman constant (approximately equal to 0.4); U is the wind velocity at height z , and z_0 is the height at which velocity is zero. When large roughness elements are present (e.g. soil clods, stones, vegetation), the roughness length may be displaced upwards. This new region of zero velocity is called the zero plane displacement height (d), and is a function of roughness element height, density, distribution, flexibility, and permeability. The shear velocity can then be described by:

$$u_* = \frac{kU}{\ln(z-d/z_0)} \quad (2.3)$$

The shear stress (τ) exerted by the wind on the surface can be related to the shear (drag) velocity and to the density of air (ρ_a) by the following expression:

$$u_* = \sqrt{\tau/\rho_a} \quad (2.4)$$

Intuitively, the expression implies that both the shear velocity and shear stress increase as wind velocity (U) increases. As the wind velocity increases, grains on the surface may begin to move. At this point, the shear velocity (u_*) overcomes a ‘fluid threshold’ which defines the

resistance of the surface to mobilisation (Bagnold, 1941). This fluid threshold is described by the threshold friction velocity (u_{*t}):

$$u_{*t} = A \sqrt{\frac{\rho_p - \rho_a}{\rho_a} g D} \quad (2.5)$$

where ρ_a is the density of air; ρ_p is the density of particles; g is the acceleration due to gravity, and D is the particle diameter. A is an empirical coefficient related to Re_p (the particle Reynolds number). The Reynolds number can be described by the shear velocity, particle diameter and kinematic viscosity ν :

$$Re_p = u_* \frac{D}{\nu} \quad (2.6)$$

Re_p provides a measure of turbulence around a particle. Particle movement, and wind erosion, therefore occur when $u_* > u_{*t}$. Of note, is that Equation (2.5) describes the threshold friction velocity in terms of the soil properties of particle diameter and packing density. In practice, this may only be applicable to loose sandy soils. Equation (2.5) does not account for the effects of inter-particle cohesion (by soil moisture or crusting), roughness effects (e.g. by stone cover or vegetation), or how these may vary over soils of different texture.

In addition to wind stress effects, bombardment by grains may also initiate movement of new grains, and results in the ability for sediment movement to be maintained at wind velocities lower than u_{*t} . The lower threshold is the dynamic or impact threshold u_{ti} :

$$u_{ti} = 680 \sqrt{D} \cdot \log\left(\frac{30}{D}\right) \quad (2.7)$$

where D is the mean surface grain diameter. Once the wind friction velocity exceeds the entrainment threshold velocity (i.e. $u_* > u_{*t}$) particle mobilisation will occur. Movement of sediment by wind occurs in a number of modes. These modes are linked and occur during the evolution of the erosion process. The modes include creep, reptation, saltation and suspension, and are illustrated in Figure 2.2. The modes in which particles are moved are a function of grain size and weight, and the wind friction velocity and transport capacity.

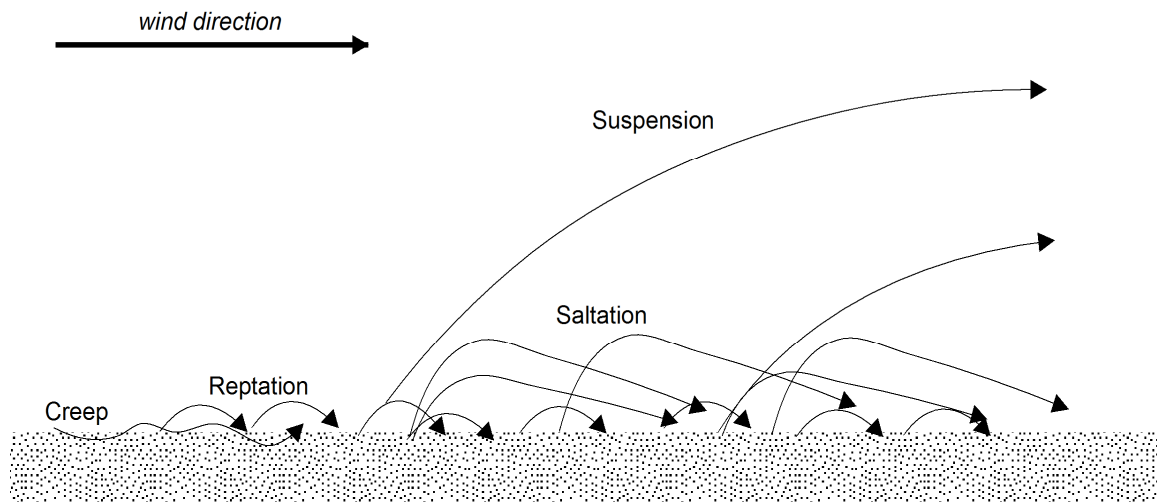


Figure 2.2 Modes of particle transport by wind, including: creep; reptation; saltation; and suspension (after Pye, 1987).

Creep describes the movements of mostly larger grains close to the surface. In particular, creep refers to the rolling motion of particles, and may be induced by impacts from finer saltating grains (Livingstone and Warren, 1996). Reptation is the splashing or low hopping of grains close to the surface. Particles pass between this mode and saltation depending on friction velocity characteristics. Saltation describes the hopping action of grains. Saltation is initiated by particle impacts that eject grains into the airstream. These particles are then carried by the wind before descending back to the surface. Saltation is responsible for mobilising static particles on the surface, and the ejection of fine particles into the air through rupturing and abrasion. Clouds of saltating particles influence surface roughness and the wind friction velocity in a positive feedback called the Owen effect (Gillette et al., 1998). Particles emitted from a surface that are carried by the wind by turbulent motion do so in the suspension mode. In general, grain size decreases with height away from the surface and through these modes of transport.

2.2.2 Climatic Controls on Soil and Land Erodibility

The dominant, regional scale, control on wind erosion is climate. In addition to windiness, sites require a significant moisture deficiency in order to become susceptible to wind erosion. Land areas susceptible to wind erosion are therefore typically located within the world's

dryland environments. These areas include sub-humid, semi-arid, arid and hyper-arid lands. Conditions driving aridity in drylands include (after Thomas, 2000:7):

- Atmospheric stability: drylands are common in the sub-tropical high-pressure belts. These are zones of stable air beneath the descending arm of the Hadley Cell.
- Continentality: increased distance from oceans reduces the frequencies of moisture-bearing weather systems and may enhance atmospheric stability.
- Topography: rain shadows may develop in the lee of mountain ranges, increasing aridity.
- Cold Ocean Temperatures: ocean currents affect atmospheric temperatures, humidity and precipitation, and the effects of this may be continual (e.g. along the west coast of South Africa) or dynamic (e.g. related to the El Niño/Southern Oscillation).

Dryland environments are characterised by low precipitation and high annual evapotranspiration rates. Thornthwaite (1931) developed a precipitation-evaporation ($P-E$) ratio to provide an indicator of regional scale aridity:

$$(P - E) = \sum_{i=1}^{12} \left(\frac{P}{t + 12.2} \right)^{1.11} \quad (2.8)$$

where $(P-E)$ is the precipitation-evaporation ratio, P is the monthly mean precipitation (mm), t is the monthly mean temperature ($^{\circ}\text{C}$) and i is the month. Chepil (1956) used Thornthwaite's (1931) index of aridity to provide a climatic indicator (C) of wind erosion based on windiness and moisture availability, a driver of vegetation cover and soil particle cohesion:

$$C = \frac{V^3}{(P - E)^2} \quad (2.9)$$

where V is the mean annual wind velocity (ms^{-1}) at 9 m, and $(P-E)$ is the index of aridity.

Analyses of dust-storm frequencies in relation to rainfall indicate a non-linear relationship between decreasing rainfall and increasing dust-event frequencies (e.g. Middleton, 1984; McTainsh et al. 1989). The effects of rainfall on land erodibility are seen at multiple spatial and temporal scales. At short time scales (days), rainfall affects the soil moisture content,

increasing inter-particle cohesion and physical crusting. At intermediate time scales (weeks), elevated soil moisture may induce biological crust growth and plant growth. Together these reduce the availability of loose erodible sediment on the soil surface (reducing soil erodibility) and increase the surface roughness and therefore u_{*t} . At longer times scales (months – seasons), rainfall effects on land erodibility are controlled by soil-vegetation feedbacks, with moisture retention in the soil, soil crust condition and surface roughness controlled by vegetation cover levels. In turn, these are controlled by rainfall amounts and frequency. The feedbacks and the capacity of the landscape to retain moisture and sustain vegetation are regulated by climate through evaporation rates. Where potential evaporation exceeds rainfall, the plant available soil moisture will be low, vegetation cover will be low, soil aggregation and crusting will be low, and land erodibility will be elevated.

2.2.3 Soil Texture and Grain Size Effects

Bagnold (1941) used wind tunnel experiments to find values of u_{*t} for a range of soil particle sizes. His results show a decreasing threshold velocity is required for entrainment given a decrease in particle diameter (Figure 2.3).

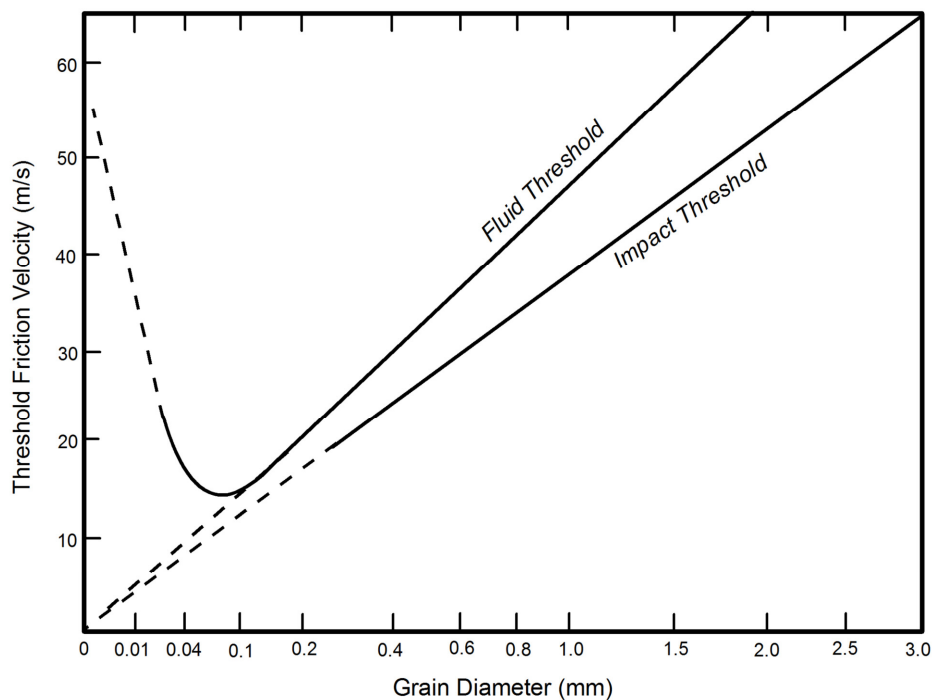


Figure 2.3 Graph of the threshold friction velocities (u_{*t}) for a range of grain/particle sizes (after Bagnold, 1941). The two curves illustrate the difference between fluid and impact thresholds.

The decrease in particle diameter causes a reduction in the roughness length (z_0), and so a decrease in wind speed (U) required to maintain $u_* > u_{*t}$. The threshold friction velocity reaches a minimum when particle diameter is approximately 0.08 mm (very fine sand). After this point, a further decrease in particle size results in an increase in u_{*t} . Bagnold (1941) suggested a mechanism determining this occurrence could be Reynolds number effects (an aerodynamic effect relating to a smoothing surface with decreasing grain size). The relationship between particle size and u_{*t} has been confirmed by a number of authors (e.g. Chepil and Woodruff, 1963; Belly, 1964; Iversen et al., 1976).

Iversen and White (1982) used wind tunnel experiments to further examine the effects of particle size and Reynolds number on u_{*t} under various atmospheric pressure conditions. They proposed that the minima in u_{*t} (Figure 2.3) can be attributed to inter-particle cohesion forces between fine grains. Factors affecting the strength of these forces include moisture films, electrostatic charges, and van der Waals forces. The effects of these factors were not accounted for in Bagnold's (1941) expression to compute u_{*t} (Equation 2.5). A number of studies have sought expressions relating particle cohesion, drag, lift and particle moment forces to u_{*t} (e.g. Iversen et al., 1976; Iversen and White, 1982; Greeley and Iversen, 1985).

Gillette (1988) used wind tunnel experiments to determine the effects of soil texture on u_{*t} . The study reported increases in u_{*t} with increasing soil clay content. This translates to a lower soil erodibility for fine textured soils, however, u_{*t} was found to decrease (increasing soil erodibility) as replicates of soils with cloddy, crusted and loose surface conditions were examined.

Numerous studies have reported the effects of soil texture on wind erosion rates (Chepil, 1953; Lyles, 1977; Gillette, 1978; Leys et al., 1996). In general, these studies have sought empirical relationships between soil clay, sand and silt content and sediment transport at various wind speeds. Chepil (1953) presented one of the first comprehensive studies of this type. The following expression describes the relationship between soil clay content and erosion rates (Chepil, 1953):

$$Q = a.(\%clay)^b . c^{(\%clay)} \quad (2.10)$$

where the Q is the transport rate ($\text{gm}^{-1}\text{s}^{-1}$) and the constants a , b , and c have the values 4.8, -5.1 and 0.09 respectively. The relationship indicates that clay content up to 15% is extremely important in lowering erosion rates. A minimum erosion rate was found at ~27% clay. Soils with higher clay content tended to show an increase in erodibility. Similar expressions were reported by Leys et al., (1996) in a comparison of cultivated and non-cultivated soils. In that study, Leys et al. (1996) found that the empirical constants a , b and c were different for the two soil treatments, reflecting a breakdown in surface structure due to disturbance (see Chapter 4 for details). The effects of this were seen in both increased erosion rates for disturbed soils, and an alteration of the rate of change in erodibility in response to a change in soil clay content (Figure 2.4).

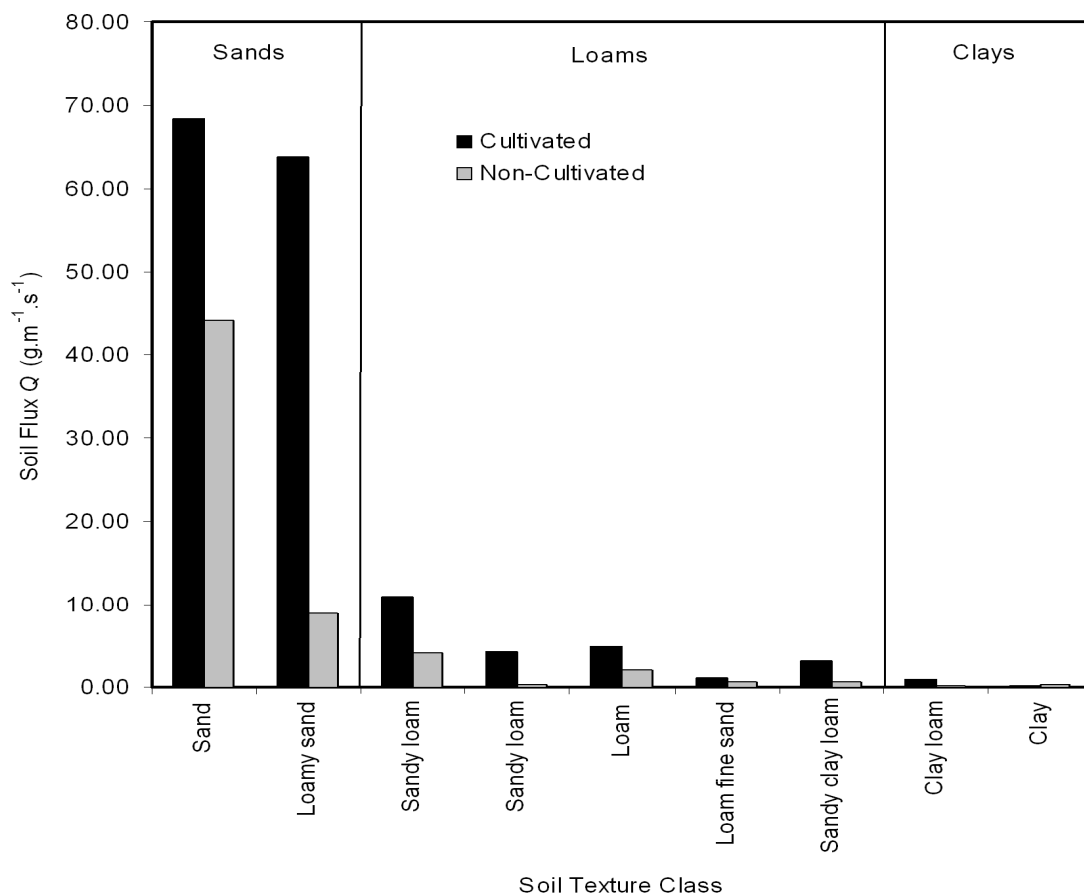


Figure 2.4 Graph illustrating the effect of soil clay content on sediment transport for soils in cultivated (disturbed) and non-cultivated (crusted) conditions (after Leys et al., 1996).

Ultimately, the effect of soil texture on soil erodibility is in determining the potential for coarse grain binding by finer silts and clays (Chepil, 1953). The effect of increasing the

portion of silt and clays in a soil is to increase binding and aggregation of particles, resulting in an increase in the shear stress required to mobilise grains on the surface. The sub-rounded shapes typical of quartz sand grains are not conducive to the formation of inter-particle bonds. This results in a loose and erodible surface. Clays are effective binding agents as their plate-like structures provide large inter-particle contact areas. Clay particles often carry some charge, which allows for electrostatic bonding and moisture retention, which facilitate grain binding (Breuninger et al., 1989). Under extreme drought conditions and disturbance (e.g. by livestock), these inter-particle bonds may be broken down, resulting in cracking and granulation of a clay soil surface (Gillette, 1978). Self-mulching clays are prone to this condition (Leys et al., 1996). Granulation of clay soils may lead to a surface structure with aggregates of a similar size to sand grains, and therefore elevated erodibility. Figure 2.4 illustrates how significant changes in soil erodibility may occur under surface disturbance, i.e. soils with the same textures may have a range of erodibility values.

2.2.4 Soil Aggregation

Soil Aggregate Formation and Breakdown

Soil aggregation is driven by soil texture (sand, silt, clay content), organic matter content, calcium carbonate and salt content, and climate (moisture availability). Perhaps the most important of these is soil texture, with the particle size and mineralogy of soil controlling the ability of soil inter-particle bonds to form and resist destruction. This ability to form bonds is driven by the nature of inter-particle contacts, with a fundamental difference between particle contacts in sandy soil to those in a soil with high clay content (Smalley, 1970). As sandy soils are least susceptible to aggregate formation, they are also the most consistently erodible. Gillette (1978) reported a model to illustrate the dependence of soils on drought to become susceptible to wind erosion (Figure 2.5).

The forces controlling the breakdown of soil aggregates are fundamental in controlling the erodibility of any soil with a non-sandy texture (Breuninger et al., 1989). Mechanisms driving aggregate breakdown include: mechanical disturbance (by animals or machinery), photo-degradation during long periods of exposure (e.g. during drought), clay lattice expansion-contraction during wet-dry cycles, freeze-thaw cycles, direct abrasion, and salt efflorescence (Harris et al., 1966; Merrill et al., 1999; Sarah, 2005).

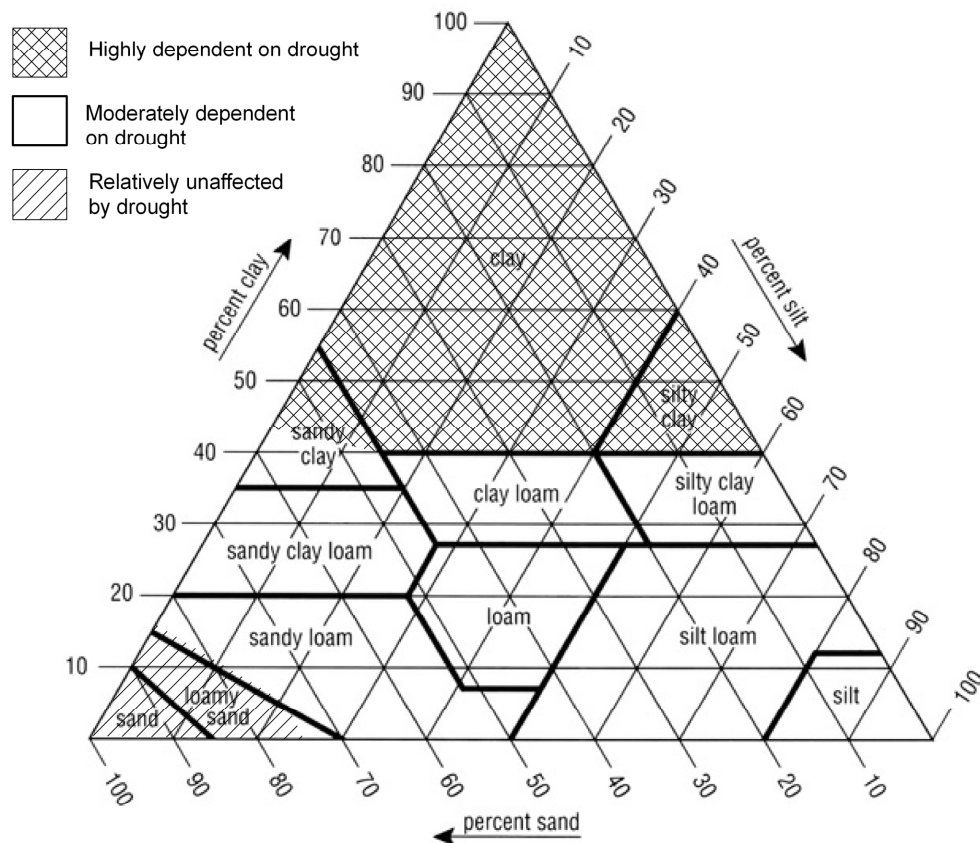


Figure 2.5 Diagram illustrating the dependence of soil textures on drought to experience a significant increase in soil erodibility. The dependence of clay textured soils on drought is due to the requirement for long dry periods that enable the breakdown of aggregates to occur (after Gillette, 1978)

Soil Aggregation Effects on Wind Erosion

Soil aggregation affects wind erosion by increasing the surface roughness length (z_0), decreasing the availability of loose erodible material, and increasing the wind shear velocity required to mobilise grains (u_{*t}). Aggregation also plays a role in fine particle emissions, with aggregate size and stability affecting the impact energy of saltating particles and the potential for abrasion to occur (Zobeck, 1991).

Chepil (1942) and Chepil and Woodruff (1954) reported on the use of dry aggregate structure to rank soil erodibility. They identified that the dry aggregate component of a soil >0.84 mm in diameter is typically non-erodible. This component drives the availability of loose erodible material (grains <0.84 mm in diameter) on the soil surface. Leys et al. (1996) used a portable field wind tunnel to further examine effects of dry aggregation on sediment flux. Sediment

flux was determined for nine soils in cultivated and non-cultivated (crusted and aggregated) states (Section 2.2.3). They found an exponential relationship between percentage dry aggregates (%*DA*) and erosion rate:

$$I_w = 8.33 \exp^{-0.078\%DA} \quad (2.11)$$

where *I_w* is the erosion rate at 65 kmh⁻¹ (tha⁻¹min⁻¹), and %*DA* is the dry aggregation >0.85 mm (%). The relationship holds for a range of soil textures, suggesting %*DA* is more physically related to erosion rates than percentage clay content, which requires one regression equation to define the erodibility of various surface conditions (Leys, et al., 1996). This physical relationship stems from the direct effect of aggregate (grain) size on *u_t* (Equation 2.5). The expression confirmed the results of a similar study by Chepil (1953).

Fryrear et al. (1994) developed a regression model for computing the erodible fraction (*EF*) of soils based on soil texture and chemical data.

$$EF = 29.09 + 0.31(Sand) + 0.17(Silt) + 0.33(SC) - 4.66(OC) - 0.95(CaCO_3) \quad (2.12)$$

where *Sand* is the soil sand content, *Silt* is the soil silt content, *SC* is the ratio of sand to clay contents, *OC* is the organic carbon content, and *CaCO₃* is the calcium carbonate content. An important characteristic of the model is that it was derived from time-averaged data, and does not account for temporal variations in aggregation and the erodible fraction driven by climate variability and land management.

Few published studies have examined temporal variations in soil aggregation in response to climate variations, land use or land management practices. The majority of existing studies have examined the effects of freeze-thaw cycles on soil aggregation in cultivated lands in North America (e.g. Chepil, 1954; Bisal and Ferguson, 1968; Merrill, 1999; Bullock et al., 2001). Despite the global significance of wind erosion in the world's hot, dryland regions, research that addresses temporal changes in soil erodibility in these environments is scarce. While the effects of soil surface conditions on erodibility have been described in some research, for example Gillette et al. (1980) and Gillette et al. (1982), a significant gap remains in our understanding of temporal changes in soil aggregation in rangeland environments.

2.2.5 Soil Moisture Effects

Soil moisture increases inter-particle cohesion and induces soil aggregate and crust formation. Moisture binds soil particles by adhesion and capillary effects (Cornelis and Gabriels, 2003). The increase in particle binding causes a reduction in the availability of loose erodible sediment and an increase in the energy required to mobilize grains on a soil surface (u_{*t}). Soil moisture content therefore plays a critical role in determining the susceptibility of a soil surface to mobilization.

Numerous experimental studies have been carried out to quantify the relationship between soil moisture content and u_{*t} . Early reports on the effect of moisture include those by Chepil (1956), Belly (1964) and Bisal and Hsieh (1966). The research demonstrated a non-linear response of increasing u_{*t} and decreasing soil particle movement with slight increases in soil moisture content for a range of desert soils in North America. Chepil (1956) established an empirical relationship accounting for the increasing bed resistance to shear stress due to soil moisture by the expression:

$$u_{*tw} = \sqrt{u_{*t}^2 + \frac{\gamma_c}{\rho_a}} \quad (2.13)$$

where ρ_a is the air density (kgm^{-3}), and γ_c is the resistance against shear stress due to cohesion between particles (due to the presence of water films between grains), and expressed as:

$$\gamma_c = 0.6 \left(\frac{w}{w_{1.5}} \right)^2 \quad (2.14)$$

where w is the gravimetric moisture content (kgkg^{-1}), and $w_{1.5}$ is the gravimetric moisture content at -1.5 MPa (kgkg^{-1}). Subsequent research has demonstrated that u_{*tw} values calculated by equation (2.14) tend to be higher than those computed using more recently derived expressions. Like Chepil, Belly (1964) used wind tunnel experiments to derive an empirical function for the computation of u_{*t} adjusted for percent soil moisture content (w).

His expression described the wet threshold friction velocity in terms of the influence of moisture content on the dynamic (fluid) threshold friction velocity:

$$u_{*t} = u_{*tw} [1.8 + 0.6 \log(100w)] \quad (2.15)$$

Belly's expression, however, was derived from a single soil type, and did not account for the effects of grain size, shape or sorting on the effect of moisture content. For this reason, the model does not perform well in predicting the effects of moisture content on u_{*t} for soil textures different to that used in the study.

Bisal and Hsieh (1966) used a laboratory wind tunnel to determine the effects of moisture on u_{*t} for the entrainment of fine sandy loam, loam, and clay soils. They demonstrated that sandy soils are least affected by moisture, while loamy and clay soils are more resistant to mobilisation when moist. This is due to the ability of the soils with higher clay content to retain moisture and form aggregates that are less susceptible to mobilization than well-drained sandy soils.

Hotta et al. (1985) developed a linear expression to compute the wet threshold friction velocity based on the dry threshold and soil moisture content. Their expression was based on experimental data for sands with grain sizes ranging from 0.2 to 0.8 mm, although the expression does not account for soil grain size:

$$u_{*tw} = u_{*t} + 7.5w \quad (2.16)$$

where w is the water content in kg kg^{-1} and $w < 0.08 \text{ kg kg}^{-1}$. Hotta et al. (1985) assumed a model to describe the presence of soil moisture between soil particles, whereby moisture is confined to wedges, which can be defined by the contact angle between grains. More recent research into the effects of soil moisture on u_{*t} have used or revised this model of moisture retention and grain binding to develop expressions for u_{*tw} . Further research developing numerical adjustments to u_{*t} as a function of soil moisture have been reported as exponential relationships by Azzizov (1977), Hagen (1984), Shao et al. (1996) and Chen et al. (1996).

Saleh and Fryrear (1995) identified a number of deficiencies in the early studies of moisture effects on particle mobilization by wind. A particular limitation is that the expressions do not account for processes that affect particle entrainment once u^* exceeds u_{*t} and the saltation process is established. Subsequently, two expressions were developed for abrading (2.17) and non-abrading (2.18) surface conditions where:

$$u_{*ta} = 0.205 + 0.182(TWC/W') + 0.375(TWC/W')^2 \quad (2.17)$$

$$u_{*t} = 0.305 + 0.222(TWC/W') + 0.506(TWC/W')^2 \quad (2.18)$$

where u_{*t} is the threshold friction velocity under non-abrading conditions, u_{*ta} is the threshold friction velocity under abrading conditions, TWC is the soil surface threshold water content (the moisture content at which particle entrainment is initiated), and W' is the gravimetric water content of the soil. Their results suggest that under abrading conditions, because of the erosive impact of saltating particles, less wind energy is required to initiate entrainment.

Recent developments in quantifying the effects of soil moisture on the process of soil mobilization by wind have combined wind tunnel experimentation with conceptual models of inter-particle cohesion. The focus of the research moved from trying to develop regression equations to quantifying the binding forces responsible for inter-particle cohesion. McKenna-Neuman and Nickling (1989) presented a theoretical model for the effect of moisture on u_{*t} . The research developed the notion that inter-particle contacts can be modelled by accounting for capillary forces between particles with angular contact geometries. The basis for the model, the inter-particle capillary force (F_c) is defined by:

$$F_c = \frac{\pi T^2}{P} G \quad (2.19)$$

where T is the surface tension of the water, P is the pressure deficiency, and G is a non-dimensional coefficient. This function was incorporated into Bagnold's (1941) static threshold model (Equation 2.5). Two expressions were developed to account for open-packed (Equation 2.20) and closed-packed grains (Equation 2.21):

$$u_{*tw} = u_{*t} \sqrt{\frac{6 \sin 2\beta \cos \beta}{\pi d^3 (\rho_s - \rho_a) g \sin \beta} F_c + 1} \quad (2.20)$$

where the particle resting angle is defined by β and is approximately equal to 30° , and

$$u_{*tw} = u_{*t} \sqrt{\frac{6 \sin 2\beta (2 \cos \beta + 1)}{\pi d^3 (\rho_s - \rho_a) g \sin \beta} F_c + 1} \quad (2.21)$$

where β is approximately 45° , ρ_a is the air density and ρ_s is the particle density (kgm^{-3}), g is gravitational acceleration (ms^{-2}), and d is the mean particle diameter (m).

The model was calibrated by wind tunnel experimentation with a range of sand particle size groups, but was not tested for soils with high clay content. The model performance decreases when adhesive forces become the primary forces holding water between particles, and capillary forces, for which the model was developed to describe, become less important. Based on this outcome, McKenna-Neuman and Nickling (1989) proposed that tension may be a better parameter to describe the effects of soil moisture than gravimetric moisture content as it is more directly related to inter-particle cohesion.

Gregory and Darwish (1990) reported a theoretical model to calculate soil moisture effects on u_{*t} . The model was developed from wind tunnel experimentation and data presented in the literature by earlier studies and uses a simpler model for grain shape and packing than the McKenna-Neuman and Nickling (1989) model, using spheres rather than cones to approximate grain packing angles. The following expression was obtained:

$$u_{*tw} = u_{*t} \sqrt{1 + w + \frac{6}{\pi} \frac{a_1}{\rho_s g d^4} + \frac{a_2}{\rho_w g d} \exp^{-a_3 \frac{w}{w_{1.5}}} (w - w_c)} \quad (2.22)$$

where ρ_w is the density of water (kgm^{-3}), w_c is the moisture content attached to clay particles or aggregates (kg kg^{-1}), and a_1 , a_2 and a_3 are coefficients (a_1 in kgs^{-2} , a_2 in $\text{kg m}^{-1}\text{s}^{-2}$ and a_3 is dimensionless). The value for a_1 was derived from data reported by Greeley and Iversen (1985), while the values for a_2 and a_3 were developed from data presented in studies by Belly (1964), Bisal and Hsieh (1989) and McKenna-Neuman and Nickling (1989). As the contact

geometry between spherical grains (one particle on top of another) was used, no trigonometric correction was required for the direction of the cohesive force as in the McKenna-Neuman and Nickling (1989) model. The model performed well for low soil moisture, but does not hold true as moisture content increases and pore spaces between grains begin to fill.

Fécan et al. (1999) developed a theoretical model to account for soil moisture effects on u_{*t} using a similar model for particle contacts (i.e. between cones) and grain binding (i.e. by capillary forces) as McKenna-Neuman and Nickling (1989). Their model was simplified to directly include soil moisture content and an expression to calculate the threshold moisture content at which entrainment is initiated. The model was developed from a regression analysis of data presented by Belly (1964), Bisal and Hsieh (1966), Saleh and Fryrear (1995) and Chen et al. (1996), and is of the form:

$$w' = 0.0014(\%clay)^2 + 0.17(\%clay) \quad (2.23)$$

where w' is the threshold moisture content for particle entrainment. The expression is used in the calculation of the wet threshold friction velocity for two scenarios:

$$\begin{aligned} \frac{u_{*tw}}{u_{*td}} &= 1 & \text{for } w < w' & \quad (2.24) \\ \frac{u_{*tw}}{u_{*td}} &= \sqrt{1 + 1.21(w - w')^{0.68}} & \text{for } w > w' & \end{aligned}$$

where u_{*tw}/u_{*td} is ratio of the wet to dry threshold friction velocity.

Cornelis and Gabriels (2003) tested a number of the models described above, finding large differences between output values for set input conditions. To address this issue, Cornelis et al. (2004a) developed another expression to quantify the effects of soil moisture on u_{*t} . Developments in the new model included: 1) it considers inter-particle forces as the sum of dry and wet binding forces, rather than computing the wet threshold friction velocity as an adjustment to the dry threshold; 2) it takes into account both capillary forces between soil particles and adhesive forces due to absorbed water films; and 3) it uses a more general

geometry for particle shapes than cones. The outcome was a model that could be applied to determine the threshold friction velocity of both dry and wet sediment:

$$u_{*tw} = A \sqrt{\frac{\rho_s - \rho_f}{\rho_f} g d} \quad (2.25)$$

where u_{*tw} is the threshold friction velocity for entrainment as affected by near surface soil moisture, ρ_s is the particle density (Mgm^{-3}), ρ_f is the fluid density (Mg m^{-3}), g is the gravitational acceleration (ms^{-2}), d is the particle diameter (m). A defines the effect of moisture on the threshold by the expression:

$$A = \sqrt{A_1 \left[1 + w + A_2 \frac{1}{(\rho_s - \rho_f) g d^2} \left(1 + A_3 \frac{\sigma^2 d}{|\psi_{md}| \exp^{-6.5 \frac{w}{w_{1.5}}}} \right) \right]} \quad (2.26)$$

where A_1 , and A_2 are coefficients associated with aerodynamic and inter-particle forces between dry particles, σ is the surface tension of the liquid (Nm^{-1}), ψ_{md} is the matric potential at oven dryness, w is the gravimetric water content (kgkg^{-1}), and $w_{1.5}$ is the gravimetric water content at -1.5 MPa (kgkg^{-1}). The coefficient A_3 is associated with the effects of soil moisture through capillary and adhesive forces, and is determined by curve fitting to experimental data depending on soil particle shape and dimensions. Cornelis et al. (2004b) calibrated the model using wind tunnel measurements of u_{*t} at a range of soil water contents then compared model predictions with those of Chepil (1956). Good agreement was found between the models, although discrepancies were found between these models and those of Azzizov (1977), Hotta et al. (1985) and Chen et al. (1996).

Limitations of the models are that they have been developed for limited particle size and soil texture ranges. Model accuracy also tends to decline outside the ranges of data from which they were developed. Cornelis and Gabriels (2004a) reported that inconsistencies in the methods and timing used in previous research to identify the moment of particle mobilization has created differences in the model performance. Further, under moist conditions surface particles may rapidly dry under high wind velocities, dropping below the entrainment threshold and mobilizing for a moment until particles are removed and the moist surface

(above threshold) is once again exposed. This has implications for modelling the effects of soil moisture on wind erosion rates, as changes to the moisture content in the topmost exposed surface particles may occur (moving to an erodible state) that are not detectable by most measurement or modelling approaches.

A number of the moisture models have been implemented in wind erosion models, including in EPIC (Chepil, 1956), AEOLUS II (Belly, 1964), TEAM (Gregory and Darwish, 1990), WEPS (Saleh and Fryrear, 1995), DPM (Fécan et al., 1999), and WEAM and IWEMS (Shao et al., 1996). While gravimetric moisture content can easily be measured in the field, and has been used in the early empirical regression functions, the use of volumetric moisture content in later models reflects developments in integrated modelling approaches that couple wind erosion prediction systems with models of land surface and soil conditions (Shao, 2000).

2.2.6 Surface Crusting and Disturbance

Crust Types, Formation and Breakdown

Surface crusts form as a result of the binding of soil particles at or near the soil surface. Soil surface crusting may be driven by physical processes or biological activity. Both physical and biological crusts play an important role in controlling soil erodibility.

Two main types of physical crust exist. These are identified by their mechanisms of formation. The first, structural crusts, form as a result of water droplet impact (e.g. during rainfall) and *in situ* particle rearrangement. The second, depositional crusts, form by deposition of fine particles transported from some source (Valentin and Bresson, 1992). Both structural and depositional crusts are associated with a number of sub-types that reflect more specific processes during formation. Sub-types of structural crusts include: slaking, infilling, coalescing and sieving crusts. Each of these is characterized by a specific lateral structure resulting from the factors driving formation. Sub-types of depositional crusts include: runoff depositional crusts, still depositional crusts, and erosion crusts (Valentin and Bresson, 1992). The formation of a particular crust type is dependent on the characteristics of precipitation (i.e. amount, frequency and intensity), soil texture, and local topography. Soil chemistry, for example CaCO_3 and salt content, and organic matter content have also been identified as key determinants in physical crust formation (Gillette et al., 1980, 1982).

A number of microbiotic (biological) crust types exist. These are characterized by the biological organisms driving crust formation. Organisms associated with microbiotic crust formation include: mosses and liverworts, lichens, fungi, algae, cyanobacteria, and other microbiota. Microbiotic organisms facilitate crust formation by binding soil particles together through secretion of gels or binding with structural filaments. Crust classification systems have been proposed based on surface micro-topography of the crusted soil (Eldridge and Greene, 1994; Johansen, 1993). These include: smooth, rugose, rolling, and pinnacle topographies. These surface structures generally reflect the climate of the region of occurrence. For example, smooth crusts are associated with hot hyper-arid habitats, whilst pinnacled crusts are associated with cold climates in which frost heaving occurs (Belnap et al., 2001a). As for physical crusts, climate, soil texture, and soil chemistry play important roles in biological crust formation and structural characteristics. The presence and type of vegetative cover also affect biological crust formation. Vegetation affects the ability of precipitation to reach the soil (or crust) surface, and in a species-species competition sense (Belnap et al., 2001b).

Physical and microbiotic crust formation are particularly sensitive to climate and soil textural characteristics. The amount, intensity, and frequency of precipitation determine the type of physical crust formation, and where present, biological crust growth or degradation (Belnap and Gillette, 1998). In addition to these factors, the precipitation efficiency is regulated by solar radiation intensity and potential evaporation. These latter factors influence the rate of soil particle transport, deposition and binding, and the ability of microbiotic organisms to utilise the moisture, or in fact survive close to the soil surface. Soil texture, in particular sand and clay content, have been found to affect the soil particle binding, and habitat suitability for microbiota (Diouf et al., 1990; Skidmore and Layton, 1992). Physical crusts do not form readily in sandy soils due to a lack of fine particles required for inter-particle binding. Loamy and clay textured soils, however, provide conditions suitable for inter-particle binding by physical (structural and depositional) processes, and stable habitats for crust forming microbiota.

While crust formation may be triggered by precipitation, crust degradation or breakdown occurs as a result of photo-degradation, burning, structural breakdown (e.g. in self-mulching soils), trampling by livestock, mechanical disturbance, and in the case of biological crusts, death (Eldridge and Green, 1994; Belnap and Eldridge, 2001). The extent and intensity of

disturbance, season, antecedent climatic conditions and the state of a crusted surface at the time of disturbance are vital in determining the level of crust degradation suffered with a particular disturbance mechanism (Marble and Harper, 1989; Scarlett, 1994). Spatio-temporal variability in climate and disturbance regimes leads to variable and spatially heterogeneous physical and biological crust cover in many landscapes. Transitions between physical and biological crust types may result from disturbance-recovery cycles, and this has implications for wind erosion processes (Strong, 2007).

Soil crust characteristics that influence soil susceptibility to wind erosion include cover, structure (arrangement of particles by grain size), thickness, strength, and modulus of rupture (resistance to breaking by saltating particles). Spatial heterogeneity in these characteristics, combined with disturbance mechanisms also affects the crust influence on wind erosion.

Crust Effects on Wind Erosion

Soil crusts reduce soil erodibility by binding soil particles (locking them into a non-erodible surface layer), reducing the availability of loose erodible sediment, and increasing surface roughness. Thick surface crusts with a high modulus of rupture can provide a non-erodible armour over a surface.

In general, an increase in surface crust cover and strength results in an overall decrease in soil erodibility. Chepil (1942) reported that crusted soils may erode at a rate about one-sixth that of non-crusted soils. This relationship was later determined to be more complex. Leys and Eldridge (1998) reported crust effects on wind erosion under three levels of simulated disturbance on a range of soil textures. They found that on the sandy soils, erosion rates were statistically the same under the three disturbance levels; however, for the loamy soils there was a significant increase in erosion rate with increasing disturbance. Rajot et al. (2003) reported a 75% decrease in loose erodible particles on the soil surface due to crust formation following rain. Various other studies have reported the effects of soil properties on crust formation and its effects on wind erosion (e.g. Belnap and Gillette, 1997, 1998; Goosens, 2004; Langston and McKenna-Neuman, 2005; Thomas and Dougill, 2007). In general these studies have examined the effects of disturbance (i.e. by livestock or mechanical processes) on crust cover and erosion rates.

Rice et al. (1996) and Rice et al. (1997) explored relationships between crusting and particle binding strength and erosion due to particle impacts (by saltation bombardment). They presented a conceptual model to define erosion rates based on probability distributions of particle impact energy and surface strength (Figure 2.6). This work was followed by studies seeking to quantify the effects of abrasion on crusted surfaces and resulting dust emission rates (McKenna-Neuman and Maxwell, 1999; Houser and Nickling, 2001a, 2001b).

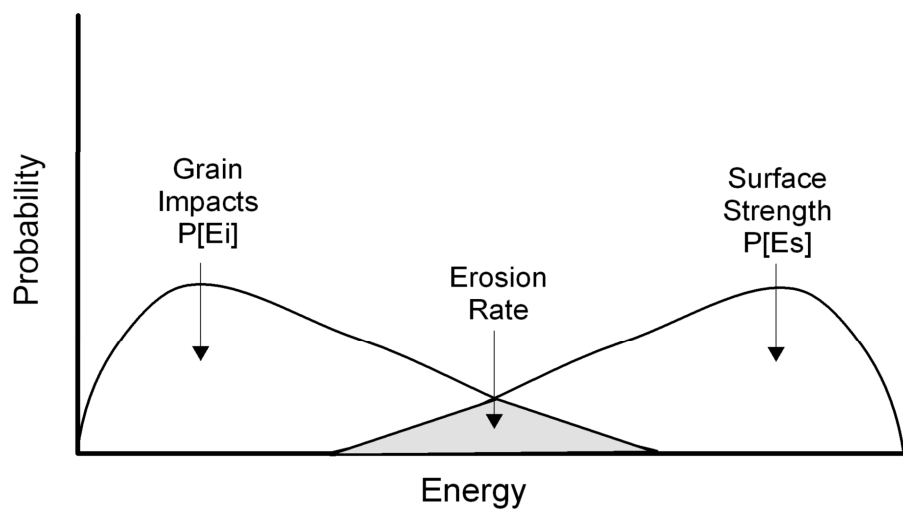


Figure 2.6 Illustration of probability distributions of the energy delivered to a soil surface by saltating grains, $P[E_i]$, and the local energy required to break surface crusting, $P[E_s]$ (after Rice et al., 1999).

Eldridge and Leys (2003) demonstrated that there is a strong relationship between crust cover and dry aggregate size distribution, for which there is an established functional relationship with wind erosion rates (Equation 2.11). Their results suggest that aggregation is a better predictor of erodibility than crust cover alone. Fryrear et al. (1998) developed a model to predict crust cover based on soil clay, CaCO_3 and organic matter contents. The empirical nature of the model, and the complex relationship between climate, soil properties and crust formation mean that the model has limited application outside the limits of the data from which it was derived. Accounting for the effects of soil surface crusting and aggregation in wind erosion models has, and continues to be, a global problem. The implications of the spatial variability in crust properties, and with time, dictate that further research is required, particularly in monitoring the temporal evolution in surface crusting and its effects on erodibility. This issue is described in detail in Chapter 4.

2.2.7 Dust Emission by Aeolian Abrasion

The entrainment of particles from an exposed soil surface takes place through two processes. These processes occur primarily as a result of impacts by saltating grains and include: 1) deflation, whereby incoherent sediment is entrained by lift and drag forces or are “splashed” into the air stream by ballistic particle impacts; and 2) abrasion, whereby consolidated, cohesive materials including both mobile aggregates and crusted surfaces are worn down (Livingstone and Warren, 1996).

Significant dust emission is dependent on a supply of loose sediment that can act through abrasion to release fine particles (Houser and Nickling, 2001a). This may occur as clay or iron oxide coatings are worn from sand grains, or as clay particles are released from a crusted surface. While sandy soils tend to have loose and mobile surfaces, soils with high clay content tend to be supply limited in particles that can saltate, abrade and emit dust. Supply limitation of particles that can abrade exposed soil surfaces results from aggregation and crusting, while an increase in supply may result from photo-degradation, mechanical breakdown and disturbance of aggregates and crusts.

The supply of sediments that can abrade a surface is critical to determining the susceptibility of a surface to wind erosion (Gillette, 1977; Hagen, 1991; Zobeck, 1991; McKenna-Neuman et al., 1996; Rice et al., 1996, 1997, 1999). Abrasion efficiency is also related to the size of saltating grains, their kinetic energy, and the physical characteristics of the surface, including crust presence, strength and modulus of rupture (Greeley et al., 1982). Houser and Nickling (2001a, 2001b) reported that abrasion efficiency is inversely related to average crust strength by a power function, implying that as crust strength increases it becomes more difficult for saltating grains to abrade a surface.

2.2.8 Roughness Effects of Vegetation

Non-erodible elements like vegetation cover affect wind erosion in a number of ways. These effects are manifested through interactions with the air stream and sheltering of the soil surface. Vegetation affects the wind shear velocity (erosivity), soil properties (e.g. organic matter content) controlling crust formation and aggregation (erodibility), and interacts with the processes of particle entrainment, transport and deposition. The effects of vegetation on

wind erosion are influenced by the nature of the cover. In particular, if the cover is prostrate (flat), or standing (Leys, 1991a).

A few approaches have been taken to quantify the effects of vegetation cover on wind erosion. These include: 1) studies to determine regression functions to describe the relationship between non-erodible roughness cover and erosion rates (soil loss); and 2) studies that seek to quantify the relationship between surface roughness and u_{*t} .

Prostrate non-erodible roughness elements reduce wind erosion by three mechanisms (Hagen, 1996). These include: 1) increasing the static threshold at which entrainment begins (u_{*t}); 2) restricting the wind transport capacity by increasing the dynamic threshold for saltation; and 3) increasing the distance required to attain transport capacity in short fields. In addition to providing direct surface protection (cover), vegetation may trap creeping, saltating or suspended particles, thereby reducing total potential soil loss relative to bare erodible surfaces.

Research into the effects of prostrate vegetation cover has been dominated by wind tunnel and field scale (10^3 m^2) experiments in cultivated environments. Early research sought to establish threshold cover levels for prostrate plant litter and stubble that could be used to control wind erosion in agricultural settings. For example, Chepil (1944) used laboratory wind tunnel experimentation with straw residue. He found a negative exponential relationship between surface residue (weight) and soil erosion rates (Q) for any given wind velocity (U) and for a range of soil types. Siddoway et al. (1965) and Lyles and Allison (1981) found similar relationships for Q with respect to various types, amounts and orientations (prostrate and standing) of stubble.

Fryrear (1985) derived a negative exponential relationship between erosion rates (soil loss ratio) and percentage soil cover. He defined the soil loss ratio (SLR) as the soil loss from covered soil to that of the soil in a bare condition, and relationships were determined for dowel and wheat stubble elements and a range of soil textures:

$$SLR = \alpha \exp^{-\beta(F_c)} \quad (2.27)$$

where SLR is the soil loss ratio, F_c is the percent soil cover by non-erodible elements, and α and β are regression coefficients describing the intercept and exponent respectively. Findlater et al. (1990) identified a limitation in Fryrear's SLR , noting that as surface cover approaches 0, the SLR approaches a value of 1.8. They developed a variant on the SLR with a maximum of 1, and presented a revised model of the form $SLR = \exp(-\beta F_c)$. The model performed well in a comparison using data from Fryrear (1985) and portable field wind tunnel experimentation. Leys (1991a) found a similar exponential relationship between cover and erosion rates (Figure 2.7), but reported further limitations with the SLR . The limitations are that: the model never falls to 0 no matter how high surface cover is, and secondly, the model gives the same SLR for all wind velocities. Leys (1991a) then presented an approach relating soil flux (Q) to surface cover using recurrence intervals for wind velocities to determine the probability of wind erosion.

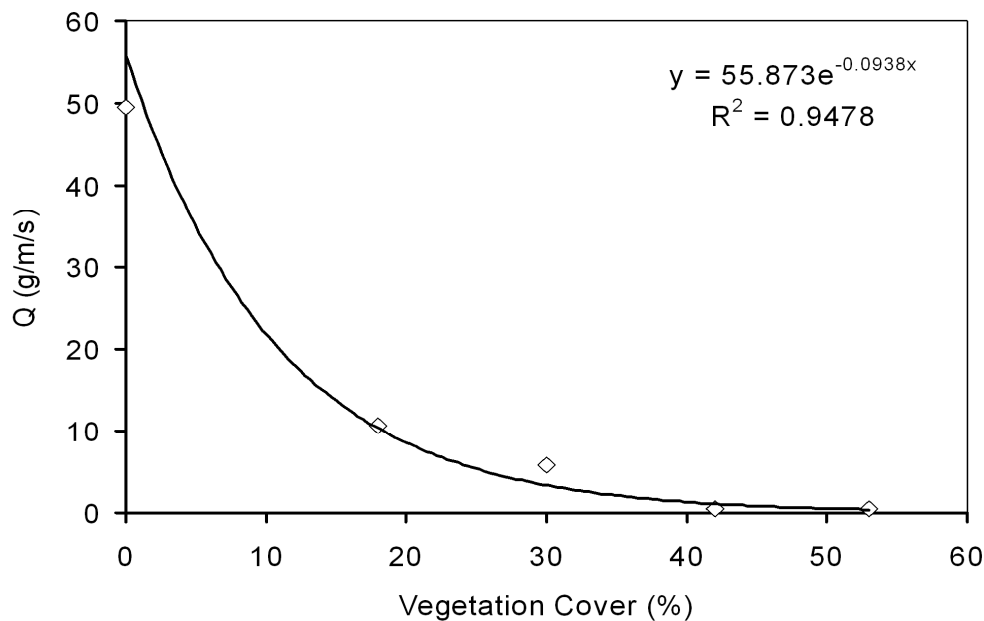


Figure 2.7 Graph illustrating the effect of wheat stubble cover on sand discharge (after Leys, 1991a).

Armbrust and Bilbro (1997) developed an approach to account for stem area, leaf area, and canopy cover of growing crops on the SLR , transport capacity and u_{*t} . They used a single parameter expression for the SLR , with surface cover being replaced by a plant area index (PAI - a composite of stem and leaf area indices). As the SLR is a function of wind velocity, a model was derived to determine the reduction in transport capacity (R) due to surface cover and as a function of wind speed:

$$R = [C_{env} / (C_{env} + T)] (Q_{cv} / Q_{cb}) \quad (2.28)$$

where C_{env} is an emission coefficient for a vegetated surface, and is a function of plant stalk diameter (mm), stalk height (m), and silhouette area (SAI) of stalks and stems per unit ground area. T is an interception coefficient calculated as the ratio of SAI to plant height, Q_{cv} is the saltation discharge transport capacity of the vegetated surface, and Q_{cb} is the measured soil loss from vegetated trays (determined by wind tunnel experimentation). While the model accounts for changes in SLR with wind speed, the input parameters to the model are not easily measured, so the model does not work well outside the limits of the experimental data from which it was derived. Ash and Wasson (1983) and Wasson and Nanninga (1986) demonstrated that wind erosion could occur over surfaces with as much as 45% surface cover. Their results showed how the relationship between vegetation cover and wind erosion varies with wind speed (Figure 2.8). Similar results have been obtained by Hagen (1996), Hagen and Armbrust (1996), and Lancaster and Baas (1998).

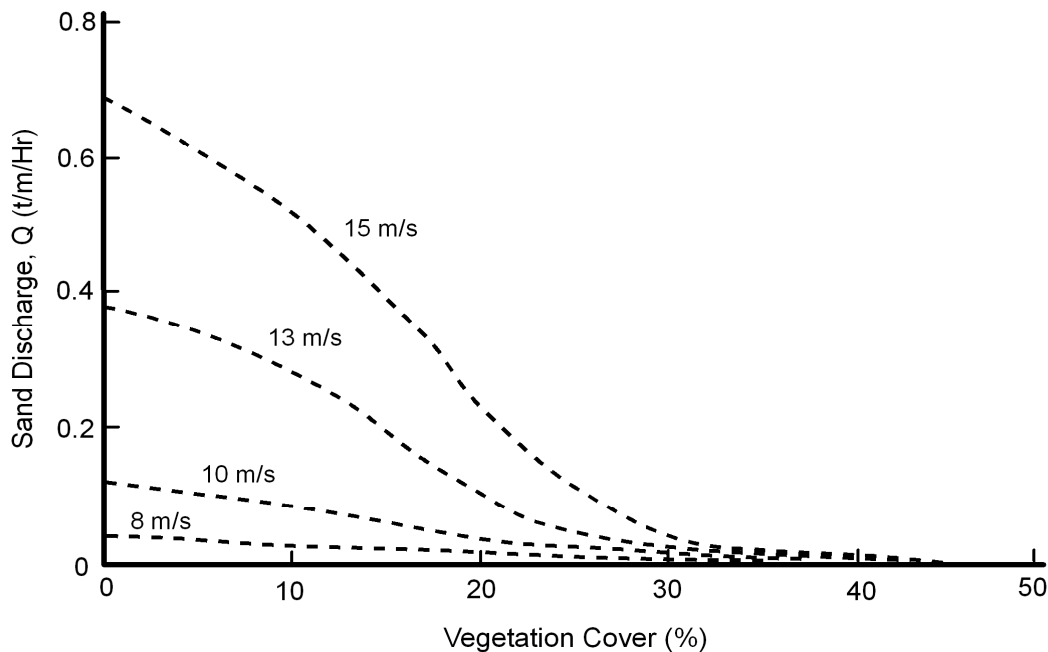


Figure 2.8 Graph illustrating the effect of Spinifex (*Triodia* spp.) grass cover on sand discharge for a range of wind velocities (after Wasson and Nanninga, 1986).

Standing vegetation affects wind erosion by reducing u^* close to the soil surface (Hagen, 1996). This occurs as a portion of the total shear stress exerted by the wind becomes absorbed by standing non-erodible elements such as trees, shrubs and grasses. Wind energy is roughly

proportional to the cube of its speed (Bagnold, 1941). Therefore, a slight decrease in wind speed will result in a significant reduction in its energy and capacity to erode (Liu et al., 1990). Factors determining the degree of protection or potential momentum reduction afforded by standing vegetation include vegetation or non-erodible element size, geometry, spacing (density), lateral cover, flexibility and porosity. In line with this momentum reduction, vegetation displaces the surface roughness length (Equation 2.3). This displacement shelters the soil surface in the lee of elements, and increases boundary layer turbulence (Shao, 2000).

Chepil (1950b) and Chepil and Woodruff (1963) introduced the critical surface constant to model the effects of roughness elements, including both soil surface roughness and standing cover. They reported that the relationship between roughness element height (H) dividend by the distance between elements (d) was constant at the point at which wind erosion is controlled by roughness. Lyles and Allison (1981) found that this was not in fact constant, but changed as a function of u^* . They determined that as surface roughness increases the surface stress absorbed by the roughness elements also increases. Marshall (1971) and Marshall (1972) further examined the effects of roughness element density and distribution on surface drag. The concepts explored in this work led to the development of schemes to adjust u_{*t} for bare surfaces to account for the presence of non-erodible roughness elements.

Gillette et al. (1989) presented a method to quantify the effect of surface roughness on the u_{*t} through the threshold friction velocity ratio $R_t = u_{*tS}/u_{*tR}$. The method computes the ratio of the threshold friction velocity of a bare surface (u_{*tS}) to that of one covered with roughness elements (u_{*tR}). Like the *SLR*, their model decreases from 1 as roughness increases over a bare surface. Raupach (1992) and Raupach et al. (1993) developed this model to predict R_t based on shear stress partitioning between the roughness elements and the surface. The premise of the model was that provided by Marshall (1971), that “*the attenuating effect of roughness on erosion is closely related to momentum absorption by roughness, which is closely controlled by the frontal area...of the roughness elements*” (Raupach et al., 1993: 3023). The frontal area index (λ) is defined by the expression:

$$\lambda = \frac{nbh}{s} \quad (2.29)$$

where n is the number of roughness elements, b and h are mean width and height of the roughness elements (giving frontal area bh), and s is the surface area. The method uses the threshold friction velocity approach, whereby the total shear stress (τ) on a land surface can be partitioned into that which is incident on roughness elements (τ_R), and that which is incident on the substrate surface (τ_S). From Equation 2.4, this can be presented as:

$$\tau = \rho u_*^2 = \tau_R + \tau_S \quad (2.30)$$

where ρ is the air density, and u_* is the friction velocity. Figure 2.9 illustrates the drag partitioning model for a bare surface (a), surface covered with sparse vegetation (b), and a densely vegetated surface (c). The underlying assumption in this theory is that the surface stress deficit behind isolated roughness elements can be described by an effective shelter area. This shelter area can be characterised by the geometry of the elements and bulk flow properties (Raupach et al., 1993). Importantly, the model allows for the effective shelter areas in the lee of non-erodible elements to be superimposed.

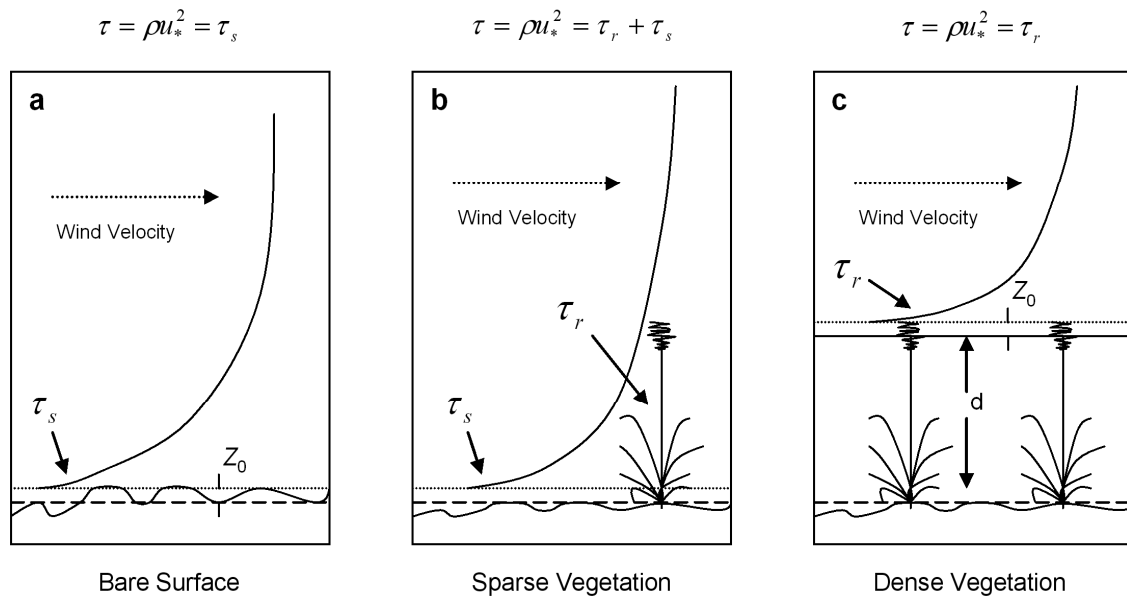


Figure 2.9 Illustration of the effects of vegetation on surface roughness and the drag partitioning model (after Chepil and Woodruff, 1963). (a) shows the relationship for a bare surface, (b) for a surface with sparse vegetation cover, and (c) for a densely vegetated surface.

Following from Equation 2.30, the total threshold shear stress (τ) on a surface covered with roughness elements, and the threshold shear stress (τ_S) of a bare surface can be defined by:

$$\tau = \rho u_{*tR}^2 \quad \text{for a surface with roughness elements} \quad (2.31)$$

$$\tau_s = \rho u_{*tS}^2 \quad \text{for a bare surface}$$

where u_{*tR} and u_{*tS} are the threshold friction velocities for rough and smooth surfaces respectively. The threshold friction velocity ratio (R_t) can then be computed as:

$$R_t = \frac{u_{*tS}}{u_{*tR}} = \sqrt{\frac{\tau'_s}{\tau}} = \sqrt{\frac{1}{(1 - \sigma\lambda)(1 + \beta\lambda)}} \quad (2.32)$$

where the factor $(1 + \beta\lambda)^{-1/2}$ accounts for the reduction in surface stress by shelter from roughness elements, and $(1 - \sigma\lambda)^{-1/2}$ accounts for the “amplification” of the stress (τ 's) on the erodible surface over the difference between total stress and that acting on the roughness elements ($\tau - \tau_R$) caused by the fraction of soil surface ($\sigma\lambda$) occupied by the roughness elements. This formulation was then modified to account for the fact that u_{*t} is controlled by the maximum stress (τ' 's) acting at any point on the surface, rather than spatially averaged stress (τ 's). The modification was thus designed to account for non-uniformity around non-erodible roughness elements, and is implemented by the inclusion of the m parameter (where $m \leq 1$). Written for the computation of the threshold friction velocity over vegetated surfaces (u_{*tR}) as an adjustment to the threshold friction velocity of a smooth surface, we have:

$$u_{*tR} = u_{*tS} \sqrt{(1 - m\sigma\lambda)(1 + m\beta\lambda)} \quad (2.33)$$

Key simplifications in the model are that: 1) surfaces are only described in terms of the roughness density, and the model does not account for differences in element shape; and 2) the model does not account for the presence of turbulent air flow close to the surface. Limitations identified in the drag partitioning scheme revolve around defining appropriate values for the m parameter, the applicability of the m parameter itself, and the roughness density range for which the model is valid $\lambda \leq 0.1$ (Okin and Gillette, 2004). The drag partitioning scheme presented by Raupach (1992) and Raupach et al. (1993) was revised by Yang and Shao (2005) to account for multiple drag partitioning (e.g. between roughness layers at different heights), and for roughness densities where $\lambda \geq 0.1$.

Okin (2008) noted that the m parameter (Equation 2.33) accounts for the spatial variability of shear stress acting on the land surface due to the distribution of vegetation. However, the parameter cannot be defined from a physical basis. While the model accounts for the amount of lateral cover on the surface (Equation 2.29), it does not account for the cover distribution. Vegetation cover in arid and semi-arid landscapes is rarely uniform. Rather, it is patchy and has an anisotropic distribution. Okin (2008) proposed a new model that explicitly accounts for the distribution of shear stress on the surface. The model considers the distribution of vegetation using a factor describing the probability distribution of the spacing between vegetation elements. In doing this the model scale is specified to the plant-interspace/gap scale. Unlike the Raupach et al. (1993) scheme, this means that sediment flux predictions can be more accurately scaled from single points to the landscape scale.

Stone cover (reg) in gibber plains or deserts may be an effective erosion control. Dust emission from land areas with dense stony cover is generally limited (Pye, 1987). While stone cover may not be as effective as some vegetation in increasing the roughness length over a surface, it certainly provides armouring and protection of fine particles from potentially erosive winds. As for vegetation, the effectiveness of stone cover in reducing wind erosion is dependent on the stone dimensions, cover and distribution.

2.3 Anthropogenic Interactions with Wind Erosion Controls

Anthropogenic interactions with the environment affect wind erosion processes. These interactions are greatest in cultivated and rangeland environments. Here, the way in which landscapes are managed affects the spatial and temporal dynamics of variations in wind erosion controls, and therefore land erodibility. Tegen and Fung (1995) suggest that approximately half of the current atmospheric dust loading can be attributed to anthropogenic disturbance of dust source areas. Understanding the implications of human-landscape interactions is therefore essential for monitoring and modelling land erodibility dynamics.

Anthropogenic interactions with the landscape affect spatial and temporal patterns in wind erosion by two main processes: 1) modification of vegetation cover and structure; and 2) modifications of soil surface condition. Gillette (1999) and Okin et al. (2006) note that the threshold dependent nature of wind erosion has a magnifying effect on the importance of the

spatial distribution of its controls within a landscape. If a landscape is modified as a result of human activity, or natural disturbance processes such as fire, it may move over some threshold, resulting in a non-linear increase in its susceptibility to wind erosion. The way in which landscapes are modified by human activities is largely dependent on land use. Land uses commonly associated with wind erosion are cultivation and pastoralism. The management of landscapes under these land uses is quite different, and so the implications of these on wind erosion processes also differ.

In cultivated lands there are often distinct boundaries between areas with particular land surface conditions. The effects of land management on land erodibility in cultivated environments are primarily seen through: 1) control of vegetation cover (crops); 2) control of field length (fetch); 3) control of soil aggregation and surface roughness through cultivation; and 4) control of soil moisture conditions through irrigation. While climate variability plays a significant role in driving the state of these conditions, intensive management will often affect the spatial distribution of erodible land areas (e.g. the location of fallowed fields) and the timing of changes in controls (Leys, 1999).

In rangelands, spatial and temporal changes in land erodibility are affected by climatic conditions and grazing pressures. Again, these affect spatial and temporal patterns in vegetation cover, moisture and soil erodibility. Boundaries of management in rangelands may appear to be less structured than in cultivated environments. However the effects of vegetation consumption by grazing animals and trampling of the soil surface can vary from creating localised patches of highly erodible land to regional scale increases in erodibility. Spatial variability in grazing pressures may lead to the formation of wind erosion ‘hot spots’ that are often significant dust emitters (Gillette, 1999).

2.4 Conceptual Model of Land Erodibility

This section provides a synthesis of the systems analysis presented in Section 2.2. It summarises the effects of controls on soil and land erodibility and combines these in a conceptual model of the land erodibility continuum.

Soil erodibility is the susceptibility of a soil to mobilisation by wind. It is controlled by the availability of loose erodible material (< 0.84 mm) on the soil surface as determined by aggregation (aggregate size distribution and aggregate stability), surface crusting and soil moisture content. Factors controlling soil erodibility are soil texture (particle size distribution), moisture content, chemistry, organic/biological content, climate variability and land management. The physical characteristics of the soil erodibility continuum are described in detail in Chapter 4.

Land erodibility is the susceptibility of a land area to erosion by wind. Land erodibility is a function of soil erodibility with the added effects of non-erodible surface roughness elements (rocks, vegetation, landforms). The effect of non-erodible elements like vegetation is through a partitioning of wind shear between roughness elements and the soil substrate resulting in an increase in roughness length and potential decrease in wind erosivity at the soil surface. Physically, the effect of non-erodible elements on wind erosivity can be described by an increase in the threshold friction velocity. Factors controlling land erodibility include those affecting soil erodibility, land type characteristics (vegetation and geomorphology), climate and management. Where non-erodible roughness elements are absent, land erodibility is controlled by soil erodibility.

The factors controlling soil and land erodibility operate at a range of spatial and temporal scales. This means that soil and land erodibility are spatio-temporally dynamic. Figure 2.10 illustrates the spatial and temporal scales over which controls on land susceptibility to wind erosion operate. While early classifications ranked soil erodibility using temporally static scaling systems, modern studies must consider the physical manifestation of erodibility as it exists within a continuum (Leys et al., 1996; Geeves et al., 2000).

Overarching controls on the susceptibility of land to wind erosion are climate, followed by soil and vegetation properties and land management. A prerequisite for wind erosion is an absence of moisture. With this dryness comes low vegetation cover and a reduced capacity for inter-particle cohesion. Factors controlling this lack of moisture include precipitation quantities and frequency, solar radiation, potential evaporation and wind speeds. Land use and land management affect wind erosion largely by controlling vegetation type and cover, and the soil surface properties.

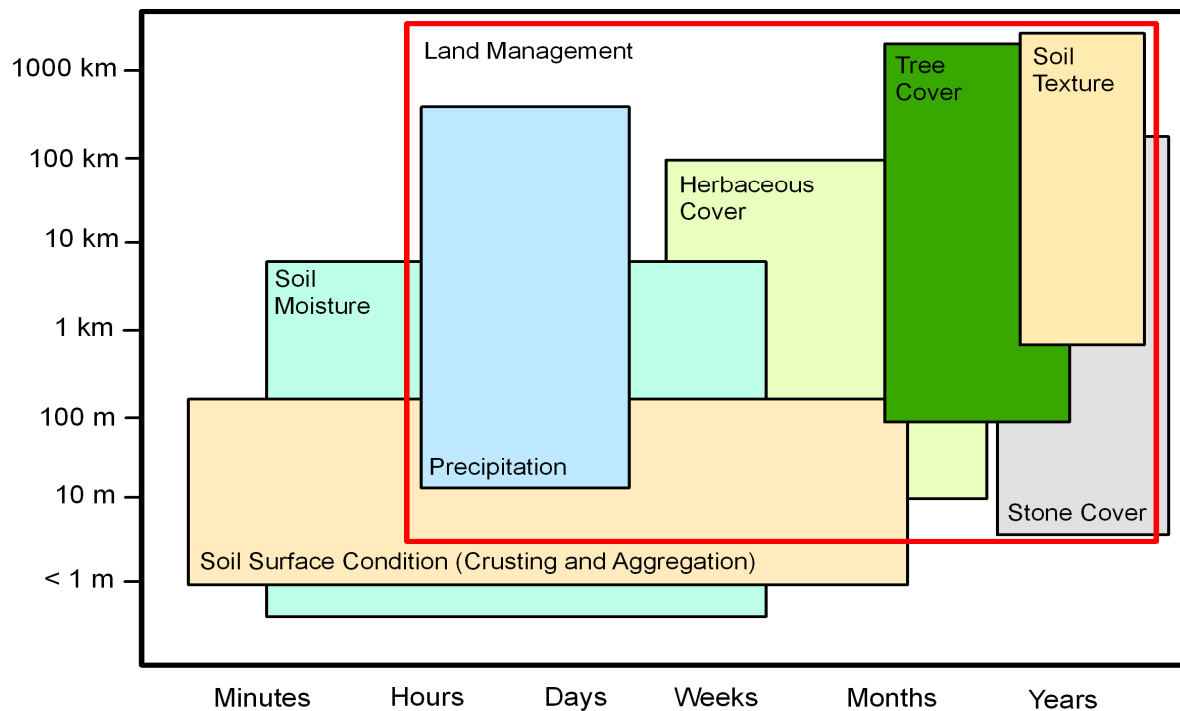


Figure 2.10 Space-time plot illustrating the spatial and temporal scales over which wind erosion controls operate.

Conditions conducive to wind erosion are those that minimize the forces holding particles to the soil surface, and maximize the wind shear velocity. Gillette (1999:75) summarised these as being:

- High wind speeds;
- A dry environment that lacks soil moisture, soil crusting and aggregation;
- Unvegetated surface free of rocks, pebbles, cobbles and boulders leading to a lack of drag partitioning and sediment trapping;
- Sandy sediments, or particle size, crusting and aggregation leading to the low threshold friction velocities ($80 < \text{grain diameter} < 120 \mu\text{m}$);
- Disturbance mechanisms to break down aggregates and surface crusts;
- High particle availability and a thick deposit so that supply limitation does not occur;
- Low binding energies of particles suspended in the soil matrix (poor aggregated);
- Long fetch – leading to the Owen Effect – sediment lines up parallel to strong winds; and
- Smooth surface ($z_0 < 0.1 \text{ cm}$) – leading to low threshold friction velocity and a lack of particle trapping.

The effects of climate, soil properties, cohesion agents, management and vegetation on wind erosion can be physically defined through the computation of u_{*t} for a land area. This threshold computation can be seen as a function employing adjustments to the principal formulation presented by Bagnold (1941) in Equation 2.5. Shao (2000) presented a pragmatic approach to defining u_{*t} based on a standard threshold for sand particles of size d_s where the soil is dry, bare and free of crust and salt:

$$u_{*t}(d_s; \lambda, \theta, sc, cr, \dots) = u_{*t}(d_s) f_{\lambda}(\lambda) f_{\theta}(\theta) f_{sc}(sc) f_{cr}(cr) \dots \quad (2.34)$$

where $u_{*t}(d_s)$ is the threshold friction velocity for sand particles of size d_s where the soil is dry, bare and free of crust and salt. The formula is a function of particle size only, which could be determined by wind tunnel experimentation with loose sand. Adjustments are made to this base threshold by incorporating functions for λ , the frontal area index for surface roughness elements, θ , soil moisture content, sc the soil salt content, and cr , a surface crust factor. Additional adjustments can be made to the formulation as required.

Conceptually, erodibility can be considered the inverse of u_{*t} . A decrease in erodibility implies an increase in u_{*t} , as more wind energy is required to mobilise particles on the soil surface. While u_{*t} has some dimension and can be physically measured, erodibility is a concept that can only be inferred through proxies like u_{*t} , or through empirical relationships between environmental conditions and erosion rates for a given wind speed.

The systems analysis presented in this chapter has demonstrated that a number of models are available that can be used to represent the effects of these conditions on soil and land erodibility. The models vary considerably in form and complexity. Approaches to integrating these models to assess land erodibility and simulate wind erosion processes are also diverse, but in general are based on the framework of Equation 2.34.

Figure 2.11 presents a conceptual model of the land erodibility continuum. The conceptual model simplifies theories of landscape dynamics (e.g. Noy-Meir, 1973; Westoby et al., 1989), presenting the range of potential vegetation and soil erodibility conditions within individual and separate continuums. In doing this, the model suggests that the state of the erodibility of a land area can be defined by the response of its fundamental controls (vegetation cover and

soil erodibility) to external forcing mechanisms. In this way the condition represents a balance of competing forces that drive changes in the system.

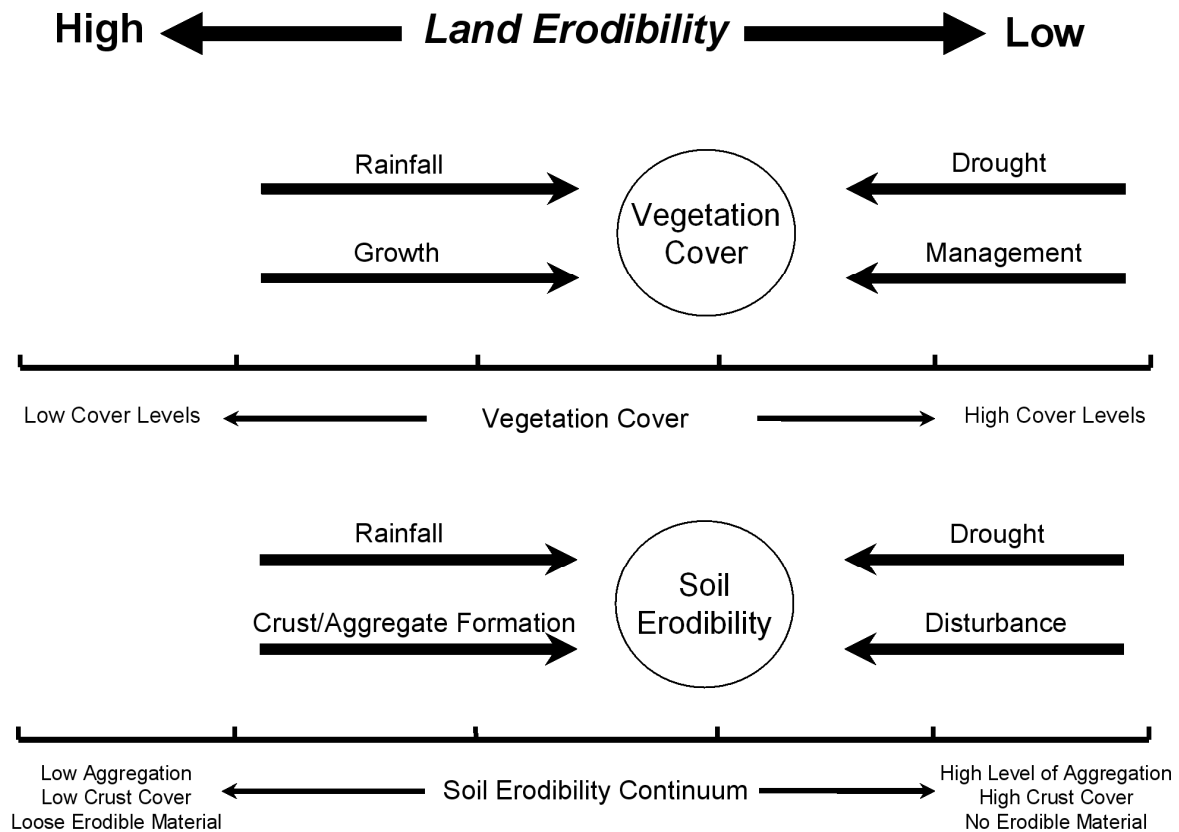


Figure 2.11 Conceptual model of land erodibility, expressed as a function of surface roughness effects due to vegetation cover (top) and soil erodibility (bottom). Land erodibility is determined by the relative conditions of surface roughness and soil erodibility. In turn, these controls are regulated by climate and land management conditions.

Land erodibility is controlled by relative positions of the dynamic vegetation cover (surface roughness) and soil erodibility factors within their respective continuums. In reality the factors are linked and interact through multiple and complex feedback mechanisms. In addition to the feedbacks, Section 2.2 has demonstrated that changing soil conditions and vegetation cover have a non-linear effect on land susceptibility to wind erosion. The implications of the physical relationships underlying land susceptibility to wind erosion are that for land types under different climate and land use/land management regimes, factors controlling land erodibility will exhibit high levels of spatial and temporal variability. The

relative importance of controlling factors and antecedent conditions will also vary, resulting in multiple potential erodibility outcomes for any soil or land area (Leys et al., 1996).

2.5 Summary

This chapter has presented a review of the concepts of soil and land erodibility. This was followed by a systems analysis of the factors controlling wind erosion. The systems analysis reviewed the state of our understanding of the relationships between meteorological, soil, vegetation and management factors and their effects on soil and land erodibility. The effects of these conditions were synthesised in a conceptual model of the land erodibility continuum. The conceptual model provides a context and scientific basis of the review of wind erosion models presented in Chapter 3. The review and the conceptual model underpin the development of the new soil and land erodibility models in Chapters 4 and 5.

Chapter 3

Approaches to Modelling Land Erodibility to Wind

This chapter addresses Objective 2 by reviewing approaches for representing land erodibility in current wind erosion modelling systems. Models are assessed that are applicable at a range of spatial scales, from the paddock (10^3 m^2) to regional (10^4 km^2) and global scales. The models integrate process relationships discussed in Chapter 2. The chapter summarises current limitations to modelling soil and land erodibility, and examines future research priorities for applying models to assess land susceptibility to wind erosion. This chapter provides a rationale for developing and applying new soil and land erodibility models in Chapters 4 to 7.

3.1 Introduction

Wind tunnel and field scale experimentation have been used to assess the effects of environmental controls on wind erosion (Chapter 2). The development of models that describe wind erosion processes has provided a means for learning more about wind erosion at multiple spatial and temporal scales. While in general these models have been developed to predict the product of wind erosion, dust emission, all require some component that describes the susceptibility of the land surface to wind erosion. The means by which this land erodibility is determined varies depending on the level of understanding of key processes, the desired complexity of the modelling systems, and the availability of input data at the required resolutions and coverage.

This chapter presents a review of a selection of wind erosion models developed over the last fifty years. Sections 3.2 to 3.4 detail how land erodibility is parameterised and modelled at different spatial and temporal scales. Section 3.5 of the chapter then synthesises the modelling approaches, limitations and aspects requiring further research. The selection of models reviewed is not exhaustive, but represents a variety of approaches that address the prediction of land erodibility. Criteria used for the selection of the models reviewed include:

- That the models selected cover those across the full range of spatial scales in wind erosion modelling applications (from the field to global scales);
- That the models contain a representative range of land erodibility and dust source parameterisations, employing emission schemes developed from a range of approaches;
- That the emission schemes contain a representative sample of empirical and theoretical formulations; and
- That the models selected have been developed for a variety of physical environments including cultivated lands, rangelands, and broader desert landscapes.

Figure 3.1 illustrates the spatial and temporal scales at which the wind erosion models reviewed in this chapter operate.

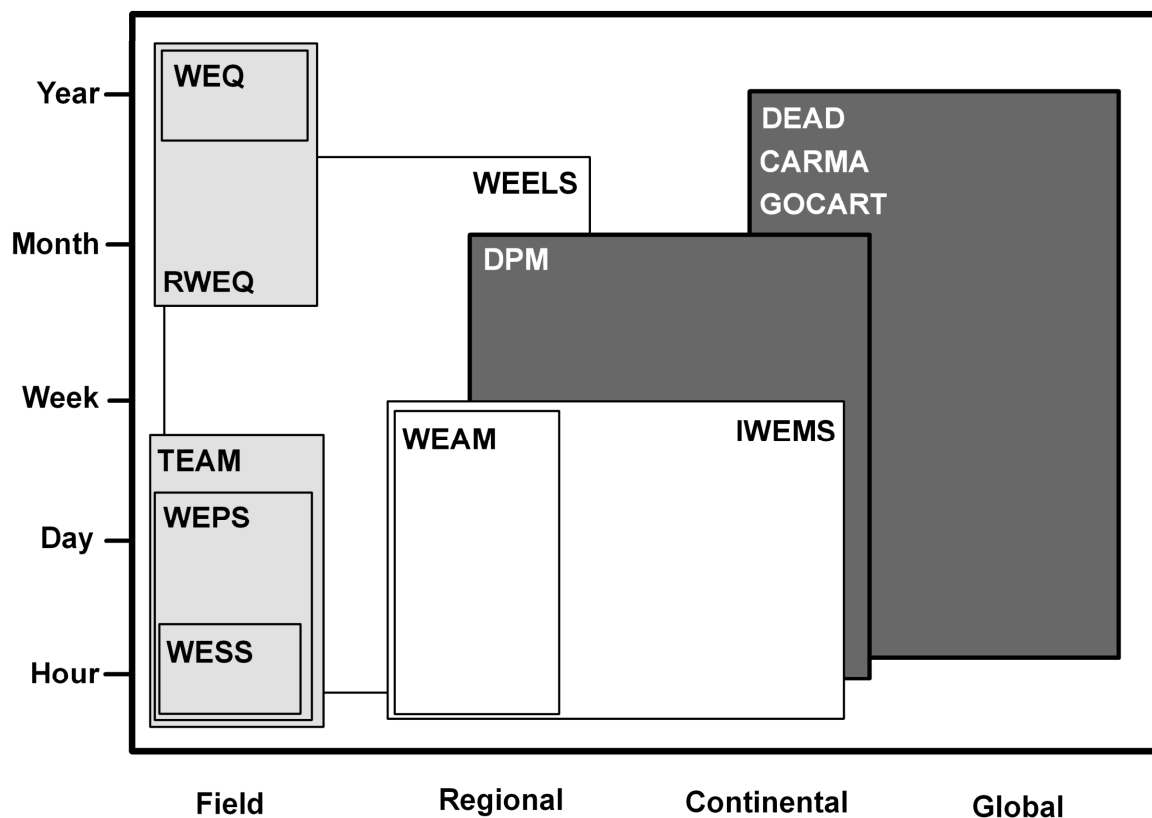


Figure 3.1 Space-time plot showing the spatial and temporal scales of wind erosion models reviewed in this chapter. Light gray boxes represent field scale models (Section 3.2), white boxes represent regional scale models (Section 3.3), and dark gray boxes represent global scale models (Section 3.4)

The first models reviewed are field scale models. These include empirical models designed to simulate single erosion events at scales of 10^2 to 10^3 m². The second group are the local to

regional scale models. For the purposes of this discussion, regional scale modelling is considered to be that where models are applied to areas $\sim 10^4 \text{ km}^2$, and operating over numerous landscapes within a region or continent. The final group are the continental to global scale models. These models operate at scales $>10^4 \text{ km}^2$, and are applied to simulate global scale dust dynamics.

3.2 Field Scale Wind Erosion Models

This section presents a review of field scale wind erosion models. The models reviewed include the Wind Erosion Equation (WEQ), Revised Wind Erosion Equation (RWEQ), Wind Erosion Stochastic Simulator (WESS), Wind Erosion Prediction System (WEPS), and Texas Erosion Analysis Model (TEAM). These models represent a range of approaches for assessing wind erosion from farm fields at different spatial and temporal resolutions. The way in which land erodibility is quantified in each model is seen as a progression from early empirical research with limited computing capability, to more recent process-based studies.

3.2.1 Wind Erosion Equation (WEQ)

The Wind Erosion Equation (WEQ) was developed by Woodruff and Siddoway (1965) to predict annual soil erosion (kg ha^{-1}) from farm fields in the United States. WEQ was designed for both the analysis and management of wind erosion, with a central aim in its development being to apply the model to determine the effects of field conditions and erosion mitigation strategies on erosion rates.

WEQ was developed from empirical relationships defining the effects of important wind erosion controls on soil loss rates. The foundation of WEQ is its soil erodibility factor. The empirical functions constituting the model were derived from field and wind tunnel experiments under a range of soil types and roughness conditions (e.g. Chepil and Woodruff, 1954). WEQ uses a relationship between five generalised factors that account for the major controls on wind erosion in cultivated environments. These factors are combined into five inputs for the model in the form:

$$E = f(I', C', K', L', V) \quad (3.1)$$

where I' is the soil and knoll erodibility; C' is the local wind erosion climatic factor; K' is the soil ridge roughness factor; L' is the field length; and V is the equivalent quantity of vegetation cover. Table 3.1 provides a description of each of the WEQ model input factors.

Table 3.1 Components of the Wind Erosion Equation (after Woodruff and Siddoway, 1965).

Control	Description
Soil Erodibility, I Knoll Erodibility, I_s	The potential soil loss (t/acre/annum) from a wide, unsheltered, isolated field with a bare, smooth, un-crustured surface. Developed from wind tunnel and field measures. I_s is used to compute erodibility of windward slopes less than 500 ft long – varies with slope.
Surface Crust Stability F_s	Considered insignificant as crust breakdown occurs due to aeolian abrasion once wind erosion has started. This factor is also transitional and is only considered significant where the erodibility of a field is to be computed for a given moment in time. I_s is usually disregarded.
Soil Ridge Roughness K_r	Measure of soil surface roughness (other than clods or vegetation).
Velocity of Erosive Wind v	Mean annual wind velocity corrected to a standard height (30 ft).
Soil Surface Moisture M	Moisture is assumed to be proportional to the Thornthwaite P-E Index (Thornthwaite, 1931).
Distance Across Field D_r	Total distance across a given field measured along the prevailing wind direction.
Sheltered Distance D_b	Distance along the prevailing wind erosion direction that is sheltered by a barrier, if any, adjoining the field.
Quantity of Vegetative Cover R'	Surface residue amounts determined by sampling.
Kind of Vegetative Cover S	Factor denoting the cross-sectional area of the cover.
Orientation of Vegetative Cover Variable K_o	A roughness variable. Values vary to describe prostrate to standing cover.

Following publication of WEQ, a number of modifications to the model were made by Woodruff and Armbrust (1968), Skidmore and Woodruff (1968), Skidmore et al., (1970), Bondy et al., (1980), Lyles (1988), and Skidmore and Nelson (1992). Bondy et al. (1980) modified WEQ such that the model could provide estimates of erosion rates for periods less than one year. While the foundation of WEQ is its soil erodibility factor, defined from field measured conditions, application of the model outside North America has been limited by its dependence on the availability of field-measured input conditions, which are expensive to acquire, and the coarse (annual) temporal resolution of its output. The model was designed for operation in cultivated lands and does not predict short-term or seasonal variations in wind erosion. This means that it does not accurately simulate wind erosion in rangelands, so its application in countries like Australia is further restricted.

3.2.2 Revised Wind Erosion Equation (RWEQ)

The Revised Wind Erosion Equation (RWEQ) was designed to predict soil loss due to wind erosion at sub-annual time scales. RWEQ combines empirical and process-based model components that became available following the development of WEQ. While WEQ provides estimates of annual soil loss, RWEQ can be applied to calculate the sediment transport mass at specific field lengths, as well as average and maximum soil loss within a field. A major development in RWEQ was the incorporation of management factors to describe soil surface conditions, vegetation state, and soil moisture changes due to irrigation (Fryrear et al., 1998; Fryrear et al., 2000).

The general input structure of RWEQ is similar to WEQ; however, the integration of model components reflects developments in field and wind tunnel studies that allowed for input conditions to be modelled (Fryrear et al., 1998). The inclusion of a scheme to compute soil surface conditions reflects these developments. Input variables of RWEQ include ER , the soil erodible fraction computed from soil properties; SCF , a soil crust factor computed from soil clay and organic matter content; WF , a weather factor; field size; COG , the crop type and orientation; and Hills, a factor used to modify wind speeds depending on field slope and height. The model computes soil loss by the expression:

$$Q_{\max p} = 109.8(WF.EF.SCF.K'.COG) \quad (3.2)$$

where $Q_{\max p}$ is the maximum amount of soil that can be transported downwind in an event. While RWEQ can be run at variable temporal resolutions, the model computes soil loss on an event basis. The input variables are computed by empirical relationships derived from field studies on agricultural soils, but also still rely on the measurement of field conditions prior to model application. The model requires inputs of soil conditions, as well as land type and land use, tillage/crop information, crop type, type of tillage tool, and amount and date of irrigation (Fryrear et al., 2001). The soil erodible fraction EF (%DA <0.84 mm) input is computed by:

$$EF = \frac{29.09 + 0.31SA + 0.17Si + 0.33\frac{SA}{CL} - 2.59OM - 0.95CaCO_3}{100} \quad (3.3)$$

where SA is the sand content (%), Si is the silt content (%), CL is the clay content (%), OM is the organic matter content (%), and $CaCO_3$ is the calcium carbonate content (%). The soil crust factor (CF) is computed using the equation:

$$CF = \frac{1}{1 + 0.0066(CL)^2 + 0.021(OM)^2} \quad (3.4)$$

The model weather factor (WF) combines measures of wind erosivity and a wind erodibility factor based on soil water content. The weather data is simulated and draws on an historical weather database (Fryrear et al., 1998). As for WEQ, the model K' factor describes the soil ridge roughness and COG (Equation 3.2) defines the percentage cover of dead, flat and standing plant material. Both the K' and COG factors can be either measured *in situ*, or derived from a series of empirical relationships (Merrill et al., 1999). Importantly, the factors in RWEQ reflect current field conditions for particular events, rather than annual average conditions (as for WEQ).

RWEQ can be applied to predict soil loss from multiple individual fields. However, the model inputs represent single value average conditions for the fields, and so the model assumes spatial homogeneity in the soil management, surface crusting, and vegetation cover conditions. The implication of this is that the model cannot be applied to accurately simulate soil loss in rangeland environments where field conditions are highly non-uniform (heterogeneous). The expressions used to compute the soil surface conditions are based on regression equations and data collected on soils in the United States (US). While RWEQ has been applied successfully outside the US (Van Pelt et al., 2004), this application is still reliant on experimental data to support model predictions of soil erodibility on soils different to those in the US (Leys, 1999).

3.2.3 Wind Erosion Prediction System (WEPS)

The Wind Erosion Prediction System (WEPS) was developed to advance wind erosion prediction through process-based modelling. While WEQ and RWEQ rely on empirical relationships and field-measured inputs, WEPS was developed as a process based, continuous, daily time-step wind erosion model that simulates weather, field conditions,

management, and erosion (Hagen, 1991). WEPS can be run to simulate single erosion events or soil loss over multiple years.

WEPS uses a database of inputs to define land surface characteristics. The inputs are processed by sub-models to compute soil and land erodibility components that are used by the erosion scheme. Inputs include data on climate, soils, management, and crop conditions. Table 3.2 provides a description of the WEPS sub-models.

Table 3.2 Definitions of the WEPS sub-models used to simulate soil loss due to wind erosion (after Hagen, 1991).

Sub-Model	Description
Weather	Contains programs that simulate wind speeds, air temperature, precipitation, solar radiation and dew point temperature. These are used as input to the remaining sub-models.
Hydrology	Simulates soil water content in the various layers of the soil profile and at the soil-atmosphere interface.
Management	Accounts for soil, surface and biomass manipulations that could affect the erosion process. These include timing and method of cultivation and irrigation.
Soil	Simulates random and oriented roughness, crust generation, cover fraction, density, stability and thickness, and loose erodible material on crusted surfaces. These temporal properties are simulated in response to weather processes like wetting/drying, freeze/thaw, precipitation amount and intensity, and time.
Crop	Simulates the amount of live biomass growth on the surface by calculating daily production of roots, stems and leaves. The growth model accounts for variations in biomass cover in response to climate and management decisions.
Decomposition	Simulates the decrease in crop residue biomass due to microbial activity. The biomass is partitioned into dead standing, surface, buried, and root. May be modified by management conditions.
Erosion	Uses input parameters supplied by the other sub-models that describe the soil surface, flat biomass cover, standing biomass leaf and stem areas, and weather to calculate threshold friction velocities and particle entrainment.

The sub-models operate at different temporal resolutions depending on the availability of input data and variability in input conditions. These temporal resolutions range from hourly to sub-hourly for the hydrology and erosion models, to daily changes in soil surface conditions that drive the erosion model (Hagen, 1991). Unlike the majority of wind erosion models, WEPS contains schemes to predict spatial and temporal variability in soil erodibility. This is in contrast to earlier models like RWEQ which rely on single value inputs to approximate soil erodibility within fields.

The WEPS erosion model operates with a scheme that computes entrainment when the wind friction velocity over a field surface (u_*) exceeds a threshold friction velocity (u_{*t}) for particle mobilisation (after Bagnold, 1941; Chapter 2). A static threshold is computed for field surfaces and employs factors to adjust for flat biomass cover, soil wetness, and aggregate size and density. The effects of standing biomass are used to adjust the wind friction velocity rather than the threshold velocity of the surface (Visser et al., 2005). The model allows for soil conditions to be updated during erosion events through changes to the aggregate distribution and crust factors.

First, a static threshold friction velocity WUB_{*ts} (ms^{-1}) is computed for a bare soil surface as a function of the fraction of the surface that can/cannot erode:

$$WUB_{*ts} = 1.7 - (1.35) \exp[-(b2)SF_{cv}] \quad (3.5)$$

where $b2$ is a coefficient computed as a function of the aerodynamic roughness height, and SF_{cv} is the fraction of bare soil that is not erodible. SF_{cv} is computed using an empirical relationship between the soil erodible fraction, rock cover, and clod and crust cover. Increases in the static threshold friction velocity WUC_{*tx} due to flat biomass are computed by:

$$WUC_{*ts} = 0.02 + SFC_{cv} \quad (3.6)$$

Further adjustments can then be made to adjust for soil moisture effects $WUCW_{*tx}$:

$$WUCW_{*ts} = 0.48 \frac{HR0_{wc}}{HR15_{wc}} \quad (3.7)$$

where $WUCW_{*ts}$ is the increase in u_{*t} due to surface wetness, $HR0_{wc}$ is the surface water content (kgkg^{-1}), and $HR15_{wc}$ is the surface water content at 1.5 MPa (kgkg^{-1}). The static threshold friction velocity is computed as an amalgam of Equations 3.5, 3.6 and 3.7:

$$WU_{*ts} = WUB_{*ts} + WUC_{*ts} + WUCW_{*ts} \quad (3.8)$$

WEPS uses a series of empirical relationships to compute changes in soil erodibility within the soil sub-model. The soil sub-model accounts for surface crusting through a linear relationship with cumulative precipitation. It then computes the amount of loose erodible material on the crust as a function of soil textural properties, organic matter, carbonate content (after Zobeck and Popham, 1992). As for surface crusting, the loose erodible fraction of soil can be adjusted within the model by accommodating precipitation effects (Hagen, 1991). Additional soil properties that affect soil erodibility such as dry aggregate size distribution and stability are also derived within the soil sub-model. Like RWEQ, WEPS has a relatively small spatial application, with the model simulation region being confined to a single field, or a few adjacent fields. The model accommodates land surface heterogeneity by dividing non-homogeneous regions into smaller homogeneous sub-regions, and running simulations for each (Hagen, 1991).

WEPS has been applied to simulate field conditions (crop residue roughness effects) and wind erosion on cultivated fields in North America and Europe (Van Donk and Skidmore, 2003; Funk et al., 2004; Hagen, 2004). Hagen (2004) compared WEPS predictions to measured data from 46 wind erosion events across North America. Across the sites, WEPS tended to under-predict soil loss, but reproduced the field data reasonably well ($r^2 = 0.71$). In a similar comparison, Funk et al. (2004) found a better match between WEPS predictions and field measured erosion rates ($r^2 = 0.91$) for a series of erosion events in Germany. Coen et al. (2004) used WEPS to map wind erosion risk of soils in Alberta, Canada. However, the model was run at a coarse (field scale) resolution, and did not provide information on high spatial and temporal resolution changes in soil erodibility.

3.2.4 Texas Erosion Analysis Model (TEAM)

The Texas Erosion Analysis Model (TEAM) was developed to have a low dependence on field-measured inputs and empirical relationships as used in WEQ, RWEQ and WEPS. TEAM simulates the detachment, maximum transport rate and emission of dust particles (Singh et al., 1999). Like WEPS, the model erosion scheme operates on the principle that entrainment occurs when u^* exceeds u_{*t} . The schemes used to compute land surface conditions draw upon external studies of the effects of land surface processes on u_{*t} . Like WEPS, TEAM computes both a static threshold friction velocity required for the initiation of entrainment, and a dynamic threshold for sustained sediment transport.

Input variables for TEAM include wind speed, relative humidity, soil particle size distribution, clay content, residue and vegetative cover, soil aggregate cover, field length, and windbreak height and porosity. The threshold friction velocity is calculated for the average particle size d (not including the clay fraction) and surface moisture conditions of the field undergoing simulation. This computation is based on wind tunnel research by Gregory and Darwish (1990), Darwish (1991), Gregory (1991) and Puri et al. (1925). The static threshold is computed by:

$$u_{*t} = 0.118 \left[21.2d \left(1 + 0.01W + \frac{0.0045}{d^2} + \frac{1.2}{d} e^{-0.1W/W_w} (W - W_c) \right) \right]^{0.5} \quad (3.9)$$

where d is the diameter of the average soil particle size not including clay, W is the water content of the soil in percentage weight, W_w is the wilting point, and W_c is the percentage of water on the surface needed to fill inter-particle spaces. The moisture content factor accounts for the change that occurs in u_{*t} once water content is above a threshold (described in Chapter 2, Section 2.2.5). The water content is based on atmospheric humidity (%). W_c is the threshold water content at which an increase in u_{*t} is initiated. If W is less than W_c set W equal to W_c so the last term does not apply (Singh et al., 1999).

The dynamic threshold friction velocity model accounts for the effects of soil surface condition, vegetation cover, and wind gustiness (Gregory et al., 2004). The dynamic threshold is computed by first adjusting u_{*t} to compensate for wind gustiness:

$$u_{*d} = 0.8u_{*t} / G_f \quad (3.10)$$

where G_f is a gust factor. The effects of vegetation cover (S) are applied to determine the maximum transport rate M , which is also a function of soil particle diameter at different stages of the particle distribution (D_{50} , D_{75} , D_g), B_d an empirical coefficient, and the dynamic threshold and u_* :

$$M = 0.004B_d \left(\frac{1}{u_{*d}} \right)^2 \frac{D_{50}}{D_R} \left(1 + 125 \frac{\sqrt{D_{75}} - \sqrt{0.08}}{\sqrt{D_R}} \right) (Su_*^2 - u_{*d}^2) u_* \quad (3.11)$$

where S is a function of the fraction of residue or aggregate cover, biomass cover, and the height of the plant canopy, residue, aggregates and other random soil roughness elements.

TEAM uses a factor to adjust for field length effects on wind erosion. The field length factor was developed by Gregory (1984) and built upon a relationship established by Chepil (1957) for field length effects on soil movement. The length factor accommodates abrasion processes in the calculation of the sediment transport rate from eroding fields, and is supplemented by an empirical abrasion factor that considers the field length, wind shear velocity, and detachment rates of aggregated/crusted and loose soils. Two factors are used to account for the erodibility of soils in the solid (crusted) and loose conditions (Wilson, 1994). The erodibility of a crusted soil surface is computed by:

$$E = \frac{\rho_{bs}}{\tau_s} f(\Theta) \quad (3.12)$$

where E is the solid state erodibility (kgJ^{-1}), ρ_{bs} is the soil bulk density (kgm^{-3}), τ_s is the soil shear strength (Nm^{-2}), and $f(\Theta)$ is a function of soil shear angle (dimensionless). For a loose soil state the erodibility is computed by:

$$E_l = N \frac{\rho_{bs}}{\rho_f u_{*t}^2} \quad (3.13)$$

where E_l is the erodibility of a loose soil (kgJ^{-1}), N is a calibration coefficient, ρ_f is the air fluid density (1.23 kgm^{-3}), and u_{*t} is the threshold friction velocity (ms^{-1}). The detachment ratio used to compute the field length factor is effectively a ratio of the solid and loose state soil erodibilities.

Testing TEAM wind erosion simulations against measurements of threshold friction velocity and sediment transport rates indicates that model performance is comparable with that of the WEPS and RWEQ models (Gregory and Darwish, 2001; Gregory et al., 2004). The model was found to perform well in comparison to measured erosion rates in bare and vegetated settings in both agricultural and industrial environments.

3.2.5 Wind Erosion Stochastic Simulator (WESS)

The Wind Erosion Stochastic Simulator (WESS) is the wind erosion module of the EPIC (Environmental Policy Integrated Climate) erosion model (Van Pelt et al., 2004). WESS is a process-based model that predicts wind erosion on an event or periodic basis. The model provides assessments of soil loss at user-specified distances within individual fields. The model uses inputs of soil surface data including texture, soil surface moisture and drying rate, erodible soil thickness, surface soil bulk density, and soil roughness parameters for large aggregates (random roughness), and ridge height and spacing (oriented roughness). This is coupled with local wind speed data and a stochastic wind speed perturbation factor to simulate dust emission (Potter et al., 1998). WESS simulates wind erosion based on daily wind speed distributions adjusted by soil, surface roughness, vegetation cover and eroding field length factors:

$$YW = (FI)(FR)(FV)(FD) \int_0^t \frac{YWR}{WL} dt \quad (3.14)$$

where YW is the wind erosion estimate (kgm^{-2}), YWR is the erosion rate ($\text{kgm}^{-1}\text{s}^{-1}$), WL is the unsheltered field length (m), FI is the soil erodibility factor, FR is the surface roughness factor, FV is the vegetation cover factor, FD is the field length factor (based on WL), and t is the duration of wind greater than a threshold velocity. YWR is calculated using the approach of Skidmore (1986). The soil erodibility adjustment factor is based on the static Wind Erodibility Groups (WEGs) reported by Woodruff and Siddoway (1965), and used in the WEQ model:

$$FI = \frac{I}{695} * \exp(-0.0001 * \sum YW) \quad (3.15)$$

where I is the soil erodibility factor (tha^{-1}) with values assigned by soil texture and the WEGs. The effect of oriented surface (cultivated ridge) roughness (FR) is calculated as a function of wind direction, and the impact angles of saltating grains:

$$FR = 1 - \exp \left[- \left(\frac{\alpha}{RFB} \right)^{RFC} \right] \quad (3.16)$$

where α is the saltating grain impact angle ($\sim 15^\circ$), RFB is based on the cultivated ridge height, wind direction and random roughness, and RFC varies with ridge height. The vegetation factor is computed as a function of modelled vegetation cover:

$$FV = \frac{VE}{VE + \exp(0.48 - 1.32VE)} \quad (3.17)$$

where VE is the vegetation cover factor (tha^{-1}). Both the soil and vegetation cover factors are scaled between 0 and 1. The field length (distance, FD) factor is modelled as a function of the field length in the prevailing wind direction.

Potter et al. (1998) described a comparison of WESS erosion simulations with measured erosion events in Alberta, Canada. Simulated erosion events corresponded to most of the measured wind erosion activity; however, the model was found to overestimate erosion for one event in seven, and occasionally predicted events when no erosion was measured. Poor model performance in these circumstances was attributed to the simplified nature of the modelling scheme, and the lack of a scheme to account for temporal changes in soil surface conditions such as crusting and the erodible fraction.

3.3 Local to Regional Scale Models

This section examines process-based regional scale wind erosion models. The models reviewed include the Wind Erosion on European Light Soils (WEELS) model, Wind Erosion Assessment Model (WEAM), and the Integrated Wind Erosion Modelling System (IWEMS).

3.3.1 Wind Erosion on European Light Soils (WEELS)

The Wind Erosion on European Light Soils (WEELS) model was developed to map wind erosion risk and to predict long-term wind-induced soil loss under different climate and land use scenarios (Böhner et al., 2003). The model was developed from the EROKLI model (Beinhauer and Kruse, 1994) and can be run at local scales in domains up to 5 x 5 km in size, using input data at a 25 x 25 m resolution. WEELS has a modular structure, and is run in a Geographic Information System (GIS) environment.

WEELS has a process-based system that simulates wind erosion when u^* exceeds u_{*t} . Four modules are combined to compute u_{*t} . The modules account for soil moisture, soil erodibility, soil roughness, and land use effects on the static threshold. Soil erodibility is calculated as a function of soil properties including soil grain size distribution and soil type. Land management conditions such as cultivation methods, crop type, crop rotations and land use scenarios are then used to adjust the soil erodibility, soil roughness and land use factors (Böhner et al., 2003).

WEELS computes soil erodibility then independently accounts for soil moisture and vegetation cover factors by integrating these into the wind erosivity component of the model. Soil erodibility is quantified using a dimensionless factor K which expresses the intrinsic susceptibility of a dry, freshly cultivated sandy soil to erosion. Values for K are computed as a function of the mean weighted diameter of the texture (by particle size distribution), organic matter content and the proportion of aggregates > 0.63 mm:

$$\log K = 1.24 - 4.21(MWD) - 0.04(\%DA) \quad (3.18)$$

where MWD is the soil particle mean weighted diameter (mm), and $\%DA$ is the weight percentage of aggregates > 0.63 mm (%) calculated as:

$$\%DA = -2.42 + 8.6[\log(\%OM)] + 44.5 \left[\frac{(\%Silt + \%Clay)}{\%Sand} \right] \quad (3.19)$$

where $\%DA$ is the aggregate portion > 0.63 mm (%), $\%OM$ is the weight percent soil organic matter, and $\%Silt$, $\%Clay$, $\%Sand$ are the soil silt, clay and sand fractions (%). Both Equations 3.18 and 3.19 are regression equations derived from field experimentation in Germany (Böhner et al., 2003). Field measurements of $\%DA$ can also be used as input to the model. Soil erodibility increases as K increases; however, K is considered static in WEELS and does not vary in response to changes in climate and management conditions. Vegetation effects on u_{*t} are considered by a logarithmic adjustment to u_{*t} . The WEELS soil moisture module relies on a simple soil water budget that operates for sandy soils, so model application is currently limited to sandy soils.

WEELS computes wind erosion as hourly assessments (thr^{-1}). The model can be used to map erosion risk in terms of the duration of erosive conditions and corresponding maximum sediment transport rate. Simulated monthly and annual time series of wind erosion conditions were compared with data from test sites in England and Germany (Böhner et al., 2003). Spatial and temporal patterns of modelled wind erosion were found to be in agreement with observations of wind erosion. However, no quantitative assessments of the model performance have been published.

3.3.2 Wind Erosion Assessment Model (WEAM)

The Wind Erosion Assessment Model (WEAM) was the first integrated model developed in Australia to quantify wind erosion rates (Shao et al., 1994). The model was designed to operate at a moderate spatial resolution and predict dust emissions from the Murray-Darling Basin in south-eastern Australia. WEAM combined current theories on the effects of climate, soil, vegetation and land use on wind erosion.

WEAM quantifies land erodibility through the threshold friction velocity, with erosion occurring when $u_* > u_{*t}$. The model computes streamwise sediment flux (Q) and vertical dust flux (F). It uses the saltation model of Owen (1964), a sand transport equation, to compute Q , which is adjusted using soil particle size distributions. The wind shear velocity (u_*) is determined from climatic conditions and the surface roughness height (Shao et al., 1996). The threshold friction velocity is calculated as a function of soil particle size, frontal area index of surface roughness elements, soil moisture, and the hardness of the surface crust:

$$u_{*t}(d_s, \lambda, w, c) = \frac{u_{*t}(d_s, 0, 0, 0)}{R(\lambda)H(w)M(s)} \quad (3.20)$$

where $u_{*t}(d_s, 0, 0, 0)$ was approximated from the model of Greeley and Iversen (1985), $R(\lambda)$ is the ratio the bare threshold velocity over the covered (rough) threshold velocity, $H(w)$ is the ratio of the threshold velocity of the wet surface over the threshold velocity of the dry surface and, $M(s)$ is a mobility coefficient describing the influence of the state of soil aggregation and crusting, and the chemical binding strength which maintains this state.

The effects of water content are defined through an exponential relationship derived from wind tunnel experiments by Shao et al. (1996):

$$H(w) = u_{*t}(0)/u_{*t}(w) = \exp(-22.7w) \quad (3.21)$$

where w has the units m^3m^{-3} . Roughness effects are computed by the drag partition model of Raupach et al. (1993). The effects of surface roughness and cover are defined by:

$$R(\lambda) = (1 - \sigma_r \lambda)^{-0.5} (1 + m_r \beta_r \lambda)^{-0.5} \quad (3.22)$$

where β_r is the ratio of the drag coefficient for isolated roughness elements over the drag coefficient for the surface, σ_r is the basal-to-frontal area ratio and m_r is a parameter ≤ 1 accounting for non-uniformity in the surface stress (Chapter 2, Section 2.2.8). The frontal area index is estimated from fractional cover f_c by the equation:

$$\lambda = -c_\lambda \ln(1 - f_c) \quad (3.23)$$

where c_λ is an empirical constant related to the distribution of roughness elements (a value of 1 is assigned implying elements are uniformly distributed and isotropically oriented). The effect of dry, loose soil on u_{*t} is computed by the scheme of Greeley and Iversen (1985):

$$u_{*t}(d) = A_l F(Re_t) G(d) (g \sigma d)^{0.5} \quad (3.24)$$

where g is the gravitational acceleration, σ is the particle-to-air density ratio, A_l is an empirical coefficient, $F(Re_t)$ and $G(d)$ are empirical functions, and Re_t is the threshold particle Reynolds number. The parameters $R(\lambda)$, $H(w)$ and Sc are estimated based on wind tunnel experiments carried out in the study area (Shao et al., 1996). The model is linked to a GIS which contains input data on soils, vegetation, land management and climate across the Murray-Darling Basin.

The M parameter (Equation 3.20) and particle size distributions are linked to the soil data in the GIS database, with the M values being assigned to soil texture classes based on descriptions of soil properties in the Australian Natural Resources Atlas. While WEAM does

not directly account for soil surface crusting and aggregation, the M parameter values are based on the likelihood of crust formation, hydraulic conductivity and the chemical composition of the soils. The model does not, however, account for temporal changes in soil surface conditions such as the effects of grazing on soil crusting (Shao and Leslie, 1997). The availability or supply of erodible material on the soil surface is not specified in the model, so the predictions are unrestrained by temporal changes in soil erodibility. The vegetation data used as input to the model are derived from satellite imagery of the Normalised Difference Vegetation Index (NDVI), and do not represent actual cover measurements (%). Climate data for the model are sourced from the Australian Government Bureau of Meteorology, and the model uses a stochastic simulator to generate synthetic time-series meteorological inputs.

Shao et al. (1996) described application of WEAM. A number of characteristics of WEAM affect its performance in predicting dust emission. The limitations of this model are related to: 1) the absence of a scheme to account for spatial and temporal changes in surface erodibility, and 2) the fact that the model for vegetation cover effects and vegetation input data does not differentiate or account for the separate effects of both prostrate and standing cover. The frontal area index (λ) describes the effects of standing cover, but does not describe the effects of prostrate cover. Given the mix of cover types across the study region, the estimate of frontal area may have been too simplistic.

Future development areas noted for WEAM were in the extension of the model to account for temporal changes in soil erodibility. In achieving this, Shao et al. (1996) noted that research is required to develop models that describe the soil aggregation state, formation and breakdown of surface crusts, transitions between transport and source-limited situations, and the effects of land management (including cultivation and grazing).

3.3.3 Integrated Wind Erosion Modelling System (IWEMS)

The Integrated Wind Erosion Modelling System (IWEMS) is an extension of WEAM (Lu and Shao, 2001). IWEMS was developed by coupling WEAM with climate and land surface simulators within a GIS framework. The model was developed with the capacity to receive input data from a weather prediction model. The emission scheme and dust transport model can be applied at a national level across Australia. The input GIS database for land surface conditions has a spatial resolution of ~25 x 25 km at surface. The atmospheric component has

10 to 31 vertical layers, and a horizontal resolution ranging from 5 x 5 km to 75 x 75 km. The atmospheric model provides forcing for both the dust emission and transport models (Lu and Shao, 2001).

In addition to the spatial expansion of model application, and input upgrades, the model estimation of u_{*t} was also revised. The revisions built upon the model presented by Shao et al. (1996) and Shao and Leslie (1997), with the inclusion of factors to account for multiple non-erodible roughness element layers, and the erodible fraction of the exposed surface (Lu and Shao, 2001). In a development from the drag partitioning scheme used in WEAM, IWEMS uses a double drag partitioning approach that allows for the independent (and combined) effects of large roughness elements (e.g. trees) and also small roughness elements to be modelled (Shao, 2000).

Lu and Shao (1999) developed a soil classification map to account for spatial differences in soil erodibility. In using this the model groups soil inputs into erodible and non-erodible classes, with the erodible soils being assigned particle size distributions based on field samples from the soil type classes (Lu and Shao, 1999). This is similar to the methodology employed by Marticorena and Bergametti (1995) to account for spatial variability in desert dust source areas (Section 3.4.1). Due to a lack of quantitative research on the temporal evolution of surface crusts in Australia, the surface crust/mobility factor remained set to a constant (1) for all soils. The model therefore still lacks a dynamic component to define soil erodibility (Lu and Shao, 2001). This means that temporal changes in u_{*t} in IWEMS are driven by variations in surface roughness and soil moisture conditions.

3.4 Continental to Global Scale Models

This section describes the prediction of land erodibility in continental and global scale dust emission and transport models. The section includes reviews of the Dust Production Model (DPM) and Dust Entrainment and Deposition Model (DEAD), and notes characteristics of the Community Aerosol and Radiation Model (CARMA), and the Global Ozone Chemistry Aerosol Radiation and Transport (GOCART) model. In general these models use erosion schemes developed for smaller scale models, with adaptations to suit their application environments.

3.4.1 Dust Production Model (DPM)

Marticorena and Bergametti (1995) developed the Dust Production Model (DPM). The model was designed with a dust emission scheme that accounts for spatial variations in dust source erodibility. Early regional to global scale models (e.g. Tegen and Fung, 1995) compute dust emission as a function of wind velocity. The DPM computes emissions in a similar way to WEAM and IWEMS, using a relationship between soil particle size distribution, surface roughness and u_{*t} . Importantly, the approach allows for the direct and relative contributions of various source areas to global dust emissions to be quantified (Marticorena and Bergametti, 1995).

The basis of the dust emission scheme is that land erodibility is strongly dependent on soil texture and surface roughness characteristics. Values for u_{*t} are assigned to soil input maps based on the size distribution of particles in different textured soils (Marticorena and Bergametti, 1995). Semi-empirical equations of Iversen and White (1982) were then modified to obtain a relationship between particle or aggregate diameter and u_{*t} . The relationship was developed as a piecewise function dependent on the Reynolds number (here denoted B):

$$u_{*t}(D_p) = \frac{0.129K}{\left(1.928(aD_p^x + b)^{0.092} - 1\right)^{0.5}} \quad \text{for } 0.03 < B < 10 \quad (3.25)$$

$$u_{*t}(D_p) = 0.129K \left[1 - 0.0858 \exp(-0.0617((aD_p^x + b) - 10))\right] \quad \text{for } B > 10$$

where D_p is the size of the soil surface aggregates (cm),. The Reynolds number (B) is described by the term:

$$B = aD_p^x + b \quad (3.26)$$

where $a = 1331$, $b = 0.38$, and $x = 1.56$. The influence of surface roughness on the loss of wind momentum is accounted for using a scheme developed by Marticorena and Bergametti (1995) and Marticorena et al. (1997). The drag partitioning scheme of Raupach et al. (1993) requires a measure of the frontal area and estimation of the m parameter, so an alternate specification of drag was included in DPM and is based on the roughness length z_0 :

$$f_{eff}(Z_0, z_{0s}) = 1 - \left(\ln \left(\frac{Z_0}{z_{0s}} \right) / \ln \left(0.35 \left(\frac{10}{z_{0s}} \right)^{0.8} \right) \right) \quad (3.27)$$

where z_0 is the aerodynamic roughness length of the overall surface (cm), and z_{0s} is the aerodynamic roughness length of the erodible part of the surface (cm). The following expression is then used for the computation of u_{*t} :

$$u_{*t}(D_p, Z_0, z_{0s}) = \frac{u_{*t}(D_p)}{f_{eff}(Z_0, z_{0s})} \quad (3.28)$$

where u_{*t} is a function of particle or aggregate size, total surface roughness length (including vegetation), and soil surface roughness length. Soil moisture effects on u_{*t} were not accounted for in early versions of the model, but were included by Laurent et al. (2006) based on the work of Fécan et al. (1999). Further, the model does not consider the effects of surface crusting and does not consider temporal variations in soil aggregation. All soils are therefore considered loose and erodible if $u_* > u_{*t}$ with field measurements of soil aggregation being used to define soil particle size distributions for non-sandy soils. As values of u_{*t} are assigned to particle size groups, the contributions of size groups to modelled dust emissions are proportional to the surface area covered by the groups.

Qualitative validation of the DPM was presented by verification of the model functions during model development (Marticorena and Bergametti, 1995). The drag partitioning scheme was validated by comparison with stress partition measurements for roughness lengths presented by Marshall (1971). Further tests of the model were conducted by application in Africa, the Middle-East and China (Marticorena et al., 1997; Alfaro and Gomes, 2001; Lasserre et al., 2005). For these applications region specific soil textural databases were used to determine dust source erodibilities, and roughness elements were considered static and assigned from vegetation cover maps. Simulated dust emissions from the DPM have been compared with satellite imagery of dust events using the Infra-red Difference Dust Index (Brooks and Legrand, 2000), field records of dust entrainment, and dust observations averaged across regions (Marticorena et al., 1997).

3.4.2 Dust Entrainment and Deposition Model (DEAD)

The Dust Entrainment and Deposition Model (DEAD) is a global dust emission and transport model (Zender et al., 2003a). The dust entrainment and mobilisation scheme used in DEAD is similar to that of Marticorena and Bergametti (1995). The model emission scheme is particle size dependent, with u_{*t} being computed as a function of soil particle size distributions. Modifications are made to u_{*t} by accounting for soil moisture effects using the scheme of Fécan et al. (1999), and surface roughness effects using the drag partitioning scheme of Raupach et al. (1993). The model considers the Owen effect, a positive feedback of saltation on the surface roughness length and friction velocity, using the model of Gillette et al. (1998).

Dust source area soil erodibilities in DEAD are defined by soil particle size distributions from a global soil texture dataset (Zender et al., 2003a). Land surface and geographic constraints are applied to the dust source areas in addition to the roughness and moisture parameterisations in the emission scheme. The fraction of bare soil surface is described by a parameter which is a function of total vegetation, snow, lake and wetlands cover. Vegetation cover is derived from satellite imagery of Leaf Area Index (LAI) at a 1 x 1 km spatial resolution. These factors are considered constant during simulations. A further adjustment is made for dust source area erodibility S . This adjustment has a basis in global dust source area characterisations reported by Ginoux et al. (2001), identified using Total Ozone Mapping Spectrometer (TOMS) aerosol optical thickness image data. Zender et al. (2003a) defined S as the upstream area from which sediment transported in surface runoff may have accumulated. The S parameter was included to account for the relationship between global dust source area location and sediment recharge in drainage depressions. Zender et al. (2003b) reported on the effects of variants on the S parameter for uniform, topographic, geomorphic and hydrological source erodibility factors on global dust load predictions. Results demonstrated that simulations of global dust emissions are highly sensitive to dust source area characterisations.

Grini et al. (2005) examined dust sources and transport with DEAD using four soil erodibility factors. The soil erodibility factors were defined to capture characteristics of global dust source areas and included the topographic and geomorphic factors used by Ginoux (2001) and Zender et al. (2003b), for example:

$$TOPO = \left(\frac{z_{\max} - z}{z_{\max} - z_{\min}} \right)^5 \quad (3.29)$$

where *TOPO* is the erodibility factor, *z* is the elevation of the relevant grid point, and $z_{\max}$ and $z_{\min}$ are the highest and lowest points in the surrounding area (10° x 10° lat./long.). For the geomorphic factor, erodibility was considered proportional to the upstream area from which dust source area sediments accumulate. The other two erodibility factors were based on linear and non-linear ratios of the surface reflectance at a grid cell to the global maximum surface reflectance (recorded in the Sahara). Surface reflectance data were acquired from the Moderate Resolution Imaging Spectroradiometer (MODIS) satellite sensors. The study demonstrated how continental dust emissions are strongly influenced by source area erodibility characterisations. A comparison of the model output with measured dust loadings indicated that the erodibility factors performed well in representing spatial patterns in dust source strengths (Grini et al., 2005).

For validation, DEAD was implemented for an analysis period from 1990 to 1999 as a component of the Model for Atmospheric Chemistry and Transport (MATCH) Chemical Transport Model (CTM). Meteorological inputs for the simulations were sourced from observational rather than simulated data (Zender et al., 2003a). Simulations of dust loading and dry and wet deposition were compared with predictions from Duce et al. (1991) and Prospero (1996), and the Global Ozone Chemistry Aerosol Radiation and Transport (GOCART) model (Prospero et al., 2002). DEAD dust emission predictions were found to be consistent with the International Panel on Climate Change (IPCC) estimate ranges for emissions (Penner et al., 2001). DEAD predictions of Aerosol Optical Depth (AOD) were also compared with measurements from the Advanced Very High Resolution Radiometer (AVHRR) and TOMS observations. Finally, DEAD dust concentrations were compared with measured concentrations from 18 stations with long-term surface dust concentration records. Good agreement was found between predicted and observed records (Zender et al., 2003a). Limitations of the model are consistent with those of other dust emission models. The most significant of these is that the model does not account for temporal variations in soil erodibility.

3.4.3 Other Global Dust Models

A number of global dust emission models have been developed, and there are similarities in their erosion and dust emission schemes. In general, three approaches have been taken to model wind erosion (Zender et al., 2003a). The dependence of the model emission schemes on factors controlling land erodibility varies by the nature of their formulations.

The first type parameterise mobilisation in terms of the third or fourth power of the wind speed or friction velocity, then impose size distribution factors on the emitted dust (e.g. Tegen and Fung, 1995; Mahowald et al., 1999; Perlwitz et al., 2001). These models are reliant on assumptions about the general characteristics of dust source areas and do not account for micro-physical entrainment processes (Zender et al., 2003a).

The second type use a microphysical specification of the land surface to predict size-resolved saltation mass flux and dust emission (e.g. Marticorena and Bergametti, 1995; Gillette and Passi, 1988; Shao, 2001). Due to the complexity of inputs required for these models, this model type has typically been applied in regional scale modelling where spatial data requirements are often better met and inherent assumptions built into the model are less likely to violate the ranges of conditions seen in global dust source areas.

The third type represents those models employing micro-physical parameterisations with a number of simplifying assumptions to account for global dust source characteristics. In general these models are not able to accommodate fine scale soil erodibility factors like the field to regional scale models. These models include the Community Aerosol and Radiation Model for Atmospheres (CARMA), the Global Ozone Chemistry Aerosol Radiation and Transport (GOCART) and DEAD (Ginoux et al., 2001; Woodward, 2001; Luo et al., 2003; Zender et al., 2003a, 2003b; Barnum et al., 2004).

Recent advances in the application of remote sensing to detect atmospheric aerosols and global source areas have enabled significant improvements to be made to global dust emission models. An example of this is in the DEAD model, where global characterisations of dust source areas by a range of geomorphic, topographic and hydrological factors have enabled more advanced dust source parameterisations to be employed. These source area

factors represent a shift forward from broad global dust source characterisations, and result in improved estimations of global dust loads.

3.5 Synthesis and Discussion

In general, two approaches have been taken to representing soil and land erodibility conditions in wind erosion models. These include: 1) integrating empirical relationships between soil surface conditions, moisture content and vegetation cover to compute rates of soil loss (e.g. WEQ, RWEQ); and 2) using mechanistic approaches that seek to integrate physical and theoretical relationships between soil and land surface conditions and u_{*t} (e.g. WEPS; DPM; IWEMS).

The development of wind erosion models has been characterised by a shift from empirically-based field scale analyses to process-based regional to global scale analyses. This progression has induced changes in the model input data requirements, which reflect both increases in model complexity and in the availability of spatial data. The temporal resolution at which wind erosion models operate has also increased, from annual to sub-hourly time-steps. These developments have led to changes in the complexity in the way in which soil and land surface conditions are represented in the models. This complexity has generated a need for greater computing power and has brought increased attention to the use of integrated modelling approaches, e.g. coupling wind erosion models with climate models (Shao, 2000).

Surprisingly, few of the models reviewed in this chapter have been applied to assess wind erosion hazard. While the WEELS and WEPS models have been applied to assess wind erosion risk in Europe and the United States (Böhner et al., 2003; Coen et al., 2004), published applications of the models for this purpose have been limited to geographically small areas ($<100 \text{ km}^2$) and have not included analyses of temporal changes in soil or land erodibility. Application of the models reviewed in this chapter has generally been with the view to assess spatial and temporal patterns in soil loss and dust transport (e.g. Song, 2004; Visser et al., 2005; Sun et al., 2006). Despite their capabilities, none of the models have been applied explicitly to assess long-term spatial and temporal patterns in soil and land erodibility dynamics. This is consistent through the application of the models from the field to global

scales, and in particular there is a significant lack of modelling research at the landscape scale (i.e. $\sim 10^3 \text{ km}^2$).

A number of common challenges and limitations have emerged in representing land erodibility in wind erosion models. The first of these represents arguably the greatest challenge in wind erosion modelling, and affects the accuracy of model representations of land susceptibility to wind erosion. They include (after Raupach and Lu, 2004):

- Reliability of control representations and ability to account for soil erodibility dynamics;
- The availability of suitable input data;
- Up-scaling models and accounting for sub-grid scale heterogeneity; and
- Validation of regional to global scale models.

3.5.1 Reliability of Control Representations

The reliability of how controls are represented in wind erosion models has been affected by: 1) a lack of research into the temporal dynamics of soil erodibility, and 2) our ability to account for sub-grid scale variations in soil and land surface conditions, in particular the heterogeneous distribution of surface roughness. The effects of sub-grid scale heterogeneity on model performance are described in Section 3.5.3.

Raupach and Lu (2004) note that deficiencies in dust source parameterisations account for a large part of the observed discrepancies in model estimations of dust emissions. The accurate representation of spatial and temporal patterns in land erodibility is therefore essential for good model performance. While field scale models, for example WEQ, RWEQ, WEPS, contain factors to account for and simulate changes in soil erodibility, such factors have not been integrated into the regional to global scale models (Zobeck et al., 2003). These models account for spatial variations in soil erodibility by estimating u_{*t} as a function of the soil particle size distribution (e.g. Marticorena and Bergametti, 1995) and/or by designating regional dust source areas using topographic, geomorphic or remote sensing based indicators (e.g. Ginoux et al., 2001; Grini et al., 2005). The rationale for omitting soil erodibility factors from these broad-scale models is that robust models to simulate temporal changes in soil aggregation and crusting simply do not exist (Shao, 2000). Empirical models, like that used

in WEPS, either do not perform well outside the regions in which they were developed, or suitable input data for the models are not available with regional to global scale coverage.

The major implication of this is that, without schemes to account for temporal variations in soil erodibility, the models cannot capture the true nature of temporal variations in land erodibility. They are reliant on representing land susceptibility to wind erosion through vegetation, soil moisture and soil textural controls. This limitation is particularly important in modelling wind erosion in dryland environments that are devoid of vegetation and soil moisture, for example over playas. In these circumstances land erodibility is controlled by the erodibility of the soil surface (Chapter 2, Section 2.4), and the models may significantly over- or under-estimate wind erosion activity based solely on soil textural characteristics.

3.5.2 Data Availability

The development of process-based models has created a requirement for high spatial and high temporal resolution model input data. The models also often have a requirement for hard-to-measure inputs such as soil textural characteristics, soil moisture content, and in particular vegetation characteristics like cover, height, density and frontal area. Where these data are available with regional to global coverage, their spatial and temporal resolutions are often too coarse to resolve local scale (e.g. $\sim 10^2$ km²) heterogeneity. This data availability issue has undoubtedly contributed to the lack of model application at moderate to high spatial resolutions and at the landscape scale.

Raupach and Lu (2004) describe methods for dealing with the lack of data required as input to the process-based models. Commonly used methods include: 1) simplifying the models such they can be run with available input data; and 2) constraining spatial patterns in the hard-to-measure parameters using correlated parameters for which spatial data is readily available. Developments in remote sensing of soil moisture content (e.g. Loew, 2008) and vegetation structural attributes (e.g. McGlynn and Okin, 2006) will undoubtedly improve our ability to account for these factors at high spatial and temporal resolutions. In the interim, however, the ability of the models to represent local-scale heterogeneity in dust source areas remains weak. Capturing this heterogeneity is important if models are to accurately represent dust emitting hot spots within dryland environments (Gillette, 1999; Okin, 2005).

3.5.3 Up-scaling Models and Sub-Grid Scale Heterogeneity

Models that account for the effects of conditions such as soil texture, soil moisture and vegetation cover are based on plot-scale wind tunnel experimentation. The relationships driving the models are therefore functional at these scales. Accuracy issues arise when the relationships are applied to coarse resolution spatial data and are used to examine wind erosion processes at the landscape to regional scales. The spatial data used as input to the models represents gridded averages of land surface conditions, including vegetation cover and soil moisture. These conditions are rarely homogeneous (Okin and Gillette, 2001). This issue links back to the nature of the model functions, which themselves may not adequately account for the non-uniform distribution of inputs (Raupach and Lu, 2004). Complicating the issue is the fact that the relationships between factors controlling land susceptibility to wind erosion are non-linear and display threshold-like behaviour (Gillette, 1999).

Shao (2000) reported on three methods for dealing with sub-grid scale variations in spatial modelling. These include: averaging surface properties, i.e. treating grid cells as homogeneous areas; representing sub-grid scale heterogeneity through multiple smaller homogeneous sub-grids; and using probability density functions to represent sub-grid scale heterogeneity. The latter approach has been adopted in a number of wind erosion models, including IWEMS and DEAD, to account for local variations in both land surface and meteorological conditions (Lu and Shao, 2001; Zender et al., 2003a). The approach has also been used to account for spatial variations in wind shear stress in a model to predict surface roughness effects by Okin (2008) (Chapter 2).

The implications of not accounting for sub-grid scale heterogeneity are that models are likely to significantly underestimate predictions of both land erodibility and dust emissions (Okin and Gillette, 2004; Okin, 2005). The issue may also complicate model validation as point observations of wind erosion activity that are typically used to validate wind erosion models do not necessarily reflect spatially averaged predictions.

3.5.4 Validation of Regional to Global Scale Models

Wind erosion models are most often validated by comparisons of output with point measurements of wind erosion (Shao et al., 1996; Fryrear et al., 1998; Gregory and Darwish,

2001; Zender et al., 2003a). This approach is surprising considering the diverse spatial scales at which wind erosion models have been developed to operate. It reflects difficulties in obtaining measurements of wind erosion activity over large areas, and also inherent difficulties associated with obtaining representative measurements of a spatio-temporally dynamic process. Global scale studies of dust transport are increasingly comparing model outputs with remotely sensed data on aerosol optical thickness to obtain spatial assessments of model performance (e.g. Mahowald et al., 2003a; Ginoux et al., 2004).

Validation of the model components is generally considered during their development (individually) or through validation of the complete wind erosion models (e.g. Marticorena and Bergametti, 1995). Explicit validation of model assessments of u_{*t} have not been reported. This reflects the fact that, despite the potential for doing so, the models have not been applied to assess land erodibility.

3.6 Summary

This chapter has presented a review of approaches for representing land erodibility in wind erosion models. The review has covered empirical and mechanistic models that have been applied to quantify soil loss and dust emission from the field to global scales. The chapter has reviewed limitations to wind erosion modelling, including: the availability of robust models to predict temporal variations in soil erodibility; the availability of suitable model input data; a requirement for robust methods to upscale models and account for sub-grid scale heterogeneity in soil and vegetation conditions; and a reliance on validating models at small point scales or very coarse resolutions due to a lack of data available to test the performance of models at intermediate (landscape to regional) scales. Most notably, and despite considerable modelling efforts, there has been a significant lack of research that seeks to apply models to assess spatial and temporal patterns in the susceptibility of landscapes to wind erosion. This is essential for accurately assessing spatial and temporal patterns in land erodibility, particularly in rangeland environments.

Considerable effort has been directed toward scaling issues associated with broad-scale spatial modelling (see Raupach and Lu, 2004; Shao, 2000) and dealing with sub-grid scale heterogeneity (e.g. Okin, 2008). Priority must therefore be given to addressing gaps in

landscape scale research to map and monitor land susceptibility to wind erosion. This can be achieved by:

- Developing models to predict temporal changes in soil surface conditions like surface crusting and aggregation that control soil erodibility.
- Developing models to assess land erodibility and wind erosion at the landscape scale.
- Developing field methods to assess land erodibility and wind erosion that can be applied to collect data for use in validating spatially distributed models at the landscape scale.
- Applying models to learn more about spatial and temporal patterns in land erodibility and landscape responses to climate variability and land management pressures.

The following chapters of this thesis systematically address these research requirements.

Chapter 4

A Framework for Modelling Temporal Variations in Soil Erodibility

This chapter addresses Objective 3 by presenting a framework for modelling temporal variations in soil erodibility. The framework draws on the systems analysis presented in Chapter 2 to characterise the temporal response of soils subject to variable precipitation and disturbance by livestock trampling, which are dominant controls on soil erodibility in rangeland environments.

4.1 Introduction

Global interest in climate change, desertification and land degradation has drawn increased attention to monitoring and modelling wind erosion processes in rangelands and cultivated environments (e.g. Hagen, 1991; Lu and Shao, 2001; Gregory and Darwish, 2001). Central to the development of functional wind erosion modelling systems is an understanding of how the mechanisms driving wind erosion interact and change through space and time in response to climate variability and land management. A recurring limitation to the development of wind erosion models is the absence of robust schemes to simulate temporal changes in soil surface conditions driving soil erodibility (Shao, 2000). This stems from the fact that the mechanisms driving changes in soil erodibility to wind are complex and are yet to be quantified in many environments (Bresson, 1995). Developing our understanding of processes forcing temporal changes in soil erodibility is therefore essential for advancing our understanding of wind erosion dynamics (Merrill et al., 1999).

Soil erodibility is defined as the susceptibility of a soil to mobilisation by wind. Soil erodibility is spatially variable and temporally dynamic and therefore exists within a continuum (Geeves et al., 2000). The erodibility of a soil is controlled by the aggregate size distribution (ASD) of soil grains, and in particular the availability of loose erodible grains (<

0.84 mm diameter) on the soil surface (Chepil, 1950a). The soil ASD is a function of the level of soil aggregation and physical or biological crusting (Zobeck, 1991). These properties are determined by soil texture, chemistry, organic matter content, and dynamic factors such as climate (e.g. rainfall, temperature) and land management (Figure 4.1). Soil moisture is an important control on soil erodibility and is a transient factor that may be considered independent of the soil inherent wind erodibility which is primarily determined by the ASD (Merrill et al., 1997).

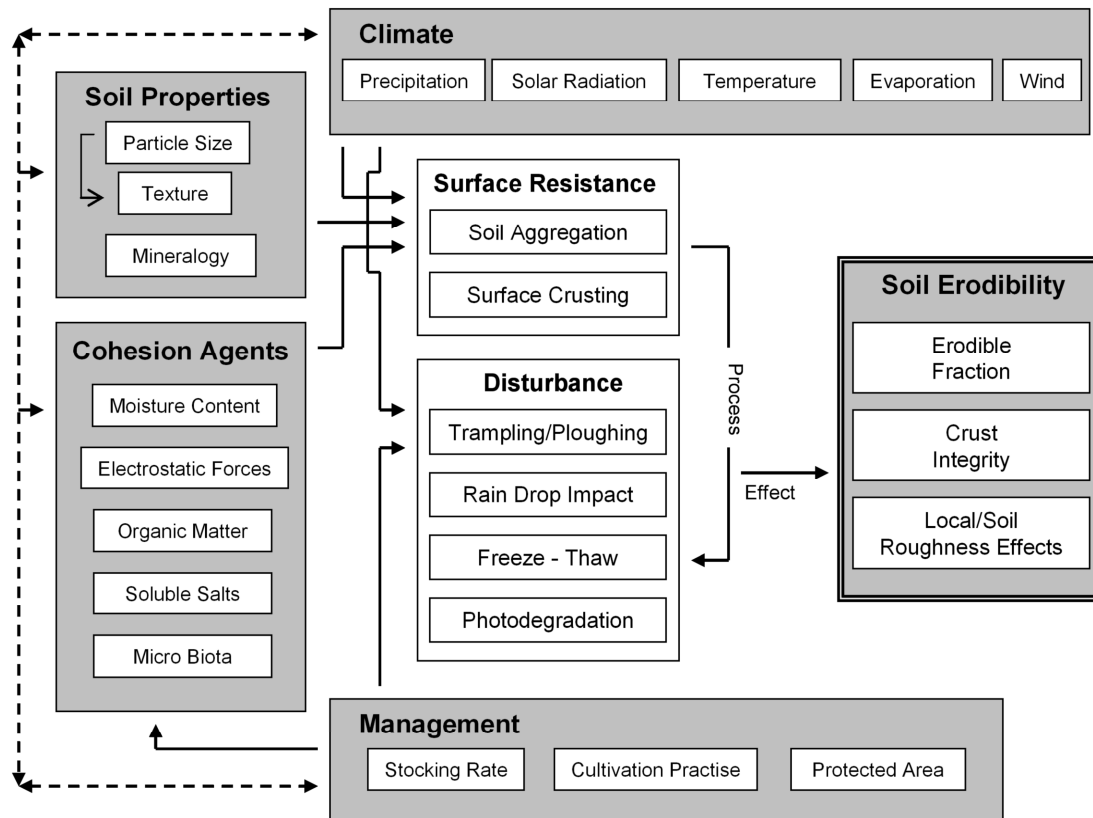


Figure 4.1 Flow chart illustrating the relationships between soil erodibility controls within a landscape. Grey boxes represent environmental conditions and processes that determine soil surface conditions and the impact of disturbance mechanisms on the availability of loose erodible material

The susceptibility of a land surface to wind erosion is highly sensitive to changes in soil erodibility. This sensitivity is most evident when crusted soil surfaces are disturbed by cultivation or trampling by livestock. Belnap and Gillette (1997), for example, reported a 73-92% decrease in the threshold friction velocity (u_{*t}) required for grain mobilisation on moderately disturbed sandy soils. Numerous studies have reported similar responses for a range of soils and types of surface disturbance (e.g. Williams et al., 1995; Leys and Eldridge,

1998). In terms of sediment transport potential, Eldridge and Leys (2003) reported a 4-fold increase in the streamwise sediment flux (Q , $\text{gm}^{-1}\text{s}^{-1}$ at 65 kmh^{-1}) for disturbed sandy soils (relative to the soil in a crusted condition), and a 26-fold increase in the streamwise sediment flux of disturbed loamy soils. Accounting for temporal changes in soil erodibility in wind erosion models is therefore critical for the accurate simulation of wind erosion processes.

As identified in Chapter 3, few wind erosion modelling systems contain schemes to compute temporal changes in soil erodibility. Field measurements of aggregate size distributions and empirical expressions relating soil texture (sand, silt and clay content), organic matter (OM) and calcium carbonate (CaCO_3) content have been used to account for soil erodibility in some models (Chepil and Woodruff, 1954; Fryrear et al., 1998; Singh et al., 1999; Böhner et al., 2003). Fryrear et al. (1994) developed a model to compute the wind erodible fraction of soils (aggregates $< 0.84 \text{ mm}$) using inputs of soil texture, OM and CaCO_3 . The model was adapted by Hagen (1991) to simulate temporal changes in soil erodibility for the Wind Erosion Prediction System (Chapter 3). Global application of the WEPS soil erodibility prediction scheme has been limited by the low availability of input spatial data for the OM and CaCO_3 model parameters, and the model applicability to soils outside North America (Leys et al., 1996). Established regional to global scale wind erosion models, such as the Dust Production Model (Marticorena and Bergametti, 1995), and Integrated Wind Erosion Modelling System (Lu and Shao, 2001) do not consider temporal changes in surface crusting or aggregation. This is due to the absence of soil erodibility models applicable at these scales, and yet is surprising given the dependence of global dust emissions on spatial and temporal variations in soil erodibility (Grini et al., 2005).

Building models to simulate temporal changes in soil erodibility is essential for the development of robust wind erosion modelling systems. In particular, this is of importance to increasing the skill of models like AUSLEM in assessing land susceptibility to wind erosion in rangeland environments. This chapter has three aims that seek to address this issue. The first is to draw on published research to develop a conceptual model of the soil erodibility continuum. The second aim is to establish a framework for modelling temporal changes in soil erodibility that could be integrated into a revised AUSLEM (Chapter 5). The framework characterises the temporal response of soils subject to variable precipitation and disturbance by livestock trampling that are dominant controls on soil erodibility in rangeland environments. The final aim is to use the model framework to highlight deficiencies in our

understanding of factors driving temporal changes in soil erodibility to wind. It is hoped that this will incite discussion and research action that is required to quantify relationships between controls on soil erodibility so that robust predictive models can be developed.

4.2 Aggregation, Soil Crusts and Soil Erodibility

Bagnold (1941) demonstrated by laboratory wind tunnel experimentation that as grain diameter decreases, u_{*t} decreases toward a minimum where grain diameter is ~ 0.08 mm. Further reductions in grain size result in an increase in u_{*t} . As described in Section 2.2.3, Iversen and White (1982) attributed the increase in u_{*t} for small grain sizes to enhanced inter-particle cohesion between fine (clay) grains. In field situations the effect of inter-particle cohesion is seen across all soil textures, with particle-bonding driving soil aggregation and surface crusting. Chepil (1950a) demonstrated that soil aggregate size is directly related to erodibility, with aggregates <0.42 mm being highly erodible, $0.42\text{--}0.84$ mm moderately erodible, and >0.84 mm non-erodible. The dry aggregate size distribution (DASD) thus drives the availability of loose erodible material on a soil surface. The physical mechanism behind the aggregation-erodibility relationship can be drawn back to grain (aggregate) size effects on u_{*t} , with increased aggregation resulting in an increase in surface roughness (z_0) and energy required for grain mobilisation. Once grain mobilisation has occurred, DASD and surface crusting will in turn affect the saltation load, abrasion efficiency, and potential dust production.

DASD characteristics are driven by processes of aggregate formation and breakdown (Breuninger et al., 1989). A primary control on aggregate formation is soil texture, with soil particle size and mineralogy affecting inter-particle bonding (Harris et al., 1966). Bond strength is subject to the nature of particle contacts. The plate-like structure of fine clay particles provides large inter-particle contact surfaces, and so a greater potential for bonding than in soils with sub-rounded or angular particles, e.g. sands (Smalley, 1970). As sandy soils are least susceptible to aggregate formation, they are also the most consistently erodible (Chapter 2, Section 2.2.4). As clay soils have the most potential for aggregate formation, they display the greatest range of variability in aggregation and therefore erodibility (Chepil, 1954).

Physical and biological crusts both influence soil erodibility (Chapter 2, Section 2.2.6). Physical crusts are identified by their mechanisms of formation: structural crusts form as a result of water droplet impact (e.g. during rainfall) and *in situ* particle rearrangement, whereas depositional crusts form by the settling of fine particles transported by surface water (Valentin and Bresson, 1992). Microbiotic crusts are characterized by the biological organisms driving crust formation, e.g. mosses, liverworts, lichens, algae and cyanobacteria. These bind soil particles with structural filaments or through the secretion of gels, and influence the availability of loose erodible material, soil micro-topography and z_0 (Eldridge and Greene, 1994; Johansen, 1993). Additional crust characteristics that influence soil erodibility include: lateral cover; structure (arrangement of particles by grain size); thickness; strength; and modulus of rupture, the resistance to breaking by saltating particles (Rice et al., 1999; Figure 2.6).

Moisture is the principal agent driving the initiation of inter-particle bonding in aggregate and crust formation (Harris et al., 1966, Section 2.2.5). The processes are therefore particularly sensitive to climate variability. The amount, intensity and frequency of precipitation, soil texture and site stability determine the type of crust formation and productivity of soil microbiota (Belnap and Eldridge, 2003). Precipitation efficiency is regulated by solar radiation intensity and potential evaporation. Field studies of freeze-thaw induced changes in the aggregate size distribution of soils in North America have demonstrated the added importance of temperature in regulating soil aggregation (Chepil, 1954; Bisal and Ferguson, 1968; Merrill et al., 1999; Bullock et al., 2001).

Mechanical disturbance of the soil surface by machinery and livestock trampling is a fundamental control on DASD and crust characteristics. In cultivated landscapes soil aggregation is affected by tillage practices, while in the rangelands aggregation is related to crust disturbance levels (Eldridge and Leys, 2003). Additional factors affecting aggregate and crust breakdown include: photo-degradation during long periods of exposure (i.e. during drought); clay lattice expansion-contraction during wet-dry cycles (e.g. in self-mulching soils); freeze-thaw cycles; fire; and salt efflorescence (Harris et al., 1966; Breuninger et al., 1989). Aeolian abrasion associated with saltation bombardment during wind erosion further affects DASD and crust integrity (McKenna-Neuman et al., 2005). These disturbance mechanisms are superimposed on the factors driving aggregate and crust formation, resulting

in soil erodibility being a spatio-temporally dynamic condition that varies through a continuum at scales from $10^{-1} - 10^6$ m over minutes to years (Figure 4.1; Figure 2.10).

4.3 The Soil Erodibility Continuum

Chepil (1953) demonstrated that soil clay content is a good texture-based predictor of soil erodibility, with clay content up to 15% of the soil weight being extremely important in increasing aggregation and decreasing soil mobility. The relationship between soil clay content (percentage clay) and streamwise sediment flux Q ($\text{gm}^{-1}\text{s}^{-1}$) was reported as:

$$Q = a.(\%clay)^b . c^{(\%clay)} \quad (4.1)$$

where a , b and c are regression coefficients with the values 63 095, -5.1 and 1.23 respectively. An erodibility minimum defined by Equation (4.1) was found to occur in soils with 15 – 27% clay. The high erodibility of soils with <15% clay was attributed to a lack of silt and clay particles to facilitate inter-particle bonding (Section 2.2.3). The low erodibility of loamy soils was attributed to enhanced bonding of particles (aggregation), but not to the extent seen in soils with higher (>27%) clay content that may suffer from granulation and aggregate breakdown (Chepil, 1953).

Leys (1991b) reported on u_{*t} and soil flux rates for selected Australian soils. The study was extended by Leys et al. (1996) to better detail percentage clay and dry aggregation effects on sediment flux for cultivated and non-cultivated (crusted) soils. The study reported similar patterns in the relationship between percentage clay and erodibility to those described by Chepil (1953). However, a significant difference in the results was evident with the increased erodibility of soils with >27% clay not found in the Australian soils. The elevated erodibility for soils with high clay content (modelled by the $c^{(\%clay)}$ tail in Equation 4.1) occurs as a result of a breakdown in soil structure (i.e. crusting and aggregation) induced by freeze-thaw processes affecting the North American test soils used by Chepil (1953). Without this tail the percentage clay- Q relationship for Australian soils can be effectively modelled by a power function of the form:

$$Q = a.(\% clay)^b \quad (4.2)$$

where a and b are regression coefficients representing the intercept and rate of change in Q with respect to percentage clay.

While Leys et al. (1996) presented two erodibility curves relating to two surface treatments (Figure 2.4), Chepil (1953) presented a single regression equation fitted to data collected on soils that had experienced a range of antecedent climate and management conditions. The elevated erodibility of soils with clay content $>27\%$ assessed by Chepil resulted from crust/aggregate breakdown – the same process separating the two sets of data presented by Leys et al. (1996). The data presented by Leys et al. (1996) demonstrates that a power function (e.g. Equation 4.2) will best model the condition of ‘minimum erodibility’ in soils that have maximum aggregation or crusting. In situations where soil moisture content is low enough not to affect u_{*t} , the equation for crusted soils can be taken to define the lower limit of the soil erodibility continuum.

Eldridge and Leys (2003) reported on biological crust cover-aggregation relationships in rangeland environments. Their results demonstrate that decreasing crust cover results in lower dry aggregation levels by a linear relationship of the form:

$$\%DA = 18.03 + 0.79(\%CC) \quad (4.3)$$

where $\%DA$ is the percentage mass of dry aggregates >0.85 mm, and $\%CC$ is the percentage biological crust cover ($r^2 = 0.72$, $p < 0.001$). The relationship was found to vary for different soil textures. While clay and loam soils were found to require low levels of crust cover to maintain high $\%DA$, sandy soils rely heavily on crust cover to provide adequate soil aggregation to minimize erodibility. Severe disturbance of crusts on loamy and clay soils is therefore required to instigate a significant increase in erodibility.

Comparing the Leys et al. (1996) regression models for the cultivated and non-cultivated soils reveals the effects of disturbance on soil aggregation, crusting and erodibility. With aggregate breakdown (increasing $\%DA < 0.85$ mm) a vertical displacement of the curve defined by Equation (4.2) takes place. For example, the curve will move from a low position

for crusted soils to a higher position for disturbed soils (raw data shown in Figure 4.4). This is physically supported by changes (decrease) in u_{*t} in response to surface disturbance measured for a range of agricultural and desert soils in the United States (Gillette, 1980; Gillette et al., 1982; Gillette, 1988).

Mathematically, movements of the erodibility curve for changes in aggregation and crusting are defined by a range of values for the power b in Equation (4.2). In terms of soil clay content then the soil erodibility continuum can be defined by the range:

$$Q_{\min} = a.(\%clay)^{b_{\min}} \rightarrow Q_{\max} = a.(\%clay)^{b_{\max}} \quad (4.4)$$

where $b_{\max}$ defines the erodibility maximum and $b_{\min}$ defines the erodibility minimum.

The availability of loose erodible sediment ultimately drives changes in the position of the curve between those at $b_{\max}$ and $b_{\min}$. The directional form of the change is therefore controlled by the relationship between soil aggregation (namely %DA > 0.84 mm) and streamwise sediment flux Q . Chepil (1953) and Leys et al. (1996) demonstrated that this relationship is consistent for all soil textures (i.e. 0 – 100 % clay) and can be expressed as:

$$Q = a \exp^{-b\%DA} \quad (4.5)$$

where a is the equation intercept and $-b$ defines the sensitivity of Q to %DA, the percentage mass of non-erodible dry aggregates (>0.84 mm). The applicability of Equation (4.5) across the soil texture range links back to the energy requirements for grain mobilization and the elementary relationship between grain (aggregate) size and u_{*t} . Progressively increasing soil surface disturbance results in weakened inter-particle bonds and an increase in the fraction of loose erodible material. Soils can therefore exist with some range of ‘grain’ sizes determined by inter-particle cohesion as a result of aggregation or crust disturbance (Breuninger et al., 1989). As soil aggregation changes through time in response to climate and management factors the position of a soil on the curve defined by Equation (4.5) will also change. Figure 4.2 illustrates a hypothetical vertical displacement of the Q -%clay curve (4.2b) in response to a change in %DA (4.2a). Differences in the regression coefficients a and $-b$ (Equation 4.5) as presented by Chepil (1953) and Leys et al. (1996) are attributed to site specific variations in

soil physical, chemical and biological properties and the density/specific gravity of the aggregates.

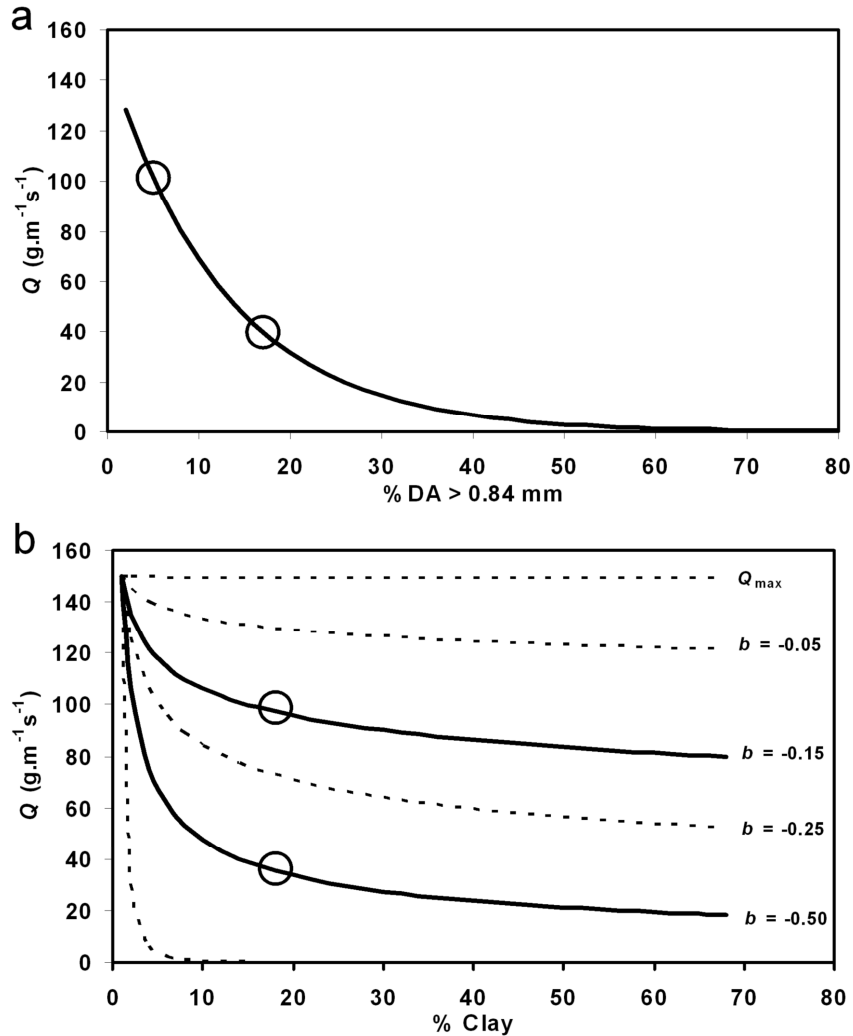


Figure 4.2 Graphs illustrating the effect of (a) changing the non-erodible fraction of a soil ($\% \text{ DA} > 0.84 \text{ mm}$) on (b) the position of the curve defining the relationship between soil clay content (percentage clay) and soil erodibility as indicated by the streamwise sediment flux (Q , represented by circles) at a wind velocity of 65 kmh^{-1} (after Leys et al., 1996).

Given that changes in soil erodibility can be defined by a shift in position of the curve defined by Equation (4.2), and the nature of this shift is determined by the $\% \text{ DA} - Q$ relationship defined by Equation (4.5), the physical dimensions of the soil erodibility continuum can now be described. To start, an expression defining the minimum erodibility limit needs to be selected. Here a power function of the form of Equation (4.2) was fitted to the data presented by Leys et al. (1996) for non-cultivated (crusted) soils. The equation ($r^2 = 0.78, p < 0.01$) is of the form:

$$Q = 737.43(\%clay)^{-2.34} \quad (4.6)$$

An assumption made in selecting this limit is that the crust conditions on the soils measured by Leys et al. (1996) are representative of the minimum erodibility of those soils, i.e. lowest fraction of loose erodible material and dry soil surface. This model operates under the assumption that soil moisture content is equal to zero, with soil erodibility being controlled by the soil aggregate size distribution.

The maximum erodibility limit now needs to be defined. The highest values for Q obtained by Leys et al. (1996) for sandy soils were around $150 \text{ gm}^{-1}\text{s}^{-1}$ at a wind velocity of 65 kmh^{-1} . For the current study this value will be considered representative of Q_{max} for all soil textures. An assumption made in setting this limit is that given the right climate and land management conditions the dry aggregate size distribution of any soil could potentially match that of the sandy soil producing Q at $150 \text{ gm}^{-1}\text{s}^{-1}$. In practice Q_{max} will not be constant for all soil textures. While sandy soils display high erodibility and a relatively small range of variation due to aggregation, soils with high clay content may experience a wide range of aggregation. The extent to which erodibility can increase for soils with high clay content is dependent on the nature and severity of disturbance mechanisms. If a hypothetical maximum disturbance were reached on clay textured soils, a decrease in surface roughness may result, causing an increase in u_{*t} and decrease in erodibility by the nature of that measured by Bagnold (1941) - relating to increased inter-particle cohesion in the very fine clay grains ($< 0.08 \text{ mm}$). Furthermore, the minima in measured soil erodibility at $\sim 27\%$ clay due to strong grain binding and aggregation in some clay loam soils (e.g. Chepil, 1953) would likely cause a dip in Q at around this point and moving up the continuum.

A three-dimensional plot can be produced from Equations (4.4) and (4.5) to conceptualise the physical dimensions of the soil erodibility continuum (Figure 4.3). Values of Q computed with inputs of clay content from 0 to 100% can be matched to values of Q computed using Equation (4.5) for a full range of soil aggregate size distributions ($\%DA > 0.84 \text{ mm}$). Here the coefficient $a = 150$ based on Q_{max} , and $-b = -0.078$ ($r^2 = 0.99$, $p < 0.01$) based on Leys et al. (1996). The lower limit for Q is defined by values computed from Equation (4.6). The model operates on the premise that the percentage of dry aggregates $> 0.84 \text{ mm}$, the non-erodible fraction, determines soil erodibility. The plot illustrates the potential erosion rates

experienced for soils under a range of potential aggregation levels. The model suggests that under heavy disturbance soils with high clay content may have erodibility levels similar to soils with low clay content that are not disturbed. This is supported by Chepil's (1953) data in which soils with 60% clay had the same erosion rates (~ 12 tons/acre) as soils containing 8% clay. It is important to note that as soil moisture content increases, it will also affect the position of a soil on the erodibility continuum. When the soil moisture content is sufficient to induce an increase in u_{*f} , soil erodibility will decrease further toward (and will reach) a value of zero.

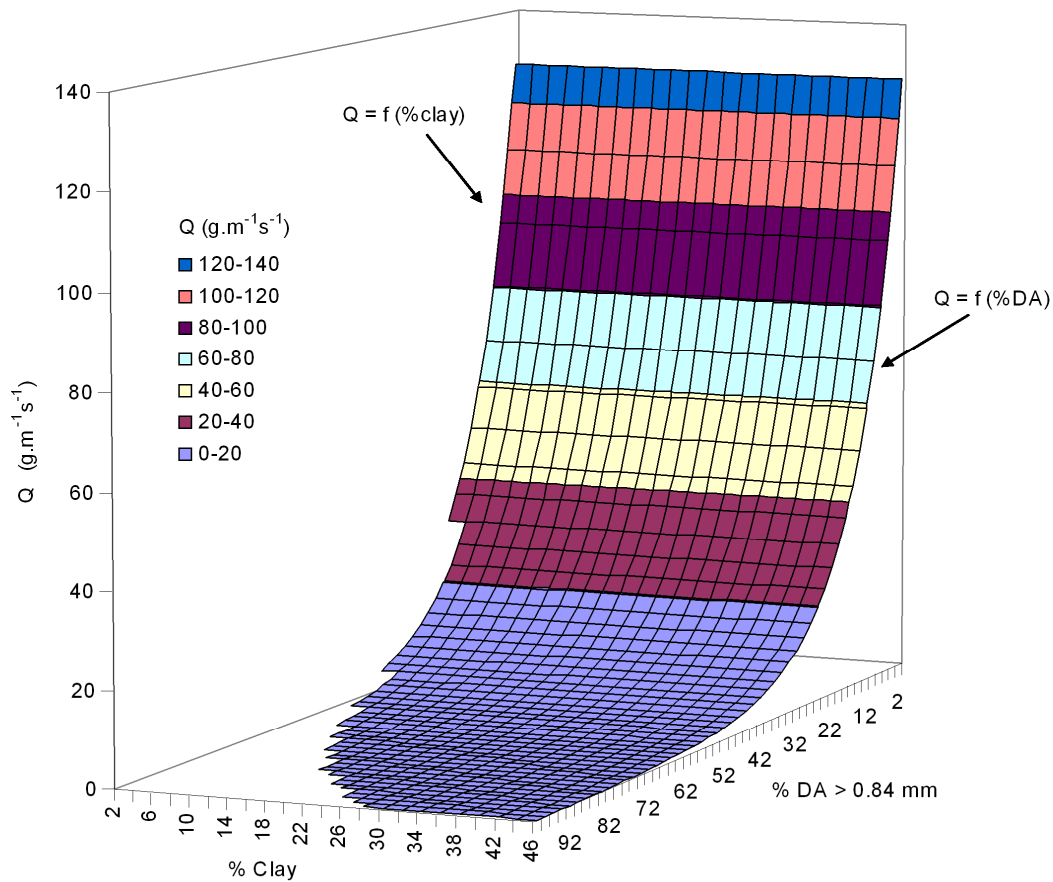


Figure 4.3 3D plot of the soil erodibility continuum as defined by the soil erodibility (indicated by $Q \text{ gm}^{-1}\text{s}^{-1}$) relationship with soil clay content (percentage clay) and aggregate size distribution that controls the quantity of non-erodible soil aggregates (%DA > 0.84 mm)

The soil erodibility continuum model can be used to describe changes in soil erodibility through time. A sandy soil with less than 7% clay may become resistant to wind erosion with stabilization due to biological crust formation. Crusts on sandy soils are, however, weak, abrade easily under saltation bombardment, and so offer a small reduction in soil erodibility.

Wind tunnel experimentation has shown that sandy soils are typically highly erodible under a range of conditions. The model suggests that even on crusted sandy soils the availability of loose erodible material will be sufficient to initiate saltation. This is supported by wind tunnel measurements of Q for crusted sandy soils (e.g. Belnap and Gillette, 1997). Soils with clay content greater than 50% may experience a greater range of variability in susceptibility to wind erosion (Chepil, 1954; Skidmore, 1994). This variability is driven by the temporal evolution of the surface particle size distribution and therefore loose erodible material that results from crust and aggregate formation and breakdown.

4.4 Modelling Temporal Changes in Soil Erodibility

4.4.1 Approach

Temporal changes in soil erodibility can be modelled by: 1) predicting soil aggregation levels (e.g. DASD or %DA >0.84 mm) then using that as input to Equation (4.5), or; 2) predicting the lateral cover of surface crusts and using that as input to Equation (4.3), which can then be input to Equation (4.5). These approaches are similar to those used in the WEPS model (Hagen, 2001; Visser et al., 2005; Chapter 3, Section 3.2.3). However, development and application of these approaches is restricted by the lack of quantitative data on factors controlling crust cover and dry aggregate size distributions in rangeland environments. The approach presented here seeks to characterise the form and rate of temporal changes in soil surface conditions between the states of minimum and maximum erodibility.

4.4.2 Temporal Model Framework

Figure 4.1 illustrates the relationships between mechanisms controlling soil erodibility. The mechanisms can be placed into three groups. The first group includes climatic factors that in the first instance may act in reducing soil erodibility. The dominant factor in this group is rainfall, the effects of which are moderated by solar radiation intensity, air temperature, evaporation rates and windiness. The second group are related to management, and include factors that may increase the susceptibility of a soil to wind erosion. In rangeland environments the dominant factor in this group is the stocking rate (animals per unit area) which drives disturbance (trampling) intensity and crust/aggregate breakdown. Climatic

factors may also play a significant role in soil aggregate destruction, for example driving photo-degradation of biological crusts and through freeze-thaw process. The third group include soil textural properties and cohesion agents. This group is influenced by the climate and management conditions, but also contains factors that are particular to specific soil types, e.g. moisture holding capacity. Movement of a soil through the erodibility continuum can be modelled through an expression of the behaviour of the soil in response to these forcing mechanisms (Chapter 2, Section 2.4).

Figure 4.4 illustrates the movement of a soil through the erodibility continuum. The movement can be considered to follow three phases, labelled *i*, *ii* and *iii*. The mechanisms controlling the position of a soil within the continuum will vary for each phase.

The first phase (*i*) defines a condition of minimum erodibility. Sufficient rainfall to induce soil surface sealing will result in a breakdown of dry aggregates and the consolidation of surface material in a saturated matrix (Kemper et al., 1987, Maulem et al., 1990). At this point reorganisation of the grains may take place, forming structural or depositional crusts and potentially rejuvenating biological crust growth (Valentin and Bresson, 1992). In terms of the erodibility continuum defined by Equation (4.4), this phase positions a soil at $Q_{\min}$, defined by $b_{\min}$, and controlled primarily by climatic factors (Figure 4.1). At the cessation of rainfall the control on erodibility will shift to being dominated by soil moisture, and the soil will remain at around the position of $Q_{\min}$ until the moisture content lowers to a position defined by Fécan et al. (1999) as w' , the minimum moisture content required to induce an increase in u_{*f} . Here the soil textural properties become important. Erodibility will remain constant during phase (*i*) for soils that seal and form physical crusts. For sandy soils an increase in erodibility will occur during this phase. This increase in erodibility can be expressed by a power function that defines a decrease in the ratio of u_{*fw} for the wet soil to u_{*fd} for the soil in a dry condition with decreasing soil moisture content (Fécan et al., 1999). The period of time that a soil remains in this phase is therefore determined by its physical properties and cohesion agents; this is likely to be shorter for well drained sandy soils than for soils with higher clay content that have higher water holding capacity (Cornelis and Gabriels, 2003).

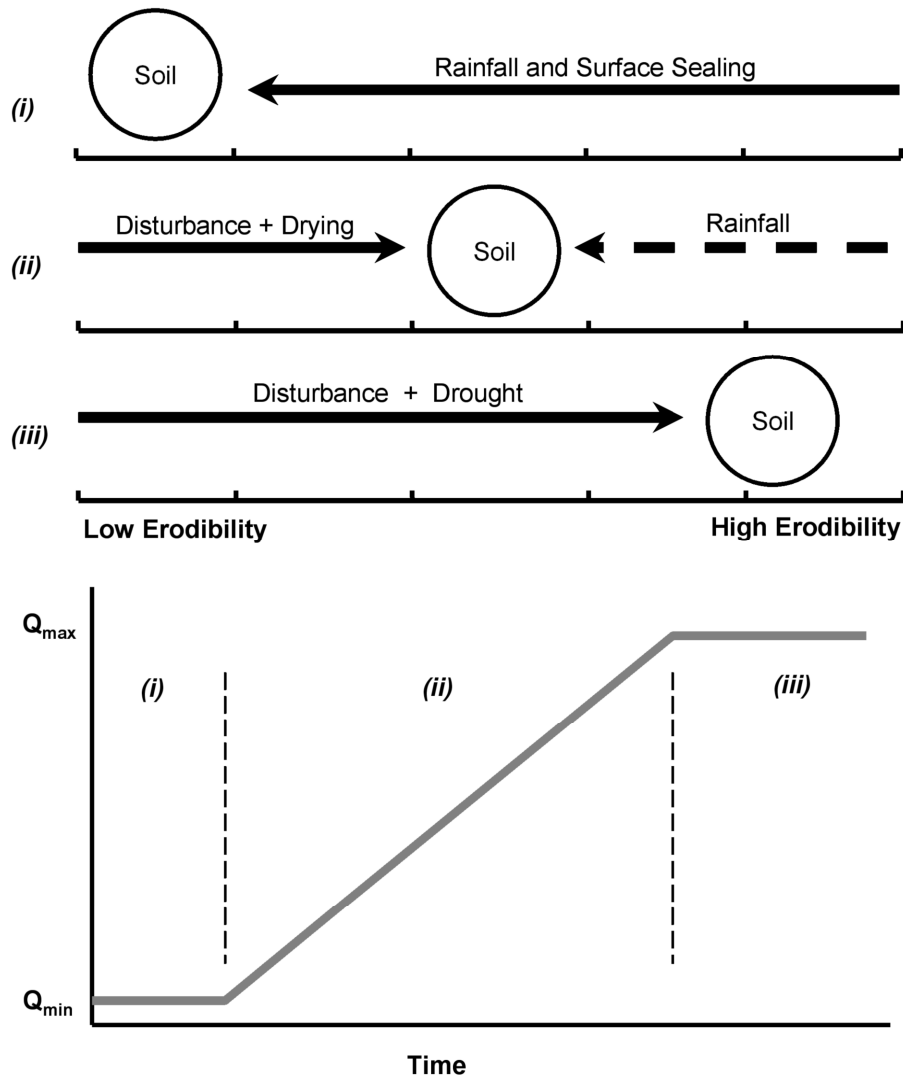


Figure 4.4 A conceptual diagram of the movement of a soil through the erodibility continuum from minimum ($Q_{\min}$) to maximum ($Q_{\max}$) erodibility states. The diagram indicates three phases of movement: (i) a condition of minimum erodibility, (ii) a transition phase of increasing erodibility, and (iii) a condition of maximum erodibility. The period of time a soil remains in each phase is determined by its textural properties, climate and management conditions

The second phase (ii) defines the transition of a soil through the continuum from a condition at $Q_{\min}$ to potentially arriving at $Q_{\max}$. This transition phase has a rate of change that is defined by the complex interaction of soil surface drying/desiccation and trampling by livestock (disturbance) which induce an increase in erodibility. During this phase rainfall events may temporarily increase the soil moisture content and aggregation and decrease erodibility. At the root of this transition phase, soil textural properties (i.e. sand, silt and clay content) and cohesive agents (e.g. OM, CaCO_3) will influence the soil response to the climate

and management factors, the slope gradient or rate of change in erodibility, and therefore the minimum period of time a soil may spend in this phase moving toward $Q_{\max}$.

Gillette (1978) reported on the dependence of soils on drought to experience an increase in erodibility to wind (Chapter 2, Figure 2.5). His results show that the erodibility of sandy soils with <15% clay is relatively unaffected by drought, soils with 15-40% clay are moderately dependent on drought to become erodible, and soils with >40% clay are highly dependent on drought to experience an increase in erodibility. This time dependence is driven by the soil particle size distributions and particle shapes which affect the strength of inter-particle bonds.

The final phase in the continuum, (*iii*) defines the condition of maximum erodibility ($Q_{\max}$). In order for a soil to reach this condition moisture content (antecedent rainfall) must be at a minimum and disturbance to the soil surface at a maximum. Soils at $Q_{\max}$ can be considered to have an effective grain diameter d of 0.08 mm to position the soil at the u_{*t} minima (Chapter 2, Figure 2.3) reported by Bagnold (1941).

Mathematically the phase shifts through the soil erodibility continuum, from *i* to *iii* (Figure 4.4), can be defined by a logistic (sigmoid) curve. Numerous expressions are available that can be used to delineate the shape of this curve (e.g. Richards, 1959; Turner et al., 1976). The logistic curve form can be approximated with few parameters, making it a good starting point for modelling temporal changes in soil erodibility to wind. The model framework presented here uses the Richard's equation (Richards, 1959), selected for its wide application in modelling vegetation growth dynamics (Tsoularis and Wallace, 2002) and more recently dunefield activation and stabilisation (Hugenholtz and Wolfe, 2005). In differential form the Richard's equation is expressed as:

$$\frac{dN}{dt} = rN \left[1 - \left(\frac{N}{K} \right)^\beta \right] \quad (4.7)$$

where N is the population size, r is the growth rate, β is a positive real number and K is the population carrying capacity reached as $\lim_{t \rightarrow \infty} N(t) = K$. The integral of Equation (4.7) is then:

$$N(t) = \frac{N_0 K}{\left[N_0^\beta + (K^\beta - N_0^\beta) e^{-\beta r t} \right]^{1/\beta}} \quad (4.8)$$

where N_0 is the population size at time $(t) = 0$. To represent temporal changes in soil erodibility we can consider changes in the power b (Equation 4.4) to follow the logistic growth curve defined by Equation (4.8) and express the model as:

$$Q(t) = 150.(\%clay)^{b(t)} \quad (4.9)$$

where $Q(t)$ is the indicator of time-dependent soil erodibility ($\text{gm}^{-1}\text{s}^{-1}$), $\%clay$ is the percentage soil clay content and the power $b(t)$ is defined by:

$$b(t) = A + \frac{C}{\left(1 + T \exp^{-r(t-M)} \right)^{1/T}} \quad (4.10)$$

where A defines the lower asymptote ($b_{\min}$), C equals the upper asymptote ($b_{\max}$) minus the value of A , r defines the growth rate, M is the time of maximum growth, T affects near which asymptote maximum growth occurs, and t equals the unit time. Parameter values for A and C can be assigned to Equation (4.10) assuming that $b_{\min}$ can be defined by Equation (4.6) and $b_{\max}$ is defined by the power (-0.05) providing a hypothetical $Q_{\max}$ at $\sim 150 \text{ gm}^{-1}\text{s}^{-1}$:

$$b(t) = -2.34 + \frac{2.29}{\left(1 + T \exp^{-r(t-M)} \right)^{1/T}} \quad (4.11)$$

For the remaining parameters, the growth rate r can be defined based on soil textural properties and the disturbance intensity, and can be modulated by adjusting the parameters T and M if changes in disturbance intensity take place through time (Bullock et al., 2001).

Under this model, temporal changes in erodibility are constrained by the limits:

$$b_{\min} \leq b(t) \leq b_{\max} \quad (4.12)$$

where $b_{\min}$ defines the minimum erodibility occurring at $t = 1$, and $b_{\max}$ defines the maximum erodibility of a soil as $t \rightarrow \infty$. Using this framework, temporal changes in soil erodibility can then be regulated by rainfall occurrence and time since rainfall, which can be used to derive values of t and input into Equation (4.11).

Antecedent rainfall conditions have demonstrated links with wind erosion activity at both short (daily to weekly) and long (e.g. monthly to seasonal) time scales. The effect of antecedent rainfall on soil erodibility at short time scales is manifested through soil wetness (moisture) conditions that drive inter-particle binding (Belly, 1964; McKenna-Neuman and Nickling, 1989; Fécan et al., 1999). The period for which soil wetness will affect erodibility is dependent on the water holding capacity of a soil, as determined by the soil texture, chemical and biological properties and vegetation cover (Cornelis and Gabriels, 2003). While the asymptotic form of the erodibility growth curve could be used to cover soil moisture effects on erodibility, the model does not explicitly account for moisture effects on grain binding (see Section 4.6 for discussion). At long time scales (e.g. monthly to seasonal) antecedent rainfall has been shown to affect both soil crust condition (Strong, 2007) and wind erosion activity (Brazel and Nickling, 1986; Yu et al., 1993; Neil and Yu, 1994).

Considering application of the model for daily simulations of soil erodibility, we can denote t_{init} as the value for t (Equation 4.11) used to initiate a simulation, t_{prev} as the value for t on the previous simulation day, and t_{new} as the value for t on the simulation day. Values for t_{init} can be approximated by considering antecedent rainfall totals for long ($\sum R$) and short ($\sum r$) periods prior to a simulation day. Values for t_{new} can be updated from t_{init} (and t_{prev}) by considering short term rainfall totals up to and including the simulation day. Under this framework conditional statements can be used to assign values for t_{init} based on $\sum R$, for example:

$$\begin{aligned} t_{\text{init}} &= W && \text{for } \sum R < W_r \\ &= X && \text{for } W_r < \sum R < X_r \\ &= \dots && \dots \\ &= 1 && \text{for } \sum R > Y_r \end{aligned} \quad (4.13)$$

where W_r and $X_r \dots$ are antecedent rainfall ranges determining values for t_{init} based on $\sum R$. For rainfall above a threshold sufficient to initiate surface crusting (denoted Y_r), t_{init} will be set to 1 such that $b(t) = b_{\min}$ and the resulting soil erodibility is equal to $Q_{\min}$. In terms of Figure 4.3,

this process can be used to determine the starting position of a simulation on the logistic curve, and whether the soil erodibility is in a minimum (*i*), transition (*ii*) or maximum (*iii*) phase.

For subsequent time-steps in a simulation, t_{new} can be computed as a function of t_{init} (that becomes t_{prev}) and short term (e.g. weekly) antecedent rainfall ($\sum r$). This can be achieved using a conditional statement of the form:

$$\begin{aligned} t_{\text{new}} &= (t_{\text{prev}} + 1) && \text{for } \sum r = 0 \\ &= (t_{\text{prev}} - \alpha) && \text{for } \sum r > 0 \end{aligned} \quad (4.14)$$

where α represents an adjustment factor for t when antecedent rainfall leading up to a simulation day is > 0 . The effect of the adjustment factor (α) can be defined by:

$$\begin{aligned} \alpha &= \alpha_1 && \text{for } \sum r < w_r \\ &= \alpha_2 && \text{for } w_r < \sum r < x_r \\ &= \dots && \dots \\ &= \alpha_n && \text{for } \sum r > y_r \end{aligned} \quad (4.15)$$

where $w_r, x_r, \dots, y_r$ are antecedent rainfall ranges determining values for α , $\alpha_1, \alpha_2, \dots, \alpha_n$ based on $\sum r$. The effect of α is on increasing ($t_{\text{prev}} + 1$) or decreasing ($t_{\text{prev}} - \alpha$) soil erodibility in response to the rainfall effect on the soil surface condition. Under this scheme, a period with no rainfall that is longer than the period over which $\sum r$ is computed (e.g. 10 days) will instigate forward movement up the erodibility continuum toward Q_{max} . Conversely, small rainfall events may temporarily decrease erodibility through a shift back down the continuum. A series of small rainfall events or a single large rainfall event may be sufficient to reset the soil surface to a position at the bottom of the continuum (Q_{min}). While the logistic curve has lower and upper asymptotes, the limits b_{min} and b_{max} (Equation 4.12) are maintained to prevent erodibility predictions being affected by circumstances of sustained rainfall events (giving a large α), or prolonged drought that may result in a large t value and no effect on erodibility with following small rainfall events.

4.4.3 Sensitivity Testing

The model framework has two components that control temporal changes in soil erodibility. These are: 1) the logistic model parameters that control rates of increases in erodibility, and 2) the model rainfall thresholds (Equation 4.15) that control the soil sensitivity to rainfall and decreases in erodibility. Figure 4.5 provides some example logistic soil erodibility ‘growth’ curves to illustrate a range of model parameterisations affecting increases in erodibility:

- (i) $r = 0.5, M = 0.5, T = 0.5$
- (ii) $r = 0.09, M = 0.5, T = 0.5$
- (iii) $r = 0.04, M = 0.5, T = 0.5$
- (iv) $r = 0.1, M = 30, T = 0.25$
- (v) $r = 0.1, M = 50, T = 0.5$
- (vi) $r = 0.15, M = 0.5, T = 0.5$
- (vii) $r \rightarrow \text{dynamic}, M = 0.5, T = 0.5$

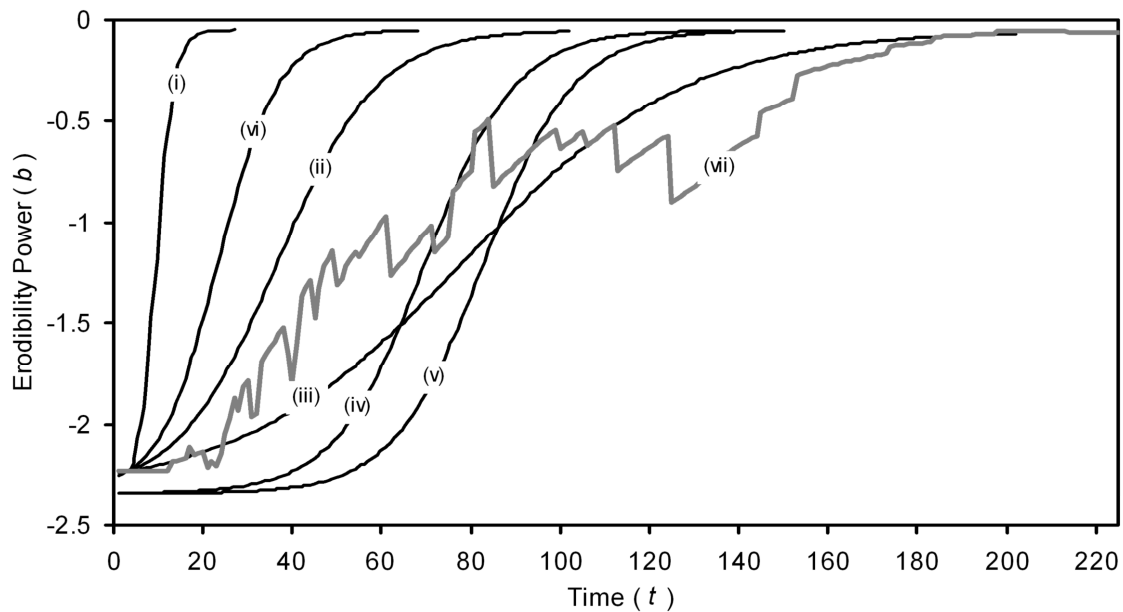


Figure 4.5 Graph illustrating the model sensitivity to changes in growth rate and growth timing parameters (i to vi), and model response to variable growth rates that can be expected under dynamic climate and management conditions (vii).

The parameterisations from (i) to (iii) demonstrate the model sensitivity to changes in the growth rate, r (Equation 4.11). Small variations in r are sufficient to induce significant

changes in the time it takes for a soil to reach a condition of maximum erodibility defined by $b_{\max}$. Parameterisations from (iv) to (vi) demonstrate the effects of varying the model components M and T that define the time of maximum growth.

In reality, temporal changes in soil erodibility are driven by dynamic variations in r , M and T . Therefore, a soil will not move up and down a single growth curve in response to rainfall and disturbance conditions. Rather, variations in the growth rate and disturbance intensities will induce an irregular/fluctuating growth pattern, for example (vii). Temporary increases in soil moisture and changes in aggregation and crust strength due to small rainfall events will induce additional variations in the growth curve. These effects will be further moderated by soil properties like organic matter content and climatic factors such as solar radiation intensity and evaporation rates. Soil erodibility dynamics could be modelled using static growth rates based on soil type dependence on drought to erode (e.g. after Gillette, 1978). However, the model output would not display realistic temporal patterns unless the effects of all dominant controls can be included in the growth rate formulation. The strength of the framework lies in this potential for incorporating a dynamic growth rate model to account for variations in soil responses to rainfall, drought and disturbance mechanisms.

Figure 4.6 demonstrates the model sensitivity to changes in the rainfall thresholds that drive decreases in erodibility. A hypothetical simulation was run with model parameters set for $r = -0.15$, $M = 0.5$, $T = 0.5$ (Equation 4.11). Rainfall thresholds (Equation 4.15) were set for:

- (a) $\sum r < 3, \alpha = 1; 3 < \sum r < 10, \alpha = -20; 10 < \sum r < 30, \alpha = -40; \sum r > 30, \alpha = -60$
- (b) $\sum r < 5, \alpha = 1; 5 < \sum r < 10, \alpha = -20; 10 < \sum r < 30, \alpha = -40; \sum r > 30, \alpha = -60$
- (c) $\sum r < 10, \alpha = 1; 10 < \sum r < 20, \alpha = -20; 20 < \sum r < 40, \alpha = -40; \sum r > 40, \alpha = -60$

Decreasing the model sensitivity to rainfall, from (a) through to (c), results in increases in the magnitude of output erodibility values and in the time for which a soil may have an elevated erodibility. Accurately defining both the model growth rates and rainfall sensitivity thresholds will therefore be essential if the conceptual framework is to be applied to model soil erodibility dynamics.

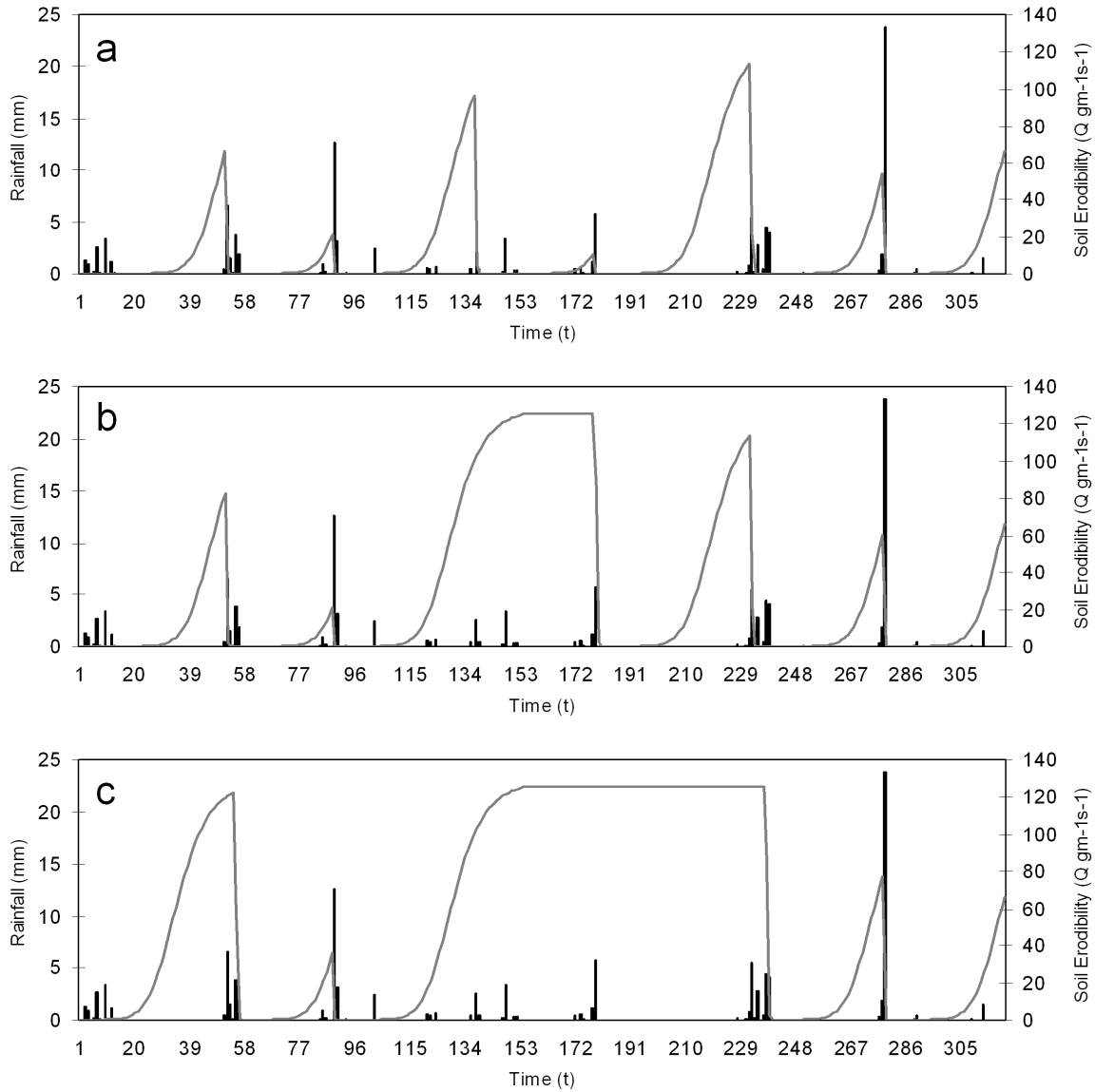


Figure 4.6 Graphs illustrating the effect of threshold changes that determine the model sensitivity to rainfall events (bars). Parts (a) to (c) illustrate decreasing model sensitivity to small rainfall events and subsequent increases in soil erodibility.

4.4.4 Model Limitations

The model framework has a number of limitations that are relevant to its future application. The primary limitations include:

Limitation 1. The model defines soil erodibility relative to Q , and not the threshold friction velocity for grain mobilisation (u_{*t}). Modern process-driven wind erosion models quantify land erodibility as a product of soil erodibility and surface roughness in terms of u_{*t} (e.g.

Marticorena and Bergametti, 1995; Lu and Shao, 2001; Hagen, 2004). Modelling soil erodibility terms of Q instead of u_{*t} limits integration of the framework into existing wind erosion models. Furthermore, soil erodibility predictions based on Q exist relative to a reference wind velocity (e.g. 65 kmh^{-1}) at which the wind tunnel experiments were conducted to obtain the model expressions (Equations 4.2, 4.3 and 4.6). This means that at higher or lower wind velocities the model may under- or over-predict soil erodibility.

Limitation 2. The model does not consider soil moisture effects on soil erodibility. An extensive body of research has been published on soil moisture effects on wind erosion (Cornelis and Gabriels, 2003). This research has considered soil moisture effects in terms of its influence on u_{*t} . Modelling soil erodibility in terms of soil crust and aggregate conditions, and separate from soil moisture, is legitimate when the soil moisture content is below the threshold water content that may induce an increase in u_{*t} , and therefore decrease in erodibility (Fécan et al., 1999). The model will fail in situations where the soil moisture content is above this threshold. Under these conditions $Q_{\min}$ will not be defined by Equation (4.6), but will be equal to zero. Integrating a factor to account for soil moisture effects on soil erodibility is feasible and could be achieved by adjusting the output by a ratio of erodibility under moist soil conditions to that under dry soil conditions.

Limitation 3. The model does not specifically account for the effects of biological soil crusts on soil erodibility. The responses of physical and biological crusts to rainfall and disturbance may vary significantly depending on antecedent conditions and disturbance types. Belnap and Gillette (1998) reported that disturbance to physical soil crusts can have a greater effect on increasing erodibility than disturbance of biological crusts due to a lower capacity to retain aggregation. Accounting for these effects under the model framework is beyond the scope of current research. In a spatial modelling context this may be addressed by incorporating factors to account for the likely distribution of biological crusts (e.g. Bowker et al., 2006) and differential rates of change in aggregation.

Limitation 4. The model considers only rainfall effects, and not freeze-thaw process effects on soil aggregate breakdown. While the focus of this paper has been to present a framework for modelling soil erodibility dynamics in hot, dry rangeland environments, the effects of freeze-thaw processes must not be ignored. The effects of freeze-thaw cycles on soil erodibility could be considered by the addition of a temperature component to the model

framework. To achieve this, quantifying temperature effects on soil erodibility at high temporal resolutions (e.g. daily sampling) is required to build on data presented in existing studies (e.g. Bisal and Ferguson, 1968; Anderson and Bisal, 1969; Bullock et al., 2001).

Limitation 5. The model does not consider (quantify) the availability of loose erodible sediment on the soil surface. This means that any crusted soil surface will be assigned a minimum erodibility and the model will not predict that crusted surfaces can erode under saltation bombardment processes. While saltation bombardment does not affect the immediate erodibility of a soil, it does affect the dust production potential. This has implications for modelling dust emissions in rangeland environments where crusted playa surfaces that are devoid of vegetation can be significant dust emitters.

The model limitations can be addressed by pursuing research to quantify climate and management effects on soil crusting, aggregation and erodibility to wind. Modelling soil erodibility in terms of u_{*t} , and integrating soil moisture effects into the soil erodibility model requires that we can first predict temporal changes in the soil aggregate size distributions. This would alleviate the issue of soil erodibility predictions being relative to a reference wind velocity, and would mean that the effects of freeze-thaw processes could be considered from a process-driven perspective. The first step required to address the model limitations is to examine application of the model and determine the effects of the limitations on the model performance.

4.5 Model Parameterisation

Evidence to support the model framework and parameterise the model components is scarce. Simulating temporal changes in the model growth rate parameter is required if the model is to accurately simulate soil erodibility dynamics. This is not possible with our current level of understanding of soil aggregation and crust dynamics, and would require making broad assumptions about the response of soils to climate variability and land management pressures. In particular, parameterisation of the model is dependent on a knowledge of how crust cover, thickness and strength and aggregate size and stability respond to rainfall, solar radiation, evaporation, trampling, and soil properties like texture, organic matter and salt content.

Considerable research has been conducted examining soil properties affecting wind erosion (reviewed in Chapter 2). The piecemeal nature of much of the research, however, has meant that it is difficult to integrate research findings to build models of ‘big picture’ processes. This characteristic is a result of the complex response of soil aggregation and crust dynamics to external drivers, and difficulties associated with extracting meaningful data on soil-climate-management interactions (Merrill et al., 1997). In particular this affects our ability to parameterise models to predict temporal changes in soil erodibility.

Table 4.1 summarises a selection of studies examining: a) soil aggregation changes in response to climate and management variability, b) soil crust disturbance effects on soil erodibility; and c) soil crust responses to trampling disturbance by livestock. Studies examining aeolian abrasion of crusts have not been included as these pertain to the process of erosion and dust emission as opposed to the immediate erodibility of a soil surface. Both qualitative and quantitative approaches have been used to examine temporal changes in soil erodibility, and passive monitoring and active manipulation of sites have been used to determine relationships between control and response variables.

Historically, soil aggregation responses to climate variability have been studied in cultivated regions where the economic and social consequences of severe wind erosion are well recognised. The majority of these studies have been conducted in North America and have focused on monitoring seasonal responses of soils to freeze-thaw cycles and under cultivation (Bisal and Ferguson, 1968; Merrill et al., 1999; Bullock et al., 2001). Standard methods for reporting on soil aggregation conditions, for example through aggregate size distributions, aggregate stability or the soil erodible fraction, have been adopted in many of these studies. This means that there is potential for comparing results between studies on different soil, management or climate conditions and parameterising a generalised model of soil erodibility response to climate, like that presented here. Few studies have used regression analyses to integrate relationships between factors controlling soil aggregation in empirical models (e.g. Zobeck and Popham, 1990; Fryrear et al., 1994; López et al., 2007).

Table 4.1 Summary of a selection of studies examining: (a) soil aggregation changes in response to climate and management variability; (b) soil crust disturbance effects on soil erodibility; and (c) soil crust responses to trampling disturbance by livestock.

(a)

Reference	Data Frequency	Sampling Period (months)	No. Soil Varieties	No. Cultivation/ Surface Types	No. Aggregation/ Crust Parameters	Surface Preparation	Erodibility Indicator
Bisal and Ferguson (1968)	Monthly	144	3	2	1	Field	Erodible Fraction
Gillette (1988) ^a	Monthly	14	52	10	5	Field	u_{*t}
Merrill et al. (1999)	Monthly	84	1	4	4	Active Field	Q
Bullock et al. (2001)	Monthly	8	1	3	3	Active Field	Erodible Fraction
Sarah (2005)	Annual	36	4	-	5	Field	-
Hevia et al. (2007)	Monthly	28	1	3	4	Active Field	Erodible Fraction

(b)

Reference	Crust Types (Phys./Biol.)	Surface Preparation	No. Soil/Crust Varieties	Disturbance Methods	Crust/Disturbance Measures	No. Crust Parameters	Erodibility Indicator
Belnap and Gillette (1997)	Biological	Active Field	4	Boot, Vehicle	Qualitative	-	u_{*t}
Leys and Eldridge (1998)	Biological	Active Field	2	Sheep Foot	Qualitative	2	Q, u_{*t}
Belnap and Gillette (1998)	Biological	Active Field	4	Vehicle, Livestock	Qualitative	-	u_{*t}
Eldridge and Leys (2003)	Biological	Active Field	2	Sheep Foot, Raking	Qualitative	4	Q
Belnap et al. (2007)	Biological	Active Field	5	Boot	Qualitative	3	u_{*t}
Gillette et al. (1982)	Physical	Field	44	Vehicle	Qualitative	1	u_{*t}
Leys et al., (1996)	Physical	Active Field	9	Cultivation	Qualitative	1	Q, u_{*t}
Rajot et al. (2003)	Physical	Field	1	-	Quantitative	2	Q
Goossens (2004)	Physical	Field	1	-	Quantitative	1	Q, u_{*t}

(c)

Reference	Crust Types (Phys./Biol.)	Sampling Type	No. Soil Varieties	Disturbance Methods	Disturbance Measures	Disturbance Measures	No. Crust Parameters
Hodgins and Rogers (1997)	Biological	Field	1	Livestock	Quantitative	Dung Density	4
Memmott et al. (1998)	Biological	Field	1	Livestock	Quantitative	Stocking Rate, Cultivation	2
Thomas and Dougill (2007)	Biological	Field	Not Specified	Livestock	Quantitative	Track Frequency, Dung Density	3
Williams et al. (2008)	Biological	Field	1	Livestock	Quantitative	Dung Density	2

A limitation to using data from studies of soil aggregation dynamics is that the temporal resolutions of these studies have varied from monthly to seasonal sampling, with data reporting at up to annual time scales (Table 4.1a). This coarse temporal resolution of sampling has provided snapshots in time of soil surface conditions. These temporally disparate samples have been pieced together to form soil erodibility time-series like that presented by Merrill et al. (1999), and have demonstrated dynamic responses of soils to inter-seasonal and inter-annual climate variability. Importantly, however, the resolution of the studies is insufficient to identify immediate responses of soil surface conditions to rainfall events or disturbance that may display the logistic behaviour in the model presented here.

Studies reporting on soil crusting effects on wind erosion have been conducted in both cultivated and rangeland environments (Table 4.1b). Unlike the studies monitoring dynamic soil aggregation effects on wind erosion, the studies of soil crusting effects have been short in duration. They have been experimentally based rather than monitoring soil crust-erodibility changes in response to climate variability and management. The focus of much of this work has been on understanding disturbance effects on biological soil crusts and resulting changes in soil erodibility. Soil crust disturbance levels in these studies have been reported using qualitative descriptors, e.g. ‘low’, ‘moderate’, ‘high’, as opposed to quantitative measures such as crust cover or crust strength (Belnap and Eldridge, 2001). These physical indicators of crust condition have not been consistently reported or acquired with standard measurement procedures (Table 4.1b).

Methods used to manipulate soil crust conditions have not been standardised. This has meant that a range of disturbance mechanisms have been used, including trampling in heavy boots to passing over the soil surface in a motorised vehicle (e.g. Gillette et al., 1982). While studies examining disturbance impacts on soil crust conditions (Table 4.1c) have used quantitative measures of disturbance intensity, these measures are proxies (e.g. animal dung density) for real disturbance intensities (e.g. stocking rates) that are required as input to spatially explicit models.

The characteristics of these studies have important implications for applying the research in parameterising models of soil crust effects on wind erosion. The coarse temporal resolution of monitoring studies (Table 4.1a), the lack of quantitative methods for defining soil disturbance intensities (Table 4.1b), and the poor transferability of disturbance methods and

measures to physical conditions that can be modelled (Table 4.1c) restrict our ability to link conditions of climate, soil types and disturbance regimes to soil aggregation, crust conditions and erodibility. This means that few models have been developed to predict temporal variations in soil erodibility (Hagen, 2004). Furthermore, there is insufficient evidence to support or parameterise the temporal model framework presented in this chapter, which requires data to define rates of change in erodibility in response to drought, disturbance, and rainfall. To address these deficiencies, research is required in three generalised areas:

- 1) In obtaining evidence to better define and support the physical boundaries of the soil erodibility continuum ($Q_{\min}$ to $Q_{\max}$).
- 2) In quantifying rates of increases in soil erodibility (growth rates) in response to drought and disturbance (of different types and intensities).
- 3) In quantifying precipitation effects on soil surface conditions under a range of surface pre-treatments (antecedent climate and disturbance regimes), soil types and climate (e.g. solar radiation intensity and evaporation rates).

Figure 4.7 illustrates environmental factors and measurement parameters that should be considered when designing experimental studies to quantify soil erodibility relationships with environmental dynamics. In addressing the research requirements, effort must be placed in conducting research at high temporal resolutions, across a range of soil types, and using methods that allow for results to be compared between studies and related to parameters that can be incorporated into spatially explicit models. Future research should extend from cultivated settings to include rangeland environments, which have received significantly less soil erodibility research attention over the last decade.

Merrill et al. (1997) described three approaches for building our understanding of soil erodibility dynamics. The first involves drawing information from the extensive existing databases on soil aggregation dynamics that have been collected in the United States. Secondly, research should continue to make use of passive monitoring of field conditions. This approach has particular application in describing soil erodibility dynamics under ‘natural’ field conditions, i.e. independently managed grazing lands. Thirdly, methods involving active manipulation of field conditions should be used to simulate changes in soil surface conditions. This approach may involve manipulating precipitation (amounts and frequency), vegetation cover, soil organic matter and disturbance types and intensities and

measuring their influence on soil aggregation and crust conditions (e.g. Bird et al., 2007). A benefit of this approach is that information gains can be accelerated as experiments are not dependent on longer-term variations in climate or management (Merrill et al., 1997). Exploring the application of remote sensing technologies to determine soil erodibility (e.g. Chappell et al., 2006; Chappell et al., 2007) may present further opportunities for monitoring erodibility dynamics over large areas.

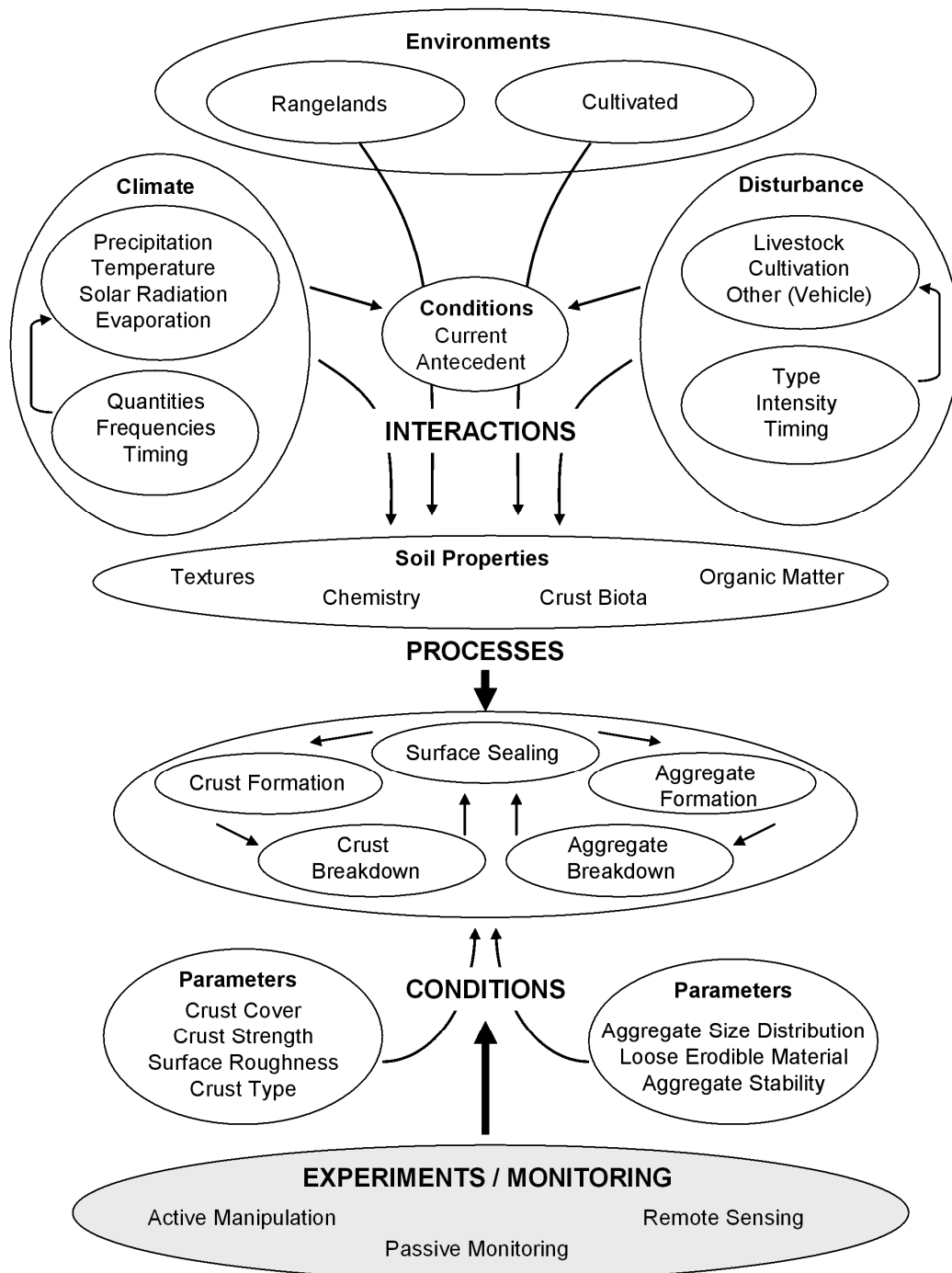


Figure 4.7 Flow chart illustrating factors that should be considered when designing new experimental studies to quantify soil erodibility relationships with environmental dynamics.

Designing field experiments to analyse soil erodibility dynamics would benefit from linkages to the model development process. Phillips (2008) reported on the application of models in deriving field-testable hypotheses that are independent of the original model. In this situation sensitivity tests could be used to determine the model response to parameterisations suiting particular environmental conditions. For example, the model could be used to simulate the effects of hypothetical stocking rates and rainfall conditions on soil erodibility dynamics. Field experiments could then be used to test hypotheses generated by the sensitivity analysis. This approach is particularly relevant for the parameterisation of the model framework presented in this chapter. The approach has multiple benefits of: providing direction for field research that would have direct application in model development; and providing baseline data that are required for the calibration, refinement and validation the model (Phillips, 2008).

4.6 Conclusions

This chapter had three aims. They were to: 1) draw on the underpinning science (reviewed in Chapter 2) to develop a conceptual model of the soil erodibility continuum; 2) establish a framework for modelling temporal changes in soil erodibility that could be integrated into a revised AUSLEM; and 3) highlight deficiencies in our understanding of factors driving temporal changes in soil erodibility to wind. Sections 4.2 and 4.3 of the Chapter developed a conceptual model of the soil erodibility continuum. Section 4.4 then presented a framework for modelling temporal changes in soil erodibility within the continuum. The model framework builds upon our existing knowledge of soil erodibility dynamics and offers an approach for assessing soil responses to variations in climate and land management conditions.

An absence of quantitative data on soil erodibility dynamics at high temporal resolutions (Section 4.5) restricted parameterisation of the conceptual framework. Alternate methods are therefore adopted to account for spatio-temporal variations in soil erodibility in the development of AUSLEM in Chapter 5. Significantly, the conceptual framework for modelling temporal changes in soil erodibility is transferable to modelling the land erodibility continuum (Chapter 2, Section 2.4). Temporal changes in land erodibility, as governed by soil moisture drying rates and vegetation growth and senescence, follow a logistic growth pattern like that described in this chapter (Hugenholtz and Wolfe, 2005). Subsequent

development of AUSLEM (Chapter 5) seeks to model these dynamics by upgrading the model functionality (after Webb et al., 2006) such that it captures this continual and non-linear response of land erodibility to changes in landscape condition.

A significant contribution of the model developed in this chapter is that it provides a framework that can be used to test and develop our understanding of the effects of static and dynamic controls on soil and land erodibility. In addition to this, the model framework is versatile in that it can be applied with varying levels of input knowledge and complexity. The development of models forces us to critically assess our understanding of environmental dynamics. This process has been important in highlighting the paucity of quantitative data relating soil properties, climate variability and land management conditions to changes in soil aggregation and surface crusting, which has restricted the development and parameterisation of models to predict temporal changes in soil erodibility, including that presented here. Future research extending beyond this thesis must use quantitative approaches in monitoring and experimentation to advance our understanding of soil erodibility dynamics. In particular, research must not lose sight of the requirement to integrate research outcomes into holistic models to simulate soil and land erodibility to wind.

Chapter 5

A Model to Predict Land Susceptibility to Wind Erosion in Western Queensland, Australia

This chapter addresses Objectives 4 and 6 by developing a functional land erodibility model that can be applied to assess land susceptibility to wind erosion across western Queensland, Australia. The model is validated at an annual temporal resolution through a comparison of point time-series records of wind erosion activity and model assessments of land erodibility between 1980 and 1990.

5.1 Introduction

Land degradation by wind erosion affects large areas of the world's arid and semi-arid lands (Oldeman, 1994; Lal, 2001). At the field scale ($<1 \text{ km}^2$), wind erosion has on- and off-site effects on crop growth, pastoral productivity, farm infrastructure and ecosystem health (McTainsh and Leys, 1993; Leys, 1999). There is a growing body of scientific evidence that dust emissions from wind erosion have numerous impacts at regional to global scales relating to: atmospheric radiation transfers; cloud nucleation; precipitation; biogeochemical processes; ocean nutrient availability, and others (e.g. Miller and Tegen, 1998; Bonasoni et al., 2004; Jickells et al., 2005). Central to the study of these processes are the requirements to monitor conditions giving rise to spatial and temporal variability in dust emissions and to integrate research across multiple spatial and temporal scales so that wind erosion mechanics can be better linked to regional and global studies of dust processes. Developing our understanding of the extent of areas susceptible to wind erosion is critical to the extension of our knowledge in these areas.

Globally, research that explicitly addresses wind erosion hazard mapping at moderate to high spatio-temporal resolutions is scarce. This is surprising given the extent of the wind erosion problem in dryland environments, and the focus of aeolian research into methods to control

wind erosion in marginal farming lands (Leys, 1999). Studies reporting the location of areas susceptible to wind erosion have foundations in land degradation surveys, analysis of dust storm frequencies and aerosol indices derived from satellite imagery, or present static erosion hazard maps based on soil texture or wind run. These methods have been applied extensively in Africa, North America, Europe, the Middle East and China (e.g. Lynch and Edwards, 1980; Kalma et al., 1988; Mezösi and Szatmári, 1998; Prospero et al., 2002; Shi et al., 2004). An important limitation of these methods is that they have not provided a means for assessing dynamic changes in land susceptibility to wind erosion at scales between the field and coarser regional scales (10^4 km^2). In Australia reports of the extent of wind erosion have been produced from assessments of landscape condition during land degradation episodes (e.g. Ratcliffe, 1937, Carter, 1985). The survey methods tend to provide snapshots of erosion hazard which reflect the regional climate at the time of survey (i.e. drought), and may not account for spatial and temporal variability in erosion controls. While numerous wind erosion modelling systems have been developed, rarely have the models been applied with the express purpose of monitoring spatio-temporal variability in areas susceptible to wind erosion. This variability can be captured through modelling and is critical for identifying wind erosion “hot spots” that are significant dust emitters (Gillette, 1999).

The development of models to assess land susceptibility to wind erosion should be seen as being essential to supporting decisions about the management of dryland environments (e.g. Bhuyan et al., 2002; Bowker et al., 2006). This also applies to the development and application of models to assess water erosion (e.g. Berlekamp et al., 2007; Miller et al., 2007). Modelling provides the opportunity to examine spatial and temporal patterns in erosion dynamics within landscapes, at different spatial and temporal resolutions and across scales. Spatially distributed models can be used to establish benchmarks of historical variations in the landscape response to climate variability and land management pressures, and to provide measures of the sensitivity of landscapes and waterways to climate and land management changes (Tegen and Fung, 1995; Baigorria and Romero, 2007). These are both requirements for enhancing land management policy. The need for research of this nature is growing, especially in sub-tropical environments in which rainfall amounts and variability, that control erosion processes, are expected to be affected by future climate change (Meehl et al., 2007).

The development of models provides a means for assessing and extending our understanding of wind erosion processes across multiple spatial and temporal scales. There are few accessible methods for modelling or mapping spatio-temporal patterns of wind erosion hazard in Australia at moderate to high spatio-temporal resolutions ($< 10^3$ km²; daily – monthly). The development of the Integrated Wind Erosion Modelling System (IWEMS) has provided a resource for addressing this deficiency, but the model is currently only applicable at a moderate (5 x 5 km) spatial resolution within south-eastern Australia (Lu and Shao, 2001). Webb et al. (2006) presented a spatially explicit Australian Land Erodibility Model (AUSLEM) for predicting land susceptibility to wind erosion. The model was applied at a 5 x 5 km spatial resolution on a monthly time-step across Australia. Qualitative comparisons of model output with an index of wind erosion activity indicated that the model could hind-cast erosion hazard with a reasonable level of success. However, the model used a threshold-based rule-set which tended to bias output predictions toward areas situated in a small range of soil textures.

This chapter addresses the limitations of the AUSLEM rule-based modelling system by deriving a new scheme to predict land erodibility. The chapter presents a description of the revised land erodibility model and evaluates model performance through a comparison of output and observational records of wind erosion activity. The focus of this research was to develop AUSLEM to predict land susceptibility to wind erosion in western Queensland, Australia.

5.2 Study Area

Analyses of long term dust-event frequencies in Australia indicate that the eastern half of the continent is the most active wind erosion region (McTainsh et al. 1990). Model development and validation were carried out for the northern part of this region, in western Queensland. Figure 5.1 provides a map of the study area, as described in Chapter 1, Section 1.6. The map shows the extent of the four bioregions covering the western Queensland rangelands and the location of meteorological stations from which observational records of dust events, used in this chapter, were acquired. Wind erosion is infrequently observed to the east of the study area due to higher annual rainfall and vegetation cover, so that area is not considered here.

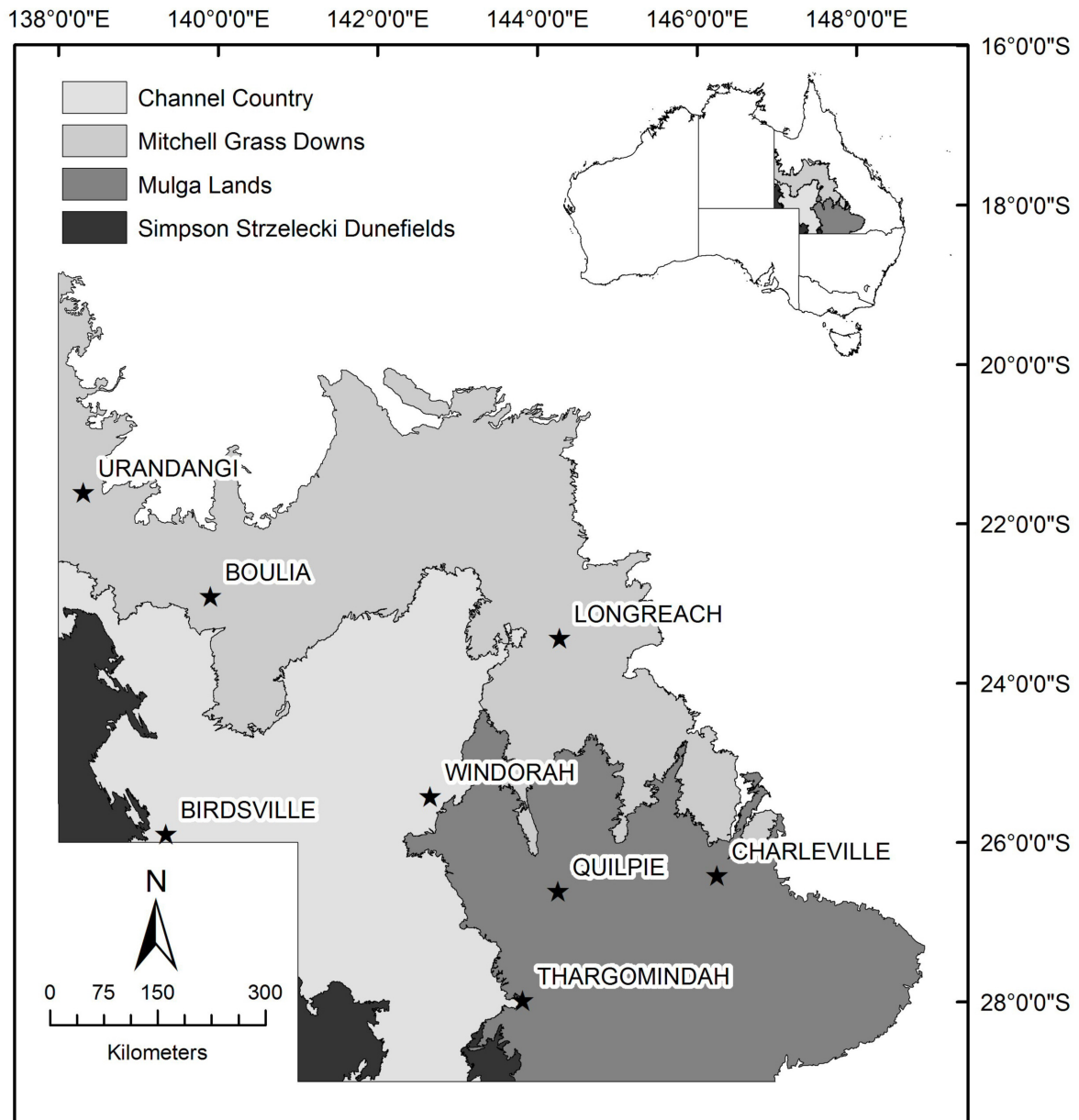


Figure 5.1 Map showing the location of the study area within Australia, the extent of the four bioregions comprising the study area, and the location of meteorological stations used for model validation.

5.3 Model Development

5.3.1 Land Erodibility Controls

Wind erosion occurs under conditions when wind shear forces at the surface exceed the energy required to mobilize soil micropeds (Bagnold, 1941). The threshold friction velocity (u_{*t}) of a land surface describes the wind shear velocity (u_*) at which particle entrainment is

initiated. As u_{*t} decreases the wind shear velocity required to entrain soil particles and fine aggregates will fall. Conceptually, land erodibility (wind erosion hazard) can be considered the inverse of u_{*t} , for as the threshold for particle entrainment decreases susceptibility or erodibility will increase.

Land erodibility is a function of soil erodibility, and the presence of non-erodible roughness elements (rocks, vegetation, landforms) that affect wind erosivity. Chapter 2 described the relationships between these controls, which are illustrated in Figure 5.2.

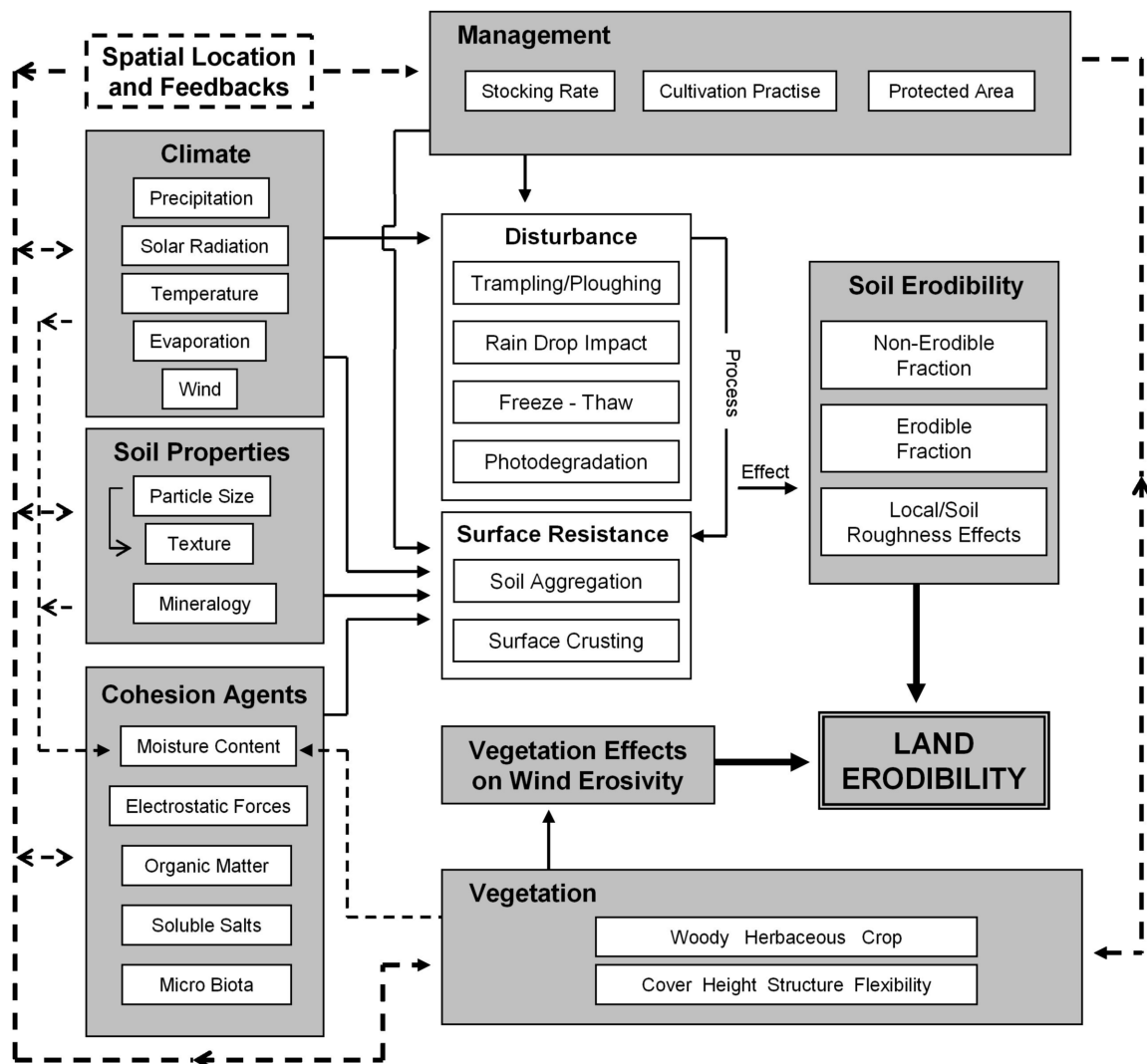


Figure 5.2 Flow chart illustrating the relationships between wind erosion controls within a landscape. Gray boxes represent environmental conditions and processes that determine soil surface conditions and the availability of loose erodible sediment, and the effect of non-erodible roughness elements on the wind shear velocity (wind erosivity)

Soil erodibility is governed by soil texture, but varies with changes in moisture, aggregation and surface crusting (Zobeck, 1991; Merrill et al., 1999). These factors affect the strength of cohesive forces between soil particles and the availability of loose erodible sediments on the soil surface. Together these conditions act against drag and lift forces associated with wind shear. The effect of non-erodible elements like vegetation is through a partitioning of wind shear stress between roughness elements and the soil surface, resulting in an increase in roughness length and potential decrease in wind erosivity (Marshall, 1971; Raupach et al., 1993). Factors controlling land erodibility include those affecting soil erodibility, land type characteristics (vegetation and geomorphology), climate (rainfall, radiation balance, windiness) and management (Figure 5.2). Where non-erodible roughness elements are absent, land erodibility is controlled by soil erodibility.

The term land erodibility implies a relative susceptibility of land areas to wind erosion. Factors controlling soil and land erodibility operate at a range of spatial (local to global) and temporal (seconds to years) scales. Climate variability and land management factors drive changes in soil erodibility and surface roughness conditions within a landscape. This variability means that soil and land erodibility are spatio-temporally dynamic through a continuum (Geeves et al., 2000; Chapter 2, Section 2.1).

5.3.2 Rationale for Model Development

Webb et al. (2006) reported the development of an Australian Land Erodibility Model (AUSLEM) through the specification of a threshold-based rule-set. The rule-set defines conditions under which the Australian landscape may become susceptible to wind erosion. AUSLEM assigns land erodibility rankings through the rule-set, which is applied to inputs of soil texture (sand, silt and clay content), and mean monthly grass cover (%), soil moisture (mm per 10 cm profile) and rainfall (mm). The model is run using monthly-average 5 x 5 km gridded inputs of grass cover, soil water content and rainfall obtained from the Aussie GRASS pasture growth model (Carter et al., 1996a) and Bureau of Meteorology. Soil textural input data were acquired from the Australian Natural Resources Atlas (see DEWHA, 2007). The model did not account for tree or stone cover effects on wind erosion.

Webb et al. (2006) applied AUSLEM to assess land erodibility on a national basis under representative “wet”, “normal” and “dry” conditions. The model demonstrated an ability to

detect erodible regions in arid and semi-arid Australia that are consistent with satellite and observational records of wind erosion activity. However, the model accuracy was affected by its rule-based structure which biased the output toward particular soil texture groups. It did not account for the dynamic nature of soil surface conditions. In addition to this, the model formulation was restrictive such that AUSLEM could not capture the full range of the erodibility continuum (Figure 2.11). The monthly temporal resolution of inputs confounded this issue and suppressed the model sensitivity to short-term (e.g. daily) rainfall events that are an important driver of land erodibility dynamics. A new scheme is therefore required which can capture high temporal resolution changes in land erodibility and through the full range of the land erodibility continuum described in Chapter 2.

Computation of u_{*t} forms the basis of the particle emission schemes in many process based wind erosion models (Hagen, 1991; Marticorena and Bergametti, 1995; Shao et al., 1996; Gregory et al., 2004). As u_{*t} is independent of but determines the wind velocity at which particle entrainment occurs it can be physically related to erosion controls. However, modelling land susceptibility to wind erosion using a physical parameterisation of u_{*t} is computationally intensive and requires continual measurement of variables that are difficult to quantify, e.g. frontal area effects of vegetation and soil particle size distribution (Shao, 2000). The lack of suitable spatial data required as input to the models has also been a limiting factor to their application (Raupach and Lu, 2004).

Land erodibility to wind is a concept and cannot be directly measured in the field. It can, however, be inferred through either u_{*t} , or measurements of wind erosion rates under a variety of conditions for some reference wind velocity. Empirical functions relating erosion rates (i.e. streamwise sediment flux in $\text{gm}^{-1}\text{s}^{-1}$) to surface conditions have been used in the development of field scale wind erosion models with applications in North America and Africa (Woodruff and Siddoway, 1965; Fryrear et al., 1998; Visser et al., 2005; Chapter 3). If derived under standard measurement conditions (i.e. wind tunnel dimensions, wind speeds, etc.) simple empirical expressions can be combined to model the land erodibility continuum. A benefit of this approach is that a model can be developed using functions for which spatial data tend to be readily available as inputs. Further, calibration of the models to improve performance can be achieved without violating assumptions inherent to physical parameterisations of u_{*t} . Caveats of using empirical soil loss functions to define wind erosion rates and land erodibility are that they must exist relative to some reference wind speed at

which they were derived; and application outside the conditions under which they were derived can result in poor modelling outcomes (Visser et al., 2005).

The approach taken to improving AUSLEM as specified by Webb et al. (2006) was to redesign its rule-based system to one using modified empirical expressions. Specific objectives were to produce a model that:

- Captures the physical nature of the land erodibility continuum;
- Does not bias output toward certain soil groups or land forms;
- Retains a computationally simple structure that functions with readily available input spatial data, without the requirement for input development; and
- Can be calibrated to suit soil and vegetation conditions in relevant application areas.

5.3.3 Model Framework

Figure 5.3 shows the revised model framework and computational procedure of AUSLEM. The model has two dynamic components accounting for daily grass cover and soil moisture effects, and three static components that provide tree and stone cover masks and a soil texture factor based on soil clay content. The computational procedure is numbered from 1 to 3. Following parameterisation of the soil erodibility model developed in Chapter 4, that model may also be incorporated into the AUSLEM framework.

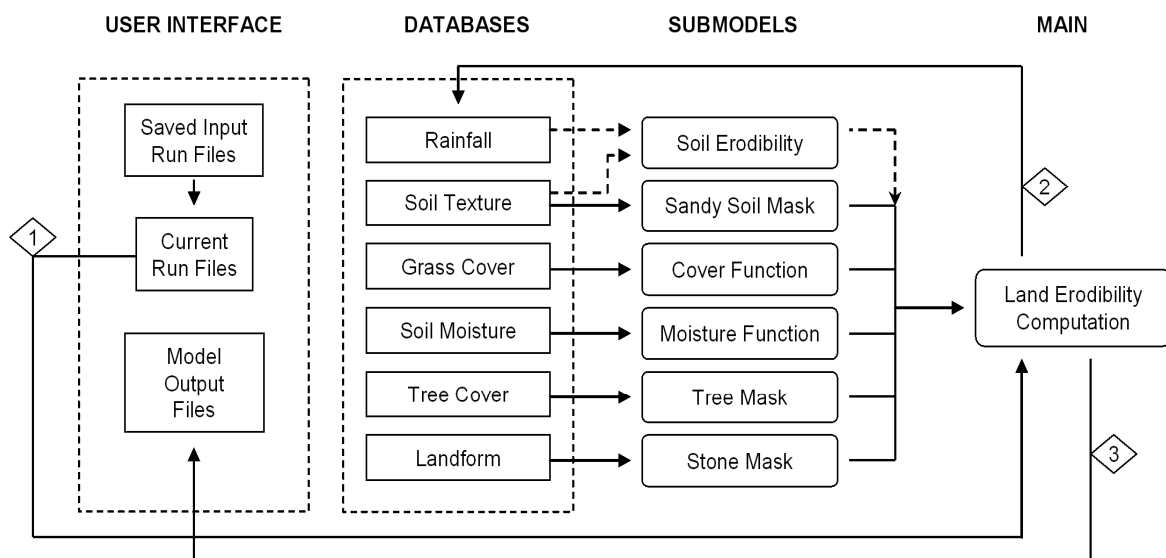


Figure 5.3 Flow chart illustrating the model framework and computational procedure (labelled 1 to 3). A texture based soil erodibility component (dotted arrows) can be included when a suitable model becomes available.

The approach for model development was to select then integrate empirical functions derived under standard measurement conditions (i.e. wind tunnel dimensions, wind speeds, etc.) that capture the relationships between wind erosion controls and the physical nature of the land erodibility continuum. That is, the model was designed to assess the range of susceptibility to wind erosion that a land area may experience based on variability in soil erodibility and surface roughness. Land erodibility was considered to exist on a dimensionless scale as the aim of this research was to produce a model to predict susceptibility to wind erosion, and not to quantify erosion rates or soil loss. The exact values of output erodibility predictions could be of any value so long as the position of a land area in the ranking is correct relative to other areas for its given input conditions. AUSLEM ranks land erodibility on a continuous scale from 0 (not erodible) to 1 (high erodibility). The integration of factors controlling land erodibility (E_r) is through a multiplicative approach of the form:

$$E_r = E_{tx}(tx)E_{gc}(gc)E_w(w)E_{tc}(tc)E_{rk}(rk) \quad (5.1)$$

where $E_{gc}(gc)$ and $E_w(w)$ define the effects of grass cover and soil water content on land erodibility. $E_{tc}(tc)$ and $E_{rk}(rk)$ account for the effects of large roughness elements (tree cover) and surface armouring over dense stony pavements found in the study area (rock cover). $E_{tx}(tx)$ defines the effects of soil texture on land erodibility. The model was composed using the ArcGIS 9.0 (ESRI) model building interface, and is run from the system command line using a batch file listing model parameter inputs.

The revised AUSLEM framework differs significantly to that of the original model. While the original model also had a 5 x 5 km spatial resolution, the revised model can be run on a daily time-step. This means that the model is now sensitive to daily rainfall events and resulting changes in soil moisture. This improved sensitivity is important for assessing landscape responses to climate variability and land management conditions, and opens the pathway for integrating a dynamic soil erodibility scheme into the framework. The removal of the soil textural rule-sets from the model, and the addition of empirical grass cover and soil moisture schemes allow the model to make more realistic assessments of land erodibility dynamics through the continuum. AUSLEM no longer has any bias toward particular soil types. Rather, the model now accounts for the fact that both soil and land erodibility are temporally dynamic conditions. Two parameters added to the revised model, to account for

tree and stone cover effects on wind erosion, provide a more appropriate parameterisation of the erodibility continuum and further improve the model skill in assessing land susceptibility to wind erosion.

Vegetation Effects

Vegetation effects on land erodibility are considered by separating the effects of woody (shrub/tree) and herbaceous (grass) cover. Herbaceous vegetation cover in arid and semi-arid environments tends to increase and decrease at rates faster than shrub- and tree cover in response to climate and management (Specht and Specht, 1999). While regional tree cover can be considered static over short periods, and its effect on wind erosion can be considered by a simple threshold, the effects of temporally dynamic grass cover are better modelled by a continuous function. The effects of grass cover (E_{gc}) on land erodibility are modelled by a negative exponential relationship:

$$E_{gc} = \alpha \exp^{-\beta(\%gc)} \quad (5.2)$$

where α and β are regression coefficients (55.873; -0.0938) denoting the equation intercept and rate of change in erodibility given a change in percentage cover (Chapter 2, Figure 2.7). The relationship was determined by wind tunnel experimentation at wind speed 18 ms^{-1} (Leys, 1991a) and is similar to that determined by Chepil (1944), and the Soil Loss Ratio (SLR) relating soil loss from a soil with cover to that of a bare soil (Fryrear, 1985; Findlater et al., 1990; Chapter 2, Section 2.2.8). While the expression was developed for prostrate (flat lying) wheat stubble, similar exponential relationships were determined by Siddoway et al. (1965) and Lyles and Allison (1981) for standing stubble. By increasing surface roughness, low cover levels of standing vegetation can reduce wind erosion to rates experienced at higher levels of prostrate cover. The mixture of prostrate and standing grass cover in the study region can be accounted for by adjusting the regression coefficients in Equation 5.2 to increase or decrease the sensitivity of land erodibility to grass cover.

Marshall (1972) determined a threshold for wind erosion control based on tree and shrub cover. The threshold exists where tree/shrub spacing is approximately 3.5 times the average maximum height of the vegetation, corresponding to roughly 20% lateral cover in the dryland shrub communities in Australia. Tree cover effects on land erodibility are modelled in

AUSLEM using a mask based on this cover threshold. The tree-cover mask (E_{tc}) assigns an erodibility rank of 0 (not erodible) to land areas with tree cover greater than 20%.

Soil Moisture Effects

Land susceptibility to wind erosion may be significantly reduced by inter-particle cohesive forces associated with soil moisture. The relationship between soil moisture and u_{*t} has been described by a number of authors, but empirical and theoretical models describing the relationship differ between studies due to differences in soil properties, experimental methods and assumptions about the inter-particle contact geometries (Chepil, 1956; Belly, 1964; McKenna-Neuman and Nickling, 1989; Chen et al., 1996; Fécan et al., 1999, Cornelis and Gabriels, 2003; Chapter 2, Section 2.2.5).

An expression was sought that could be applied with the available input spatial data representing soil moisture in the top 10 cm of profile (in mm of water per 10 cm). An analysis of dust-event frequencies was used to derive an appropriate function as wind tunnel experimentation was not available for use in this study. Local dust-event (Meteorological Code 07) frequencies at 16 Australian Government Bureau of Meteorology stations in western Queensland, Australia, were related to event day soil moisture conditions within 25 x 25 km windows representing potential dust source areas (PDSAs) surrounding the stations. The window size was selected based on the likely visible range within which blowing dust could be seen from the stations. Soil moisture data for the PDSAs was extracted from the AussieGRASS pasture growth model (Rickert and McKeon, 1982; Littleboy and McKeon, 1997). A database was then established listing the mean PDSA soil moisture conditions for each dust event. Local dust-events were used as this is the only event type for which potential source areas can be defined with confidence; potential source areas for dust storms (Codes 09, 30-35) and dust hazes (Codes 05, 06) could not be identified using the available observational data (code definitions in Goudie and Middleton, 2006). Local dust-event frequencies were plotted for 10 soil moisture classes defined by an arbitrary even breakdown of the distribution of 170 events, with soil moisture values from 0 to 36 mm per 10 cm, between 1991 and 2006. The data can be represented by a negative exponential relationship of the form:

$$E_w = \exp^{-\beta w} \quad (5.3)$$

where β is a regression coefficient (-0.236) denoting the sensitivity of local dust-event frequencies to soil moisture content. The relationship was found to be consistently strong and statistically significant ($r^2 = 0.94$; $p < 0.0001$) for the stations and across the sandy to clay textured soils in western Queensland. The mean wind speed associated with the events was 8.29 ms^{-1} . The large difference to that associated with the grass cover model (18 ms^{-1}) is due to: 1) the wind speeds used in the wind tunnel analysis by Leys (1991a) being more typical of those associated with dust storm as opposed to local dust events; and 2) local differences exist between wind speeds measured at the meteorological stations and the actual dust source areas. An implication of combining the relationships is that the skill of the model will be very sensitive to the representativeness of Equation 5.2 and may in fact underestimate erodibility in circumstances when wind speeds are $<18 \text{ ms}^{-1}$. The strength and statistical significance of the relationship indicates that the soil moisture data representing 0 – 10 cm of soil profile adequately captures the surface moisture – wind erosion relationship. While the expression relates soil water content to dust-event frequencies rather than wind erosion rates, it is the only expression that could be used with the available model input data. The expression has a form similar to that identified by Shao et al. (1996) and used by Lu and Shao (2001), and can be calibrated within AUSLEM to better represent the soil moisture – land erodibility relationship.

Soil Erodibility Effects

The effects of soil texture on soil erodibility can be modelled by calculating u_{*t} based on the particle size distribution of soils. This approach has been applied in a number of regional and global scale wind erosion models including DPM (Marticorena and Bergametti, 1995), WEAM (Shao et al., 1996), IWEMS (Lu and Shao, 2001) and DEAD (Zender et al., 2003a). While empirical models are available that relate soil texture to wind erosion rates (i.e. Chepil and Woodruff, 1963; Gillette, 1978; Leys et al., 1996), these expressions capture soil erodibility under one of two surface conditions (i.e. crusted or disturbed), and do not account for temporal changes in aggregation and surface crusting that result in variations in soil erodibility (Chapter 4). Temporal changes in soil erodibility have been considered in some field scale models (Hagen, 1991; Fryrear et al., 1998), but are not considered in most others (Gregory and Darwish, 2001; Lu and Shao, 2001; Lasserre et al., 2005). This is because the dynamic and interactive nature of crust and aggregate formation and degradation processes are yet to be quantified in many environments (Bresson, 1995).

Chapter 4 of this thesis established a framework for modelling soil erodibility. Due to an inability to parameterise the framework, this factor is kept spatio-temporally constant in AUSLEM at short (daily to monthly) time scales. This creates a problem in circumstances where vegetation cover and soil moisture are low, and where land erodibility is controlled by soil erodibility, for example on a dry lake bed or playa. If a crust is present on the soil surface and the supply of saltation material is limited, soil erodibility will be considerably lower than if the surface were disturbed (Houser and Nickling, 2001a; Langston and McKenna Neuman, 2005). These effects vary depending on soil texture and crust characteristics. An *a priori* assumption in modelling land erodibility with AUSLEM is that at seasonal to annual time scales regional variations in soil surface conditions in rangeland environments are reflected in grass cover and soil moisture conditions. During wet periods when cover and moisture are high, soil aggregation and crust cover are likely to be higher than in dry seasons and periods of drought (Chapter 4). At these times inter-particle binding by moisture will be low, and crusts and soil aggregates are likely to be most disturbed by photo-degradation and trampling by livestock (McTainsh and Strong, 2007). So, at seasonal to annual time scales regional trends in land erodibility modelled with vegetation cover and soil moisture are likely to reflect actual land erodibility. Interpretation of AUSLEM output should therefore be confined to the examination of trends in output at time-scales longer than one month.

A soil texture mask (E_{tx}) has been included in the current model formulation (Figure 5.3). The mask assigns a scaling factor (provisionally set to 3.0) to areas with soil clay content less than 7.0 %. All other soils are assigned a factor of 1.0. The scaling factor forces sandy soils for which crusting and aggregation are likely to play a minor role in determining erodibility to have a consistently higher susceptibility to mobilisation than soils with high clay content for which crusting and aggregation have greater potential effects on erodibility (Breuninger et al., 1989; Belnap and Gillette, 1997; Leys and Eldridge, 1998).

Stone Cover Effects

The effects of stone cover on land erodibility are considered by the addition of a mask that assigns an erodibility value of 0 (not erodible) to areas with extensive stone cover. The inclusion of the mask was considered important as significant areas (i.e. the Sturt Stony Desert) of the study region are covered by a dense stony pavement. Experimental wind tunnel research by Gillette et al. (1980) demonstrated that this type of dense stone cover is effective in protecting surfaces from deflation. While wind erosion on stony floodplain and playa

surfaces has been observed by the authors, these areas which occur along the river systems dissecting the study area were not included in the mask.

5.4 Model Input Data

Soil texture data (clay content in topsoil) at a 1 x 1 km spatial resolution was acquired from the Australian Natural Resources Atlas (ANRA), as used by Shao et al. (1994). Foliage projective cover (FPC – a vertical projection of tree canopy cover) data for 2003 were used as a representative measure of tree cover. FPC data at a 30 x 30 m spatial resolution were sourced from the Statewide Landcover and Trees Study (SLATS), and were derived from Landsat ETM+ satellite imagery by the methods of Danaher et al. (2004). Aerial photograph assessments (1:250 000 scale) of the extent of stony pavements within the study area were used to derive the stone cover mask and were sourced from the Western Arid Regions Land Use Study (WARLUS; QDPI, 1974). Daily spatial data of grass cover (%) and soil water (mm per top 10 cm of profile) were obtained from the Australian pasture growth model AussieGRASS (Carter et al., 1996a,b; Littleboy and McKeon, 1997). AussieGRASS is run at a 5 x 5 km spatial resolution on a daily time-step and can be used to hind-cast pasture conditions from 1890 to present. All data except FPC were rescaled to a 5 x 5 km resolution for modelling with AUSLEM. The FPC data were used at the 30 x 30 m resolution as rescaling resulted in a significant loss of tree cover detail (due to sub-grid scale averaging of the sparse tree cover in the study region), and poor model performance in regions where tree cover is a significant control on wind erosion.

Calibration and validation of AussieGRASS was carried out in Queensland pasture communities, so the model can be expected to perform best in the current study area. However, model performance may vary between pasture communities due to plant physiological and growth characteristics. AussieGRASS output is most accurate over areas 25 to 50 times the resolution of its 5 x 5 km inputs (Carter et al., 1996b). This has implications for AUSLEM output validation and interpretation, which must also be conducted at these scales.

5.5 Model Application and Validation

Land erodibility cannot be directly measured in the field, so proxy measures that indicate wind erosion hazard must be used for validation. While wind erosion rates are relatively easily measured at small scales (i.e. individual fields), and have been used to validate models such as RWEQ, WEPS, TEAM and IWEMS (Hagen, 1991; Fryrear et al., 1998; Lu and Shao, 2001), finding indicators of land erodibility at the regional scale at which AUSLEM operates is problematic. The objective of the validation was to compare time series trajectories of mean annual AUSLEM output with trends in dust-event frequencies and wind speeds at locations across the study area. The validation hypothesis was that at annual time scales dust-event frequencies are highly dependent on regional land erodibility and suitable wind speeds, so temporal trends in time-series of these should display similar patterns. Cross-correlation analysis of dust-event frequencies, mean wind speeds and modelled land erodibility was identified as a means for quantifying the similarity of their temporal trends and therefore model performance.

5.5.1 Methodology

Dust-event records (Observation codes 05-09 and 30-35) and mean annual 3 pm local wind speeds (ms^{-1}) were extracted from the Bureau of Meteorology (BoM) database for 16 stations within the study area for the period January 1980 to December 1990. This period was chosen to avoid overlap with the dust-event data used to derive the model soil moisture function (1991-2006). Mean annual wind speeds were selected to provide an indicator of windiness at the same temporal resolution as the model output. Annual total event frequencies were computed for each station, and these were linked in a database to the mean wind speed data. Eight of the 16 stations had recorded >1 event in every year, and these were used for comparison with AUSLEM output predictions (Figure 5.1). Four classes of dust-event observations were created to assess the effects of using different event types in the validation. The classes included: 1) all dust-event types; 2) events with hazes (Codes 05, 06) removed; 3) events with hazes and whirls (+ Code 08) removed; and 4) the Dust Storm Index (DSI), a composite index of local, moderate and severe dust-event frequencies (McTainsh and Tews, 1998).

AUSLEM was run to predict land erodibility on a daily time-step for the period 1980 to 1990. Output was then aggregated into monthly and annual means to match the temporal resolution of dust-event frequencies. A caveat of using observational dust-event data is the inherent difficulty in identifying source areas of non-local dust events (i.e. dust hazes, and dust storms). AUSLEM output was compared with the station data at multiple spatial scales to determine if an optimal scale exists for this validation approach. Firstly, circular source areas of interest (AOIs) were delineated around each of the eight stations to act as potential dust source regions with radii of 25 km, 50 km, 100 km, and 150 km. The AOI size range was determined based on the ranges of maximum visibility around the stations within which dust events could be observed, and the requirement to test AUSLEM performance across multiple scales. A second half-circle AOI from (25 to 150 km radii) was then used, with AOIs aligned to the north, south, east and west around the stations. The purpose of this second directional test was to determine whether a better comparison between model output and dust-event frequencies could be made if only dominant dust sources upwind of the stations were considered. Summary statistics (mean, min, max, standard deviation, mode) of mean annual AUSLEM output were computed for each AOI at each station. Cross-correlation analysis was then used to compare temporal trends in model output statistics with the measures of dust-event frequencies.

5.5.2 Results - Annual Land Erodibility Predictions

Figure 5.4 presents mean annual land erodibility predictions for western Queensland from 1980 to 1990. Results show that regions experiencing high susceptibility to wind erosion occur in the south-western corners of the state in the Mulga Lands, Simpson Desert Dunefields, and western Channel Country. The Mitchell Grass Downs in the north-east of the study region tend to have consistently low land erodibility. The regions of high land erodibility in the south-west lie in the most arid section of the study area (< 250 mm rainfall per annum), while the low erodibility lands to the north-east are consistent with that area receiving higher mean annual rainfall (~450 mm). The distribution of modelled land erodibility is in agreement with the general spatial pattern of wind erosion activity across western Queensland (Middleton, 1984; Burgess et al., 1989; McTainsh et al., 1990).

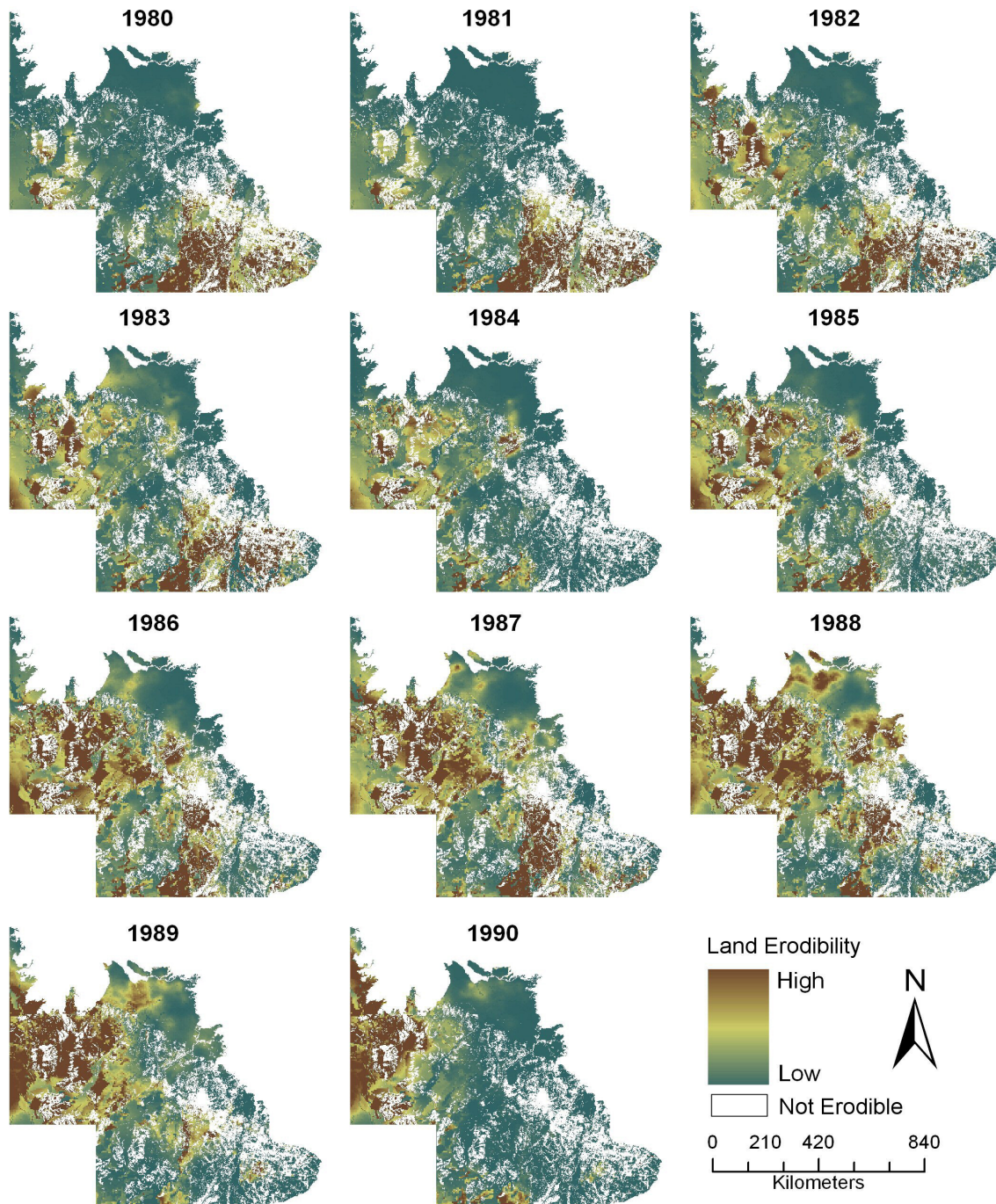


Figure 5.4 Mean annual land erodibility predictions from AUSLEM for the period 1980-1990. White areas are not erodible due to tree and stone cover being above the model thresholds.

5.5.3 Results - Station Comparisons

Annual dust-event frequencies and DSI values for the eight BoM stations are listed in Table 5.1. The spatial distribution of stations allows for analysis of model performance in each of the four study area bioregions (Figure 5.1).

Table 5.1 Dust-event frequencies at stations used for model validation. Dust event classes listed for each station include dust-event frequencies for all event types (All); events with hazes removed (NoHz); events with hazes and dust whirls removed (NoHzWr); and Dust Storm Index (DSI) values

Station*	Class	1980	1981	1982	1983	1984	1985	1986	1987	1988	1989	1990
Birds	All	9	2	33	51	22	35	30	47	64	74	42
	NoHz	7	1	32	49	21	32	25	44	64	69	33
	NoHzWr	5	1	32	46	21	29	24	42	62	66	33
	DSI	0.3	0.05	13	11.7	6.85	1.35	1.1	15.1	2	13.45	8.05
Boul	All	4	0	10	12	3	5	12	20	11	20	16
	NoHz	3	0	8	3	0	2	10	10	7	14	10
	NoHzWr	3	0	8	3	0	2	10	10	7	14	10
	DSI	5.05	0	1.3	1.1	0	0.1	7.2	8.1	5.3	5.25	2.15
Char	All	436	369	271	93	38	36	28	9	18	11	38
	NoHz	118	94	65	35	14	20	20	2	5	9	20
	NoHzWr	8	11	8	6	5	7	13	0	1	5	5
	DSI	4.75	8.4	3.95	4	2.55	0.9	5.5	0.1	1.2	0.35	0.85
Thargo	All	11	3	5	6	1	3	11	14	6	14	8
	NoHz	10	1	2	2	1	3	8	14	6	13	2
	NoHzWr	10	1	1	2	1	3	8	13	6	13	2
	DSI	11.3	1	0.05	1.05	0.05	5.05	3.15	3.2	1.2	1.4	0.1
Uran	All	9	3	8	4	0	17	11	17	25	70	63
	NoHz	9	3	6	4	0	16	11	14	24	69	63
	NoHzWr	9	3	5	3	0	16	10	12	23	67	34
	DSI	0.45	0.15	6.2	1.1	0	3.45	4.25	7.45	2	13.75	13
Wind	All	35	10	20	10	15	26	44	55	75	15	12
	NoHz	28	3	12	7	15	26	44	54	75	15	12
	NoHzWr	23	3	12	6	12	20	27	42	69	15	11
	DSI	1.3	0.15	0.55	0.3	0.65	2.2	4.85	8.45	14.35	12.4	0.4
Long	All	5	2	130	63	27	8	5	8	6	8	56
	NoHz	0	2	14	16	13	8	4	1	5	0	14
	NoHzWr	0	2	3	3	0	7	3	0	3	0	11
	DSI	0	0.05	1.45	0.6	0.5	0.2	1.1	0	1.15	0	0.45
Quil	All	6	5	7	0	2	3	3	5	5	3	7
	NoHz	6	5	5	0	2	2	3	4	5	2	5
	NoHzWr	6	5	5	0	2	2	3	4	5	2	10
	DSI	1.2	6.1	0.2	0	0.1	0.1	0.1	1.15	2.15	5.05	3.05

* Birds (Birdsville); Boul (Boulia); Char (Charleville); Thargo (Thargomindah); Uran (Urandangie); Wind (Windorah); Long (Longreach); Quil (Quilpie).

The range of land surface characteristics (soil, vegetation types), management and climate variability between the stations means that they have distinct dust-event frequency time-series trajectories (Table 5.1). The data also show that the types of dust events comprising the frequencies varied between stations. The western stations (Birdsville, Urandangie, Boulia, Windorah) show little difference in frequencies between event classes. On the other hand, the eastern stations (Longreach, Charleville) have larger differences between class frequencies.

Removing dust hazes and whirls from the western stations records has little effect on the frequencies, indicating that the majority of events recorded at these stations are dust storms (i.e. Synop Codes 09, 30-35) or locally blowing dust (Code 07). At the eastern stations removing hazes and whirls significantly reduces the remaining event frequencies, indicating that the majority of events recorded at these stations are non-local (i.e. hazes) or are not representative of lateral wind erosion activity (i.e. dust whirls).

Figure 5.5 presents examples of time-series trajectories of mean annual land erodibility for Quilpie, Thargomindah and Windorah. Figure 5.5a presents trajectories for the circular AOIs at scales from 25 to 150 km. Figure 5.5b presents trajectories extracted from the directional half-circle AOIs with a radius of 100 km. The patterns in trajectory changes between scales in Figure 5.5a are typical of those found across all stations. Differences were found between the magnitude of output values between scales, but no consistent differences were found between the trends in the trajectories. Data for the year 1980 at the 150 km scale for Quilpie and Thargomindah (5.5a), for example, indicate that land further from these stations had low erodibility during that year. Variations in mean annual output values for the directional half-circle AOIs also exhibit significant differences in magnitude and not trajectory. The trends in mean annual land erodibility extracted from model output on one side of a station (i.e. to the west) were found to be similar to those extracted from other areas around the stations (i.e. to the north, east or south). This outcome is interesting as the BoM stations are surrounded by multiple soil and vegetation types so the land erodibility trajectories could be expected to have different trends.

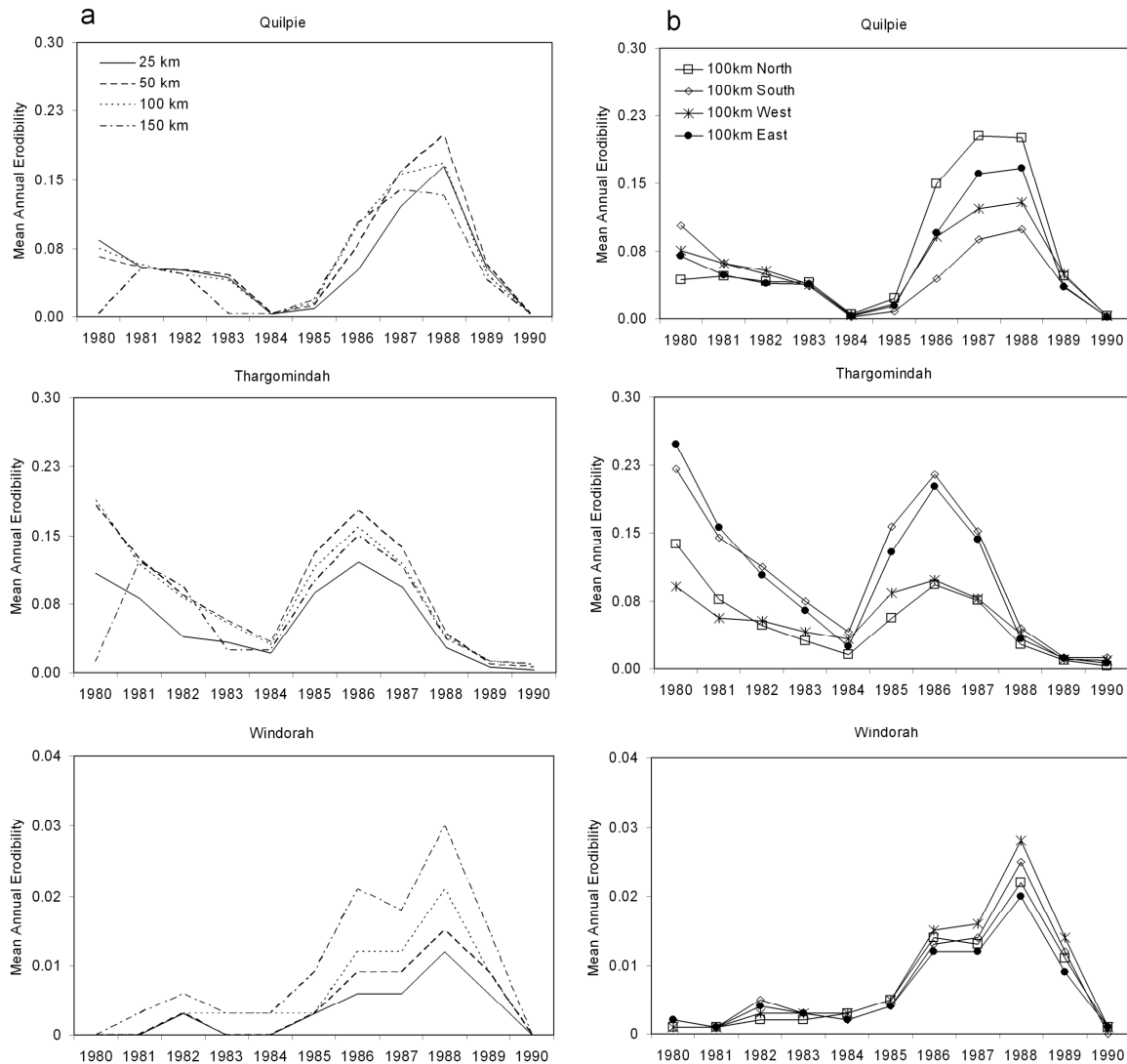


Figure 5.5 Examples of time series trajectories of mean annual AUSLEM output for three stations (Quilpie, Thargomindah and Windorah). 5a (left hand column) presents trajectories for circular potential dust source areas with radii from 25 to 150 km. 5b (right hand column) presents trajectories for potential dust source areas (radius 100 km) oriented to the north, south, east and west around the stations

Table 5.2 presents results of the cross-correlation analysis between dust-event frequencies and mean and maximum AUSLEM output at multiple spatial scales for the circular AOIs. Like the dust-event frequencies, modelled land erodibility at each station has a distinct time series trajectory. Significant improvements in correlations with the directional AOI data were not found due to the similarity of the time series trends to those from the circular AOIs. The comparison of model output with the directional AOI data therefore did not provide further information on how the model was performing so the results are not shown here. AUSLEM

output trajectories show strong correlations ($r^2 > 0.67$, $p < 0.05$) with annual dust-event frequency groups at four stations: Boulia, Charleville, Urandangie, and Windorah.

Table 5.2 Correlation coefficients (r^2) for the cross-correlation analysis between mean and maximum (max) AUSLEM output at multiple spatial scales (25 to 150 km) and dust-event frequencies for eight stations within the study area. Correlations between AUSLEM output and dust-event frequencies that are statistically significant ($p > 0.05$) are **boldfaced**

Station	Class	25 km		50km		100km		150km	
		Mean	Max	Mean	Max	Mean	Max	Mean	Max
Birdsville	All	0.22	0.30	0.36	0.55	0.48	0.58	0.49	0.66
	NoHz	0.20	0.28	0.33	0.52	0.46	0.55	0.47	0.66
	NoHzWr	0.17	0.26	0.31	0.51	0.44	0.53	0.46	0.65
	DSI	-0.36	-0.27	-0.25	-0.14	-0.15	-0.07	-0.13	-0.06
Boulia	All	0.82	0.83	0.85	0.81	0.86	0.72	0.79	0.58
	NoHz	0.77	0.71	0.82	0.81	0.84	0.75	0.85	0.69
	NoHzWr	0.77	0.71	0.82	0.81	0.84	0.75	0.85	0.69
	DSI	0.28	0.27	0.36	0.29	0.45	0.28	0.70	0.63
Charleville	All	0.92	0.87	0.88	0.41	0.71	0.26	0.45	0.19
	NoHz	0.94	0.86	0.87	0.34	0.67	0.19	0.37	0.08
	NoHzWr	0.44	0.51	0.45	0.15	0.36	-0.07	0.26	-0.03
	DSI	0.64	0.78	0.66	0.37	0.56	0.19	0.35	0.09
Thargomindah	All	-0.18	0.02	0.19	-0.19	0.19	0.25	-0.02	-0.15
	NoHz	-0.25	0.06	0.26	-0.07	0.26	0.29	0.03	-0.11
	NoHzWr	-0.31	0.03	0.26	-0.10	0.26	0.28	0.00	-0.15
	DSI	-0.25	0.52	0.71	0.54	0.76	0.79	-0.03	-0.09
Urandangie	All	0.93	0.92	0.92	0.90	0.92	0.86	0.87	0.73
	NoHz	0.92	0.92	0.92	0.89	0.91	0.85	0.86	0.72
	NoHzWr	0.80	0.84	0.79	0.80	0.81	0.79	0.77	0.68
	DSI	0.85	0.86	0.85	0.86	0.87	0.86	0.79	0.63
Windorah	All	0.82	0.84	0.81	0.85	0.85	0.77	0.83	0.72
	NoHz	0.84	0.87	0.84	0.88	0.88	0.76	0.86	0.71
	NoHzWr	0.87	0.89	0.84	0.89	0.88	0.73	0.84	0.71
	DSI	0.91	0.94	0.93	0.92	0.88	0.61	0.87	0.64
Longreach	All	-0.35	-0.39	-0.34	-0.42	-0.35	-0.38	-0.34	-0.26
	NoHz	-0.35	-0.33	-0.33	-0.33	-0.32	-0.15	-0.23	0.11
	NoHzWr	-0.17	-0.10	-0.16	-0.09	-0.14	-0.05	-0.13	-0.01
	DSI	0.32	0.33	0.32	0.31	0.30	0.29	0.35	0.31
Quilpie	All	0.24	0.30	0.18	0.19	0.18	0.29	0.17	0.21
	NoHz	0.41	0.44	0.32	0.37	0.34	0.48	0.26	0.26
	NoHzWr	0.08	0.11	0.03	0.03	0.04	0.07	0.02	-0.02
	DSI	0.10	0.04	0.05	0.00	0.00	-0.08	0.05	0.01

Correlations at Boulia and Windorah strengthen at the 100 km scale, while correlations at Charleville and Urandangie strengthen towards the 25 km scale. At Birdsville significant strong correlations were only found at the 150 km scale (max values). At Thargomindah correlations were weak, but strengthened toward the 100 km scale with DSI. Correlations at Quilpie are weak for all data combinations, but show evidence of strengthening toward the 25 km scale. No match between modelled land erodibility and observed dust-event frequencies was found at Longreach.

Correlations between model output and DSI varied significantly between stations (Table 5.2). The DSI trajectory matched AUSLEM output trends at Thargomindah while dust-event frequencies did not. At Windorah, correlations with DSI strengthened toward the 25 km scale while at Urandangie they strengthened at the 100 km scale. At Boulia dust-event frequency trends correlated with AUSLEM output, but DSI did not. No strong positive correlation was found between model output and DSI at Birdsville, Quilpie or Longreach.

Table 5.3 shows that consistently strong correlations between dust-event frequencies and mean annual 3 pm wind speeds were only found at Urandangie. As the correlations between dust-event frequencies and modelled land erodibility were also strong (Table 5.2), this indicates that during the period 1980 to 1990 wind erosion activity around Urandangie had a mutual dependence on concurrently elevated windiness and land erodibility. Results in Table 5.3 suggest that for the remaining stations annual variations in land erodibility are a stronger forcing mechanism behind variations in dust-event frequencies, which is supported by the stronger correlations with model output (Table 5.2).

Table 5.3 Correlation coefficients (r^2) for the cross-correlation analysis between mean 3 pm wind speeds (ms^{-1}) and dust-event frequency groups and DSI for meteorological stations within the study area. Correlations between mean 3 pm wind speeds and dust-event frequencies that are statistically significant ($p < 0.05$) are **boldfaced**

Station	Dust Event Class			
	All	NoHz	NoHzWr	DSI
Birdsville	0.15	0.13	0.12	0.20
Boulia	0.03	0.10	0.10	-0.20
Charleville	-0.33	-0.40	-0.25	-0.16
Thargomindah	-0.25	-0.23	-0.24	0.53
Urandangie	0.79	0.81	0.78	0.60
Windorah	0.51	0.47	0.37	-0.10
Longreach	-0.32	0.03	0.38	-0.04
Quilpie	-0.05	-0.07	-0.16	0.17

Figure 5.6 presents time series trajectories of mean AUSLEM output for 100 km circular AOIs and annual total dust-event frequencies (all events group).

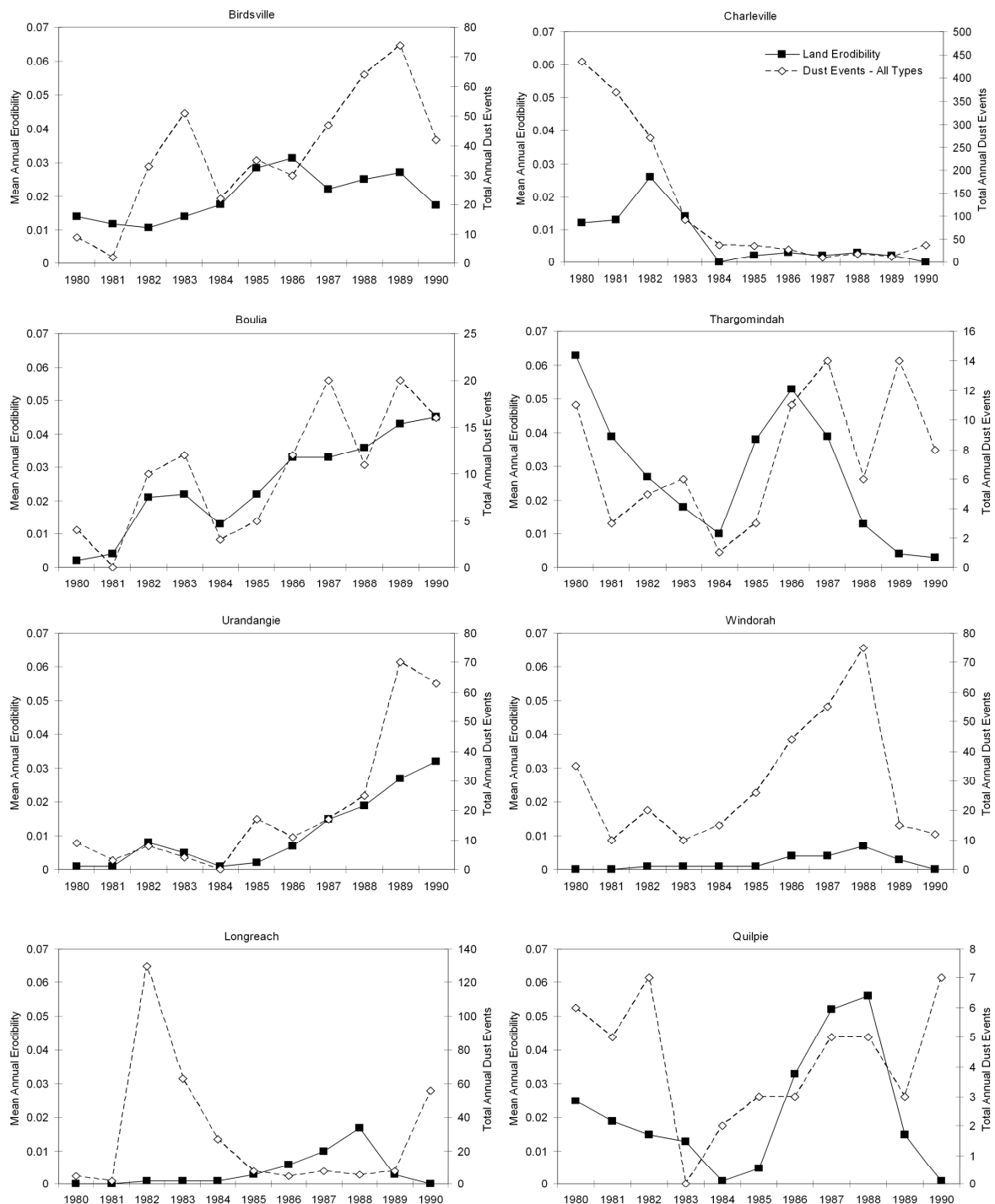


Figure 5.6 Time series trajectories of mean annual AUSLEM output for 100 km windows and annual total dust-event frequencies (all event types) for the eight meteorological stations used in model validation. Solid lines represent AUSLEM output trajectories and dashed lines represent dust-event frequencies

A pattern was found in the years in which model output trends went against those of dust-event frequencies. For example, an increase in dust events was linked to a decrease in land erodibility. This occurred at three and then four stations in 1981 and 1982, at three stations in both 1986 and 1987, and at five stations in 1990. The poor synchronisation of trends in these years was found to occur across each of the spatial scales from 25 to 150 km, and resulted in the weak correlations at Quilpie, Thargomindah and Birdsville (Table 5.2). The synchronisation of trajectories in other years was found to be consistently good, except at Longreach where the trends did not match in any years.

5.6 Discussion

5.6.1 Model Performance

Cross-correlation results (Table 5.2) show that AUSLEM performs well at half of the stations used for validation. The correlations between model output and dust-event frequencies were found to vary with scale at all of the stations except Longreach, at which there was no match between modelled land erodibility and dust-event frequencies. The differences in correlations with scale indicate that either: 1) the dust source areas responsible for the observed dust events are located within different areas and at different distances from the stations; 2) that AUSLEM performs best at a particular scale depending on the land type characteristics around a station, or; 3) a combination of these factors is responsible.

Figure 5.5a indicates that model output predictions can display similar temporal trends across spatial scales. Consistent model performance across scales (i.e. Urandangie) indicates that grass cover and soil moisture conditions may be uniform around the station, and/or that the dust source areas were widespread. Variations in the trajectories at particular scales are indicative of changes in land erodibility at a certain distance from a station. An increase in mean output values at larger scales, for example, suggests an increase in land erodibility further away from a station. By comparison, lower land erodibility values such as around Windorah can be attributed to high vegetation cover on the Coopers' Creek channels close to the town. The strengthening of correlations between model output and dust-event frequencies at Charleville and Quilpie can be explained by local tree clearing (Charleville) and the

presence of barren floodplains (Quilpie) which provide dust source areas close to these stations.

Results indicate that at annual time scales land erodibility can be effectively modelled without incorporating a specific soil erodibility sub-model. However, AUSLEM did not perform well at small scales (i.e. within areas 25 x 25 km) when land erodibility was driven by soil erodibility. The similarity of land erodibility trajectories for different areas around the stations (Figure 5.5b) suggests that AUSLEM shouldn't be used to differentiate specific variations at the land type scale ($\sim 50 \text{ km}^2$). Differences in land erodibility detected by AUSLEM are due to spatial and temporal variations in grass cover and soil moisture. If one land type is depleted of vegetation and soil moisture it will have a higher indicated erodibility than an adjacent land type. This would lead to the differences in magnitude of erodibility values at different AOI orientations (5.5b). Once a land type becomes bare, and land erodibility is determined by soil erodibility, AUSLEM cannot detect specific temporal changes in susceptibility to wind erosion. Hence, the land erodibility trajectories were similar for various AOI positions around the stations. The result demonstrates that the assumption that vegetation cover and soil moisture conditions are adequate predictors of land erodibility (Section 5.3.3) is scale dependent, and is only applicable when assessing landscape (10^3 km^2) to regional ($>10^4 \text{ km}^2$) scale patterns in land erodibility. At smaller spatial scales the importance of soil erodibility in driving temporal variations in land erodibility will increase and the performance of models that do not account for this will suffer. This outcome clearly demonstrates the requirement for research to parameterise soil erodibility models like that developed in Chapter 4.

Model output and dust-event frequency trajectories were consistently found to be dissimilar in 1981-82, 1986-87 and in 1990. These were years in which rainfall was anomalously low or high at areas within the study region relative to the preceding year. The poor match of trajectories reflects either poor model performance, or differences in land erodibility driven by these rainfall conditions, with areas of high erodibility outside the station AOIs affecting dust-event frequencies recorded at the stations. This is shown for 1982 when a short intense drought affected the study region. In that year dust-event frequencies increased at Quilpie, Thargomindah, Birdsville and Longreach while land erodibility did not as residual grass cover around the stations was still high from the previous year. Table 5.1 indicates that during the periods when the trajectories were not synchronised there were significant

differences in the types of dust-events being recorded at the stations. Increases in land erodibility outside the station AOIs was manifested through an increase in observed dust hazes, rather than dust storms or locally blowing dust. This should be reflected in improved correlations of model output with the dust-event groups where dust hazes and whirls have been removed. However, the coarse temporal resolution of the comparison led to short periods of high erodibility and potential dust production being “lost” in the annual averaging of model output. This further contributed to the poor synchronisation of trajectories with the few remaining events in these years (potentially with local source areas).

The strongest correlations between model output and dust-event frequencies were found at stations where events were dominated by local wind erosion and dust storm activity (Boulia, Urandangie and Windorah). The poor synchronisation/correlation at Longreach can be attributed to the fact that the dust events recorded at that station were predominantly hazes for the entire analysis period (Table 5.1). These events most likely had non-local origins, and so do not provide a good measure of local wind erosion activity by which model performance can be measured. Interpreting a poor correlation of model output with dust-event frequencies as indicating poor model performance is therefore highly subjective and dependant on the dust events representing wind erosion activity within the potential dust source AOIs. A further issue with the dust-event comparison is indicated by the significant differences between the dust-event frequency and DSI trajectories and the differences in correlations between these and model output. While all observed dust events were used to compile the frequency data, DSI is computed using the severest (lowest visibility) dust event recorded on a particular day. A smoothing of event frequencies by the index would have contributed to the differences in correlations. Attempting to assess model performance at shorter time scales (i.e. weeks to months) using dust-event frequencies would create further problems in the analysis as the dependence of wind erosion activity on windiness (relative to land erodibility) increases and because wind erosion activity can be low when land erodibility is high if $u_* < u_{*t}$.

5.6.2 Model Limitations

A number of limitations to the research were found to affect model performance. These limitations relate to: the absence of a quantitative model to predict the temporal evolution of soil surface conditions driving soil erodibility; the coarse spatial resolution of the model input

data decreasing land erodibility in heterogeneous landscapes where small erodible features (i.e. dune crests in the Simpson Desert region) are “lost” by sub-grid scale averaging; and at short time scales model output must be interpreted relative to a reference wind velocity (18 ms^{-1} - determined by the grass cover function), otherwise erodibility may be over- or under-predicted by the model.

Using dust-event data for model validation following its use in model development did not allow for a completely independent test of model performance. The dust-event data was, however, the only data available for testing AUSLEM performance at comparable spatial and temporal scales. The validation period (1980 – 1990) was selected to avoid using the same event data for model development and validation. Comparisons of model output with assessments of erosion hazard on long-range ($> 100 \text{ km}$) transects have potential to be useful for validation (as per Hassett et al., 2000; described in Chapter 6). However, efforts by the authors to develop such methods have been restricted by scaling issues (relating field based observations to coarse resolution model output) and the limited availability of higher spatial resolution model input data required for the comparison.

Where possible, these limitations should be addressed in future research. As detailed in Chapter 4, attention must be directed toward developing models to simulate temporal changes in soil erodibility. A comparison of model output with field assessments of wind erosion hazard and predictions from alternate models like the Integrated Wind Erosion Modelling System (Shao et al., 1996, Lu and Shao, 1999) should also be conducted to independently test model performance.

5.7 Conclusions

This chapter has reported on the development of a model to predict land susceptibility to wind erosion in western Queensland, Australia. A rationale for model development has been presented, and the model has been validated by a comparison of annual output time series with dust-event frequencies and DSI over an 11 year period. The model output had strong correlations with dust-event frequencies at half of the validation stations. Poor correlations at the other stations were linked to limitations of the modelling scheme, model input data characteristics and problems with testing model performance using dust-event data which in

itself is based on many generalisations at a range of spatial scales. The model agreement with dust-event frequencies was found to vary across spatial scales and was highly dependent on land type variability around the reference stations, and on the types of dust events used in the validation process. Validation of AUSLEM with field assessments of land erodibility, presented in Chapter 6, seeks to provide a further independent test of the model performance.

Developing a land erodibility modelling capability with AUSLEM has potential to significantly contribute to our knowledge of wind erosion processes in Australia. The resolution of its spatial data inputs (5 x 5 km resolution, daily time-step) and their availability (1890 to 3-month forecasts) set AUSLEM apart from other wind erosion models in terms of its application potential. Further research, presented in Chapter 7, will provide an analysis of spatio-temporal patterns in modelled land erodibility and relate these to landscape characteristics, climate variability and management practices within the study area. This will provide a significant advancement over evaluations of static wind erosion hazard maps, and allow for quantitative assessments of potential land degradation by wind erosion to be made at scales appropriate for land management.

Chapter 6

Assessing Land Susceptibility to Wind Erosion: Validation of the Australian Land Erodibility Model

This chapter addresses Objectives 5 and 6. The chapter describes the development of a method for visually assessing land erodibility at the landscape scale that can be used to monitor spatial and temporal changes in landscape condition. The data are then used to validate the land erodibility model through a point (observation) to pixel (model output) comparison.

6.1 Introduction

Wind erosion models have been developed to assess dust emission and transport processes at a range of spatial ($\sim 1 \text{ km}^2$ - global) and temporal (minutes - years) scales. Calibration and validation of the models is essential if the output is to be used with confidence in research and land management applications (Janssen and Heuberger, 1995). This has created a requirement for procedures to test and validate model performance across a variety of application areas, including cultivated and rangeland environments. Historically, validation procedures have relied on comparisons of model simulations with measurements of wind erosion rates (soil loss per unit time), or of simulated and recorded dust concentration time series data from observation networks (Zobeck et al., 2003). Wind erosion is a dynamic process that displays considerable spatial and temporal variability (Gillette, 1999). Using isolated point samples to validate spatial models is therefore problematic, as significant variations in emissions and erodibility may occur at small scales (Okin, 2005). Accordingly, there is a requirement to develop methods for monitoring indicators of wind erosion at the landscape scale (i.e. $>10^3 \text{ km}^2$), and to integrate this data into the calibration and validation of spatially explicit wind erosion models.

This chapter reports on research to: 1) develop a field monitoring approach for visually assessing land susceptibility to wind erosion at the landscape scale; and 2) apply the data in validating predictions from the Australian Land Erodibility Model - AUSLEM (Webb et al., 2009; Chapter 5). The chapter defines criteria suitable for evaluating land susceptibility to wind erosion and describes application of the criteria to visually assess land erodibility over long distances (10^3 km) using vehicle-based transects through the western Queensland study area (Figure 6.1). Finally, data from the transect studies are used for model validation.

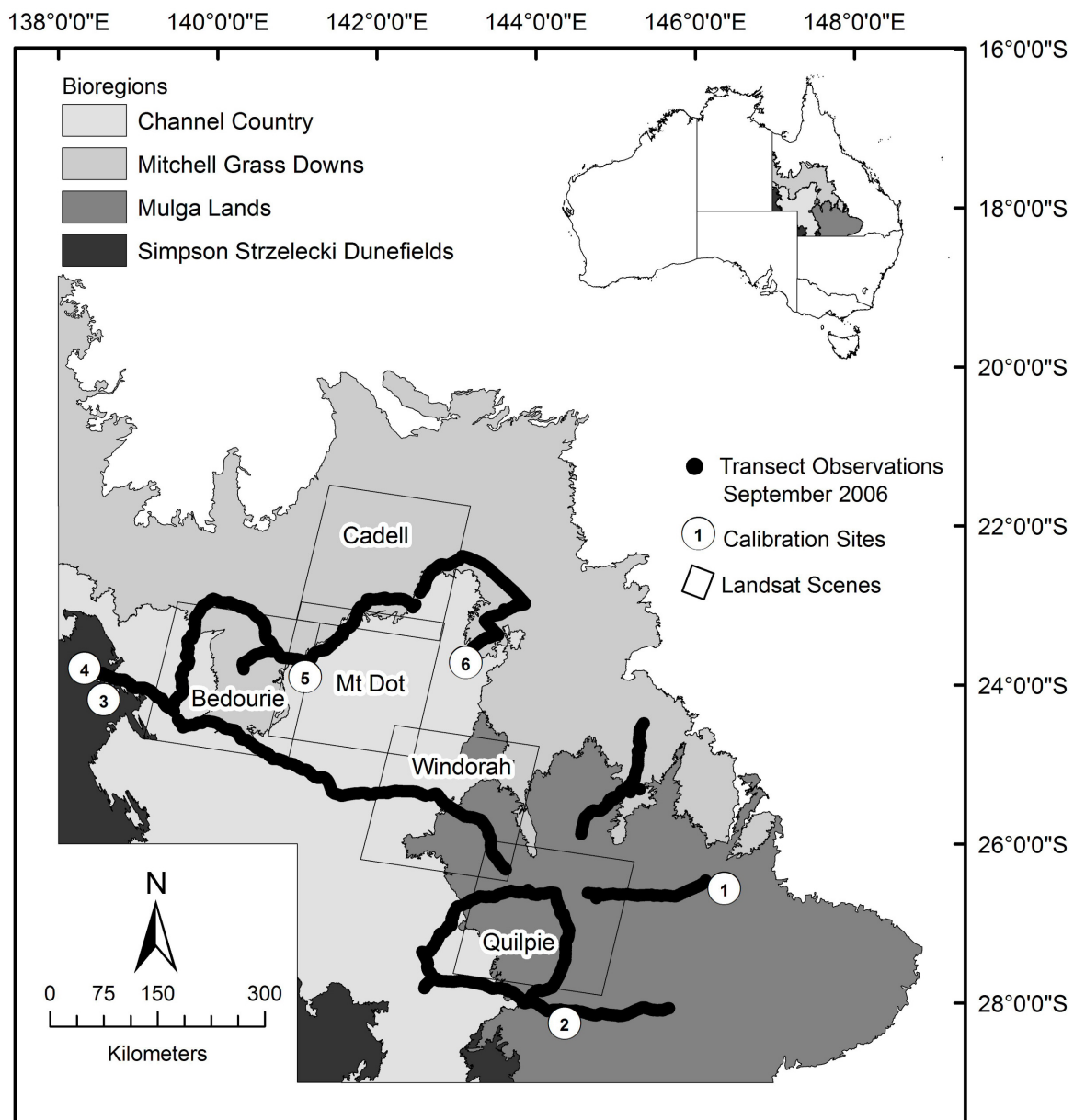


Figure 6.1 Location map showing major bioregions, Landsat ETM+ image scenes used for model validation, transect observation tracks for data collected in September 2006, and vegetation cover calibration sites: 1) 'Croxdale', 2) 'Lake Bindegolly', 3) 'Ethabuka' (sand dune crest), 4) 'Ethabuka' (dune swale), 5) 'Diamantina National Park', 6) 'Spoilbank'

6.2 Field Monitoring Approach

6.2.1 Land Erodibility Assessment Criteria

A rule-set was established (Table 6.1) to provide criteria for the visual assessment of land susceptibility to wind erosion, i.e. land erodibility. The criteria were employed to minimise error in the subjective nature of visually assessing land erodibility. Criteria used to assess the susceptibility of the soil surface to wind erosion (soil erodibility) were adapted from the Wind Erodibility Groups (Skidmore et al., 1994; Table 2.2). Soils were assigned an erodibility ranking based on their textural properties. Temporal variations in soil erodibility were assessed by disturbance levels (due to livestock trampling) and adjusting the ranking accordingly. For example, a crusted loam soil would be considered non-erodible, whereas the same soil in a heavily disturbed state would be assigned a high erodibility ranking.

Table 6.1 Criteria used for the visual assessment of land susceptibility to wind erosion.

Erodibility Ranking				
Indicator	High (3)	Moderate (2)	Low (1)	Not Erodible (0)
Soil Texture	Sandy	Loamy	Clay	-
Surface Condition	Heavy trampling disturbance	Moderate trampling disturbance	Light trampling disturbance	Crusted
Grass Cover	0 – 20 %	21 – 30 %	31 – 50 %	> 50 %
Tree Cover	Erodible		Not Erodible	
	0 – 20 %		> 20 %	
	Sparse / None Present		Dense (> 70 % cover)	
	No rain during survey or prior week		Rain during survey	
Land Type	Descriptor			
	Woodland, Open Downs/Plains, River Channels, Claypan (playa), Sandplain, Dunes, Gibber (stony mantle)			

Criteria for assessing the influence of grass cover were adapted from the rule structure applied by Webb et al. (2006) to model wind erosion hazard in Australia (after Wasson and Nanninga, 1986; Leys, 1991a). Tree cover effects on land erodibility were considered by a simple threshold of 20% cover, over which the landscape was assigned a ranking of ‘not erodible’ (Marshall, 1972).

Threshold criteria were applied for assessments of soil moisture and stone cover effects. Land erodibility was assigned a value of 0 (not erodible) where rain was actively falling, or had fallen in the week prior to survey. For all other areas surface soil moisture content was considered negligible and land erodibility a function of soil and vegetation conditions. Stone cover effects were considered by an on-off criteria: where stone cover was dense ($> 70\%$) a ranking of 'not erodible' was applied (Gillette et al., 1980). The final land erodibility ranking was based on an average of the component observation ranks on a four-class scale: high; moderate; low; and not erodible.

6.2.2 Sampling Methodology

The sampling strategy used was adapted from Hassett et al. (2000) and involved applying the above criteria to record visual assessments of land erodibility. Assessments were made from a moving vehicle that was driven through the study area. Computerised Information Gathering System (CIGS) software (Boyce Industries, Brisbane, Australia) was used for data recording. CIGS provides an interface that links observations recorded via computer function keys to latitude and longitude coordinates from a global positioning system (GPS). The software was run from a laptop, while a GPS mounted on the vehicle dashboard provided positional information that was read directly into the CIGS logging interface.

Definition files compiled using a text editor were used to assign a unique observation and data entry space to the laptop function keys. Function keys were used to record assessments by the criteria in Table 6.1, and additional comments (i.e. actively blowing dust). The CIGS interface allows for raster and vector format data to be displayed as a moving background with the observer position and logged data points displayed as an overlay. Soil texture data (Australian Natural Resources Atlas, 2007) was used as a background to facilitate evaluations of soil erodibility.

Visual observations were restricted to areas $\sim 100 \times 100$ m in size. This window was selected to ensure that assessments were made over a regulated area, and would not be significantly influenced by changes in topography or vegetation density. Vehicle speed for data recording averaged $80 - 100 \text{ kmh}^{-1}$. Assessments were made approximately every 60 seconds and more frequently when close to boundaries between land types. Each entry was assigned a date/time

stamp, position (latitude, longitude), positional accuracy and number of satellites for the position reading, followed by the conditional assessments.

6.2.3 Calibration of Observations

A trial survey of 168 observations in May 2006 was used to test the assessment procedure and refine the methodology. Calibration of the vegetation cover estimates was performed by comparison of visual estimates with data collected using a point-intercept transect survey (after Hassett, 2006). Six calibration sites were established across the study area (Figure 6.1). Calibration data were collected at each site in October 2005, September 2006, November 2006 and May 2007. Six 250 m transects were run at 60° intervals radiating from a fixed central point at each site. Observations of plant (lower and upper stratum) and plant litter presence/absence were recorded at 2 m intervals along the transects. The cover fraction of plant litter and herbaceous and woody vegetation were computed by dividing the number of records for each cover type by the total number of observation points per site (750). Visual assessments of cover made at the sites prior to the surveys were then compared with recorded cover fractions (Figure 6.2). Good agreement was found between observed and measured grass cover ($r^2 = 0.85$). The regression equation (Figure 6.2) was used to calibrate the visual cover assessments.

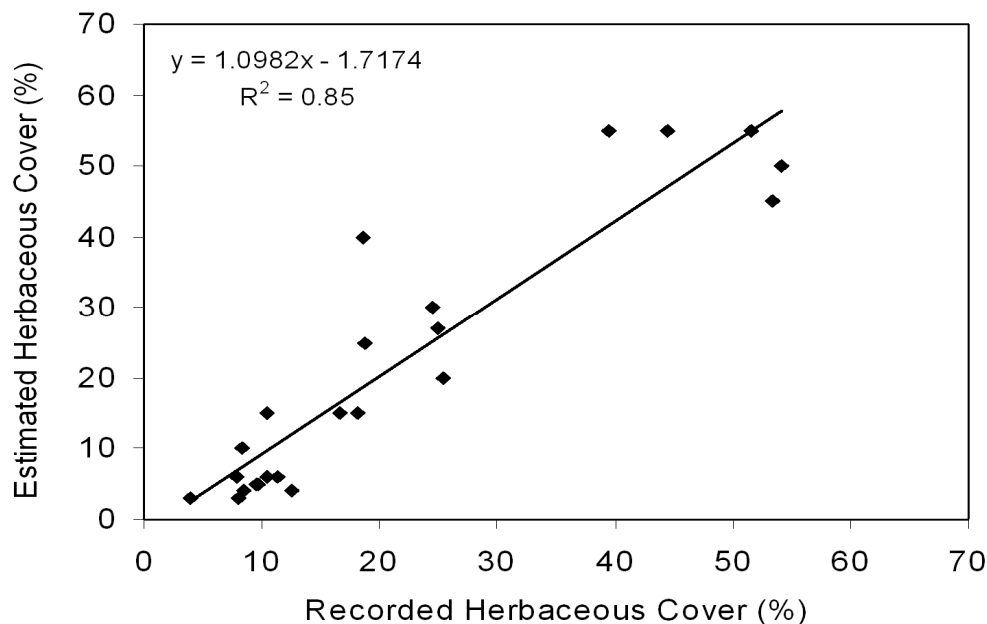


Figure 6.2 Calibration regression of field estimates of herbaceous vegetation cover versus recorded cover based on 24 calibration tests (October 2005 to May 2007).

6.3 Field Data and Model Simulations

6.3.1 Field Data Collection

Three trips were conducted to collect land erodibility assessments across the western Queensland study area: September 2006 (1171 observations); November 2006 (962 observations); May 2007 (1521 observations). Routes were surveyed based on the requirement to assess the condition of all of the major land types within the study area on each trip. This included surveying country in the Mulga Lands, Channel Country, Mitchell Grass Downs, and along the eastern margins of the Simpson Desert (described in Chapter 1, Section 1.6). The routes were also selected so that a range of land types within the bioregions were surveyed. This meant that the transects traversed river channel systems, floodplains, gibber plains, dunefields, open downs, woodlands and escarpments. Routes were revisited to provide multi-date samples of the same areas to build a database of spatial and temporal change in the erodibility of the landscape.

6.3.2 Model Simulations

The Australian Land Erodibility Model (AUSLEM) was used to simulate land erodibility across the study area of western Queensland. The model operates at a 5 x 5 km spatial resolution on a daily time step with inputs of grass and tree cover, soil moisture, soil texture and surficial stone cover (Webb et al., 2009). Previous validation of the model (in Chapter 5) was conducted by comparison of time-series predictions with point observational records of wind erosion activity (Webb et al., 2009). Attempts to validate the model at the 5 x 5 km resolution using visual assessments of land erodibility were restricted by vegetation heterogeneity in the study area which compounded scale differences between the validation data (~100 x 100 m) and model output (5 x 5 km). There is therefore a requirement to test the model performance using higher spatial resolution inputs.

In the current study AUSLEM was run at a 200 x 200 m resolution. Landsat ETM+ derived bare ground (%) and foliage projective cover (%) data were used for the grass (inverse of bare ground) and tree cover model inputs (after Danaher et al., 2004; Scarth et al., 2006). The data were rescaled from 30 x 30 m to the 200 x 200 m spatial resolution to minimise data scale differences and ensure that the validation data points would correspond to individual

model output pixels. The Landsat scenes covered five regions of the study area (Figure 6.1) and were captured on 10, 12, 17, 19 and 26 October 2006. Australian Bureau of Meteorology rainfall records for western Queensland indicate that no rain fell in the validation areas for five weeks prior to data acquisition, so in the absence of high resolution soil moisture data surface soil moisture content was considered static for the simulations.

6.3.3 Model Validation Approach

AUSLEM was run to predict land erodibility on a continuous scale (0: not erodible to 1: highly erodible) for each of the five input data scenes (example shown in Figure 6.3).

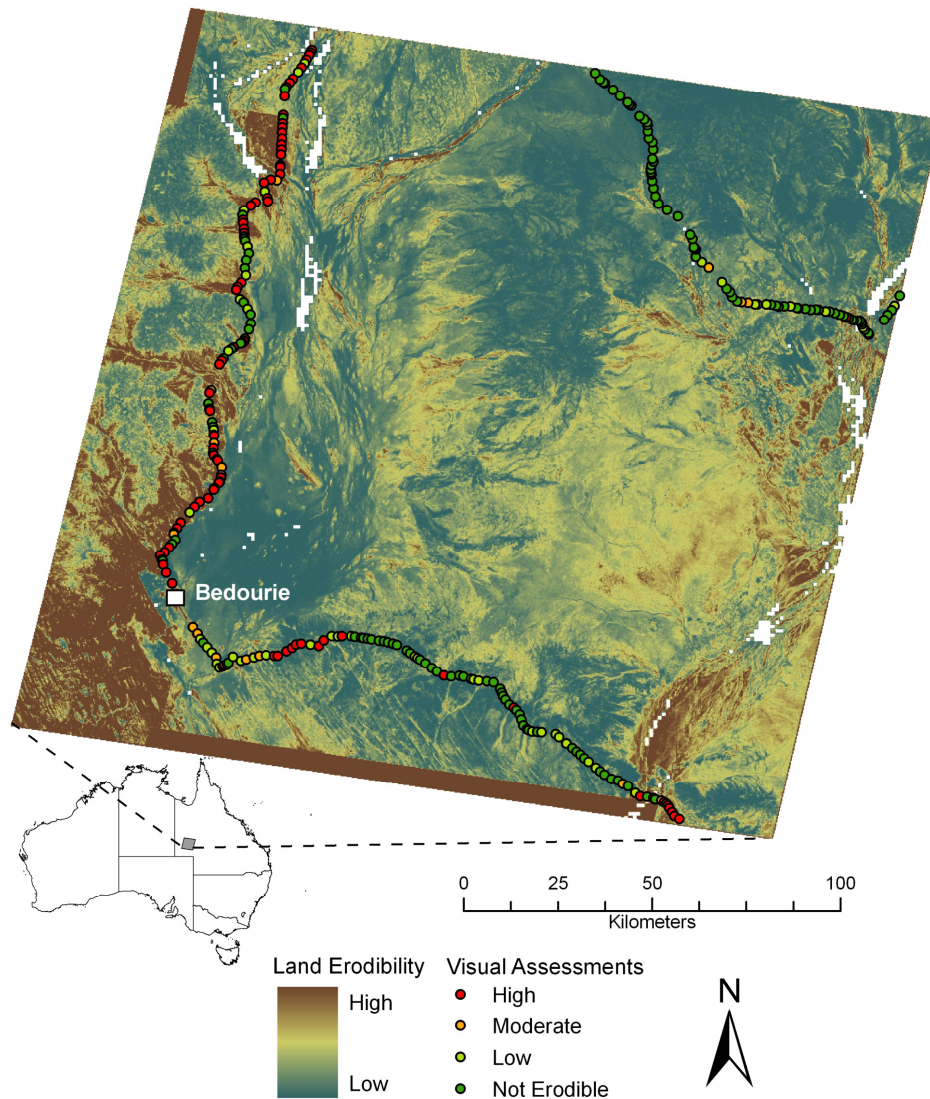


Figure 6.3 Example model output image for the Bedourie scene showing visual assessments of land erodibility as an overlay to the model assessments. White areas are non-erodible with tree cover greater than 20%.

The model output was then compared with visual assessments of land erodibility acquired from 11 to 21 September 2006. The time difference between validation data collection and model simulations was unavoidable and dictated by access to the study area (September) and Landsat image acquisition dates (October). Evaluation of 5 x 5 km grass cover data from the AussieGRASS pasture growth model (Carter et al., 1996a,b) revealed that mean cover change (a net decrease) over the study area was ~2 % between 1 September and 30 October 2006, so the effect of this time difference on the validation outcome is not believed to be significant.

A point-to-pixel extraction method (ArcGIS 9.2, ESRI) was used to retrieve values from model output pixels coincident with the transect point observations (Figure 6.1). Model output values were then plotted against the visual assessments of land erodibility. The validation relies on a comparison of categorical (point observations) and continuous (model output) data, so summary statistics (mean and standard deviation) of the model output were computed for each validation data class (highly erodible to not erodible). Model performance was then evaluated by visual assessment of the distributions of predicted values across the validation data classes.

6.4 Results and Discussion

Figure 6.4 presents plots of predicted versus observed land erodibility for the five output scenes. For each scene the rate of change in the equation for the line of best fit can be used as a qualitative indicator of agreement in the data. For the Mt Dot, Bedourie and Cadell scenes there is agreement in the data, with predicted and observed values increasing from low to high. The trend indicates that for the area covered by these scenes (Figure 6.1) AUSLEM captures the susceptibility of the landscape to wind erosion. There is a weak positive trend in the data for the Windorah scene. Data for the Quilpie scene shows poor agreement between predicted and observed values. Here the model over-predicts land erodibility for the low erodibility class and consistently under-predicts across the moderate and high classes.

The variability in predicted land erodibility values for each observation class differs between the classes and output scenes (Figure 6.4). There is significant overlap in the model predictions between observation classes, and this was expected in comparing the continuous

and categorical data. Two factors affect the class separation and this complicates assessments of model performance.

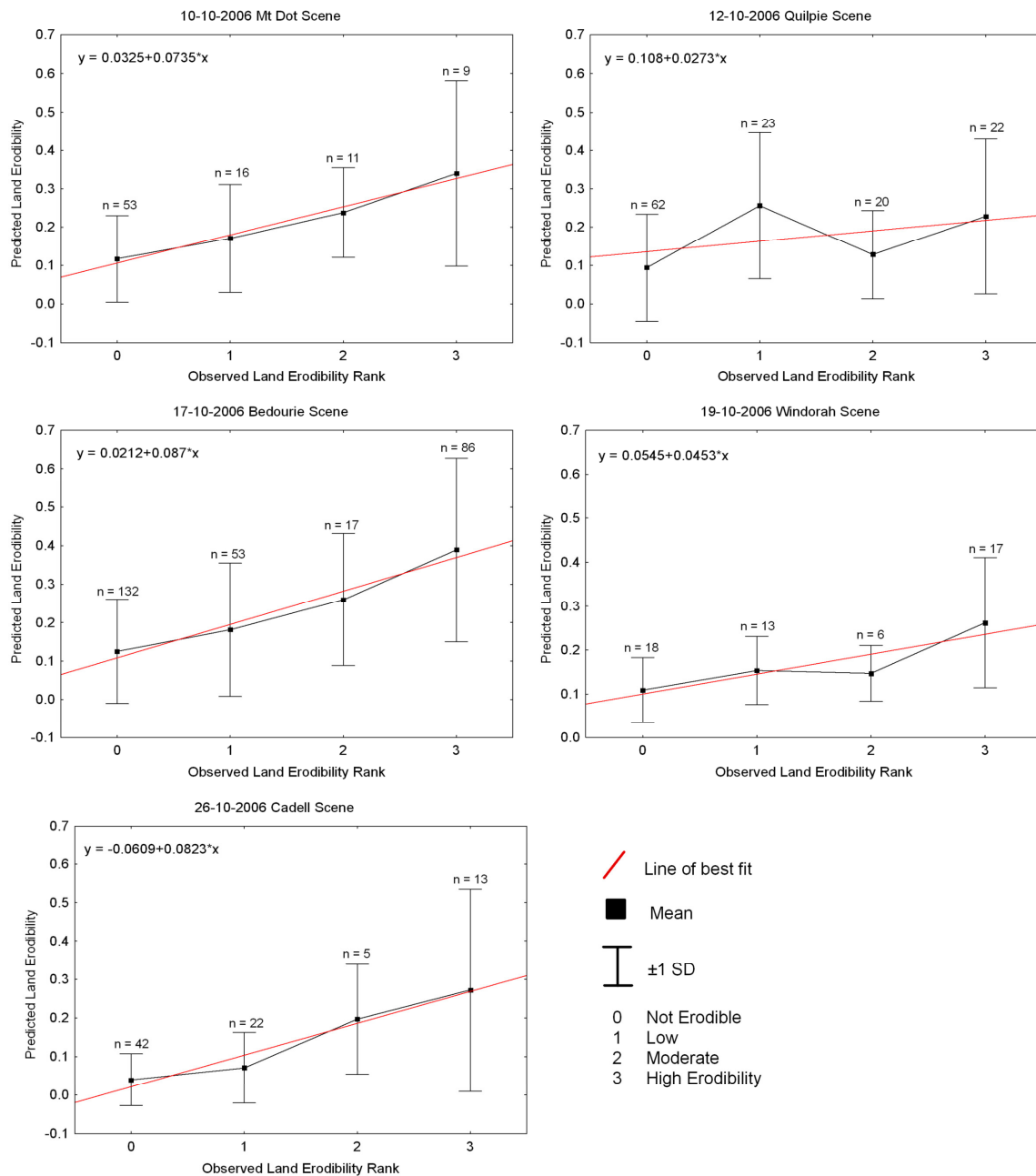


Figure 6.4 Modelled versus observed land erodibility for the five Landsat scenes. The data show the mean ± 1 standard deviation (SD) of the predicted values for each observed land erodibility class. The number of observations (n) is shown for each class.

Firstly, the scale in the model output is non-linear, increasing exponentially between 0 and 1. This non-linearity is driven by a high sensitivity of the model (and wind erosion) to small

changes in vegetation cover at the low end (0 – 15%) of the cover range. This means that modelled erodibility values increase slowly through the ‘no’, ‘low’ and ‘moderate’ classes, then rapidly into the ‘high’ erodibility class. The flat distribution of the data is explained by the variability in predicted values for the high erodibility classes, which indicate that the model has difficulty in representing land erodibility at the high end of the ranking.

Secondly, data scaling issues affect the accuracies of the visual assessments and model predictions of land erodibility. Both are dependent on the arrangement of vegetation in the landscape and are affected by its anisotropic distribution in semi-arid rangelands (Okin, 2005). Assuming a uniform distribution of cover, the model has a tendency to under-estimate erodibility when input pixel vegetation cover values are high. Conversely, visual assessments of erodibility may account for vegetation patchiness, leading to relative over-estimates of land erodibility. This issue is particularly relevant in the Mulga Lands (Windorah, Quilpie scenes), which are characterised by small erodible patches within a broader matrix of well-vegetated land (Pickup, 1985). Future research to address this limitation should examine: implementing the validation at multiple spatial scales; accounting for vegetation distribution in the visual assessments of land erodibility; and including measures of vegetation distribution in the model simulations (Okin and Gillette, 2001).

6.5 Conclusions

This chapter has demonstrated that acquiring visual assessments of land erodibility over large distances using long-range transects can be a useful approach for monitoring landscape condition and testing the performance of regional scale ($>10^4$ km²) land erodibility models. Results suggest that AUSLEM performs better in the western Mitchell Grass Downs and Channel Country (Mt Dot, Cadell, Bedourie scenes) than in the Mulga Lands (Windorah, Quilpie scenes). To the knowledge of the author the long-range transect approach described in this chapter has not previously been applied to test the performance of a wind erosion model. Further development and application of methods for assessing land erodibility at the landscape scale will improve our capacity for monitoring and modelling land degradation processes in remote desert environments.

Chapter 7

Simulations of the Spatio-Temporal Aspects of Land Erodibility in the North-East Lake Eyre Basin, Australia, 1980 – 2006

This chapter addresses Objectives 7 and 8. The chapter describes application of the land erodibility model to assess spatial and temporal patterns in land erodibility in western Queensland from 1980 to 2006. The spatio-temporal dynamics are then related to climatic processes driving regional scale variations in the condition of the western Queensland rangelands.

7.1 Introduction

Wind erosion and mineral dust emissions play important roles in land surface and atmospheric processes at local ($<1 \text{ km}^2$) to global scales (Miller and Tegen, 1998; Lal, 2001; Jickells et al., 2005). While dust emission and transport mechanisms have received much attention, the spatial extent of dust source areas and conditions driving temporal variations in land erodibility remain poorly understood (McTainsh et al., 1999; Mahowald et al., 2003a). Analyses of atmospheric dust concentrations indicate that land erodibility is a dynamic condition that is highly variable through space and time (Brooks and Legrand, 2000). Spatial and temporal variations in land erodibility at the landscape scale (10^3 km^2) are important controls on regional scale (10^4 km^2) dust transport (Gillette, 1999). While numerous wind erosion modelling systems have been developed to assess wind erosion dynamics at regional to global scales, e.g. DPM (Marticorena and Bergametti, 1995), DEAD (Zender et al., 2003a), GOCART (Ginoux et al., 2004), there remains a significant lack of research into the application of models to assess land erodibility dynamics at the landscape scale.

Mapping and monitoring land susceptibility to wind erosion is required to enhance land management to combat land degradation and desertification of dryland environments (Oldeman, 1994, Sivakumar, 2007). Quantifying rates of anthropogenic (accelerated) and naturally occurring wind erosion is a fundamental component of this work. This is dependent on an understanding of factors driving variations in dust emissions at the landscape scale (Mahowald et al., 2003b). Modeling spatial and temporal patterns in land erodibility can be used to establish baseline levels of variability in the landscape response to climate and land management conditions. The effects of climate and land management changes on potential erosion can then be assessed with knowledge of the likely response of landscapes to these external drivers. Increasing pressures on natural resource use in rangelands (Galvin et al., 2008), and increasing aridity and rainfall variability in the sub-tropics (Hu and Fu, 2007; Meehl et al., 2007) make understanding these land surface-climate-management dynamics essential for the sustainable use of the world's drylands.

A number of techniques have been employed to determine the location and extent of areas susceptible to wind erosion. Prospero et al. (2002) and Washington et al. (2003) used aerosol optical thickness data from the Total Ozone Mapping Spectrometer (TOMS) to characterize persistent global dust source areas. The studies demonstrated an association of source areas with internally draining river systems and topographic depressions. Modeling dust source areas using topographic erodibility indicators was subsequently employed by Ginoux et al. (2001), and Zender et al. (2003b) in a comparison of dust emission simulations using geomorphic and hydrological erodibility factors. The application of surface reflectance factors derived from Moderate Resolution Imaging Spectroradiometer (MODIS) data to define erodible land areas has also been investigated (Grini et al., 2005). A limitation of these approaches is their coarse spatial resolution (typically $\sim 1^\circ$ lat./long.) and inability to detect local temporal variations in source strength (land erodibility).

At the regional scale (10^4 km^2), maps of areas affected by wind erosion tend to be interpretive and based on climatic indicators (e.g. aridity), or on the frequency at which dust storms are recorded over a particular area (Leys, 1999). Similarly, temporal changes in land erodibility have been inferred from seasonal and annual trends in observed dust-storm frequencies (Goudie and Middleton, 1992). The detail of spatial and temporal variations in land erodibility is therefore very coarse, as vast distances ($>100 \text{ km}$) often separate stations recording dust events in arid and semi-arid areas (McTainsh and Leys, 1993). The growing

availability of remote sensing products and model simulations of land surface conditions, e.g. vegetation cover and soil moisture, presents the opportunity to map and monitor land erodibility dynamics at moderate to high resolutions ($<10^3 \text{ m}^2$, daily – monthly), and address the lack of research in this area.

This research seeks to quantify patterns in land erodibility dynamics in the north-eastern portion of the Lake Eyre Basin, Australia; arguably the largest dust source in the Southern Hemisphere (Goudie and Middleton, 2006). The aim of this chapter is to analyse simulations from the Australian Land Erodibility Model - AUSLEM (Webb et al., 2009) to: 1) identify spatial patterns in land erodibility in the north-east sector of the Lake Eyre Basin, western Queensland; 2) identify the frequency at which areas are susceptible to wind erosion in this region; and 3) identify temporal patterns in land erodibility, with a focus on documenting mechanisms driving seasonal and inter-annual variations in land erodibility dynamics.

7.2 Study Area

The study area for the model application is the arid and semi-arid rangelands located in the north and northeast of the Lake Eyre Basin in western Queensland (Figure 7.1; Chapter 1).

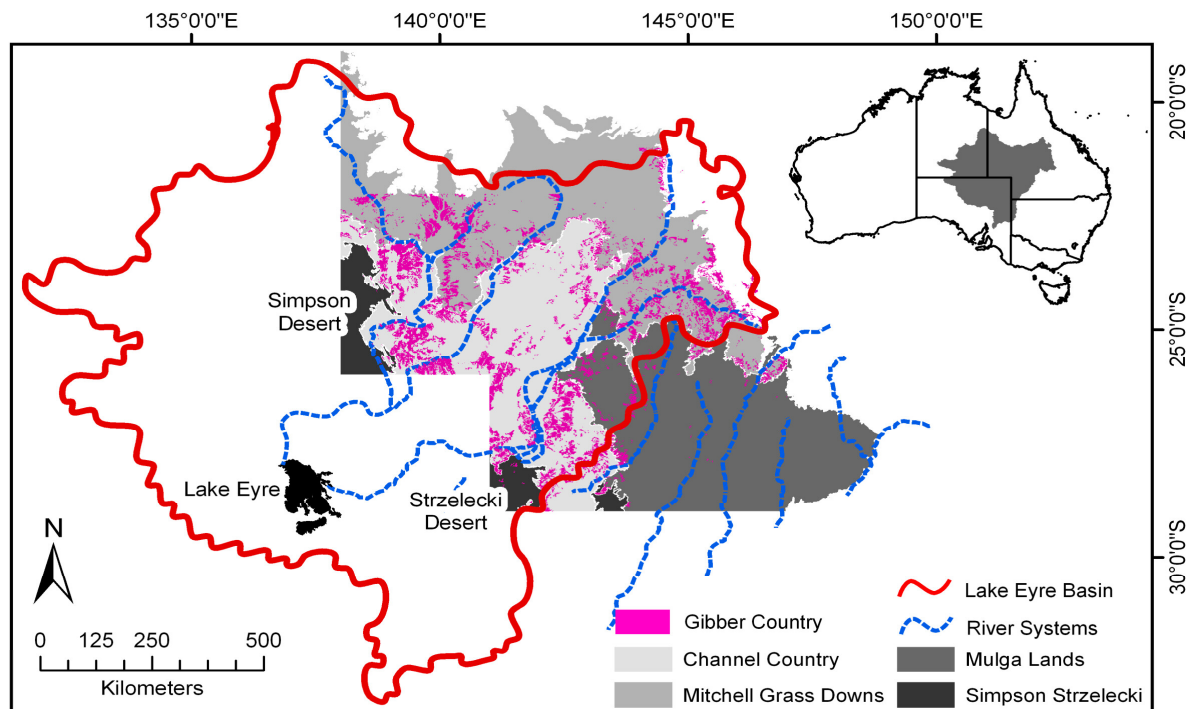


Figure 7.1 Map showing the study area location within the Lake Eyre Basin, Australia, and major geomorphic features relevant to this study.

7.3 Model Description

Land susceptibility to wind erosion, i.e. land erodibility, was modelled with the Australian Land Erodibility Model (AUSLEM). AUSLEM was developed by Webb et al. (2009), and a description of the model formulation, input parameters and performance characteristics are presented in Chapter 5. For the current analysis the model was run at a 5 x 5 km spatial resolution on a daily time-step. In the absence of an operational model to predict temporal changes in soil erodibility (Chapter 4), soil textural effects, $E_{tx}(tx)$, were considered static (after Lu and Shao, 2001) at time scales less than one month. In doing this the analysis of model output was restricted to scales $\sim 10^4$ km² at which the input conditions were found to best model land erodibility (Webb et al., 2009; Chapter 5, Section 5.5.3).

7.4 Model Simulation and Analysis Methods

7.4.1 Spatial Patterns

AUSLEM was run to assess land erodibility on a daily time-step from January 1980 to December 2006. Output was then aggregated into monthly and annual means. Patterns in the distribution of erodible land were examined by assessing the frequency at which model output was recorded in four land erodibility classes. Arbitrary class boundaries were defined for land that was highly erodible (1–0.15); moderately erodible (0.15–0.0375); had low erodibility (0.0375–0.0091); and was not erodible (0.0091–0). The class boundaries were necessarily non-linear to account for the exponential increase in land erodibility with decreasing vegetation cover and soil moisture. Mean monthly model output was classified into the four erodibility groups, and the percentage of years (out of 27) in which land occurred in each class was computed on a per-pixel basis.

7.4.2 Seasonal Variability

Seasonal variations in land erodibility were examined by analysis of point time-series data. Mean monthly model output values were extracted from areas with 50 km radius around meteorological stations at Birdsville, Boulia, Charleville, Longreach, Thargomindah, Quilpie, Urandangie and Windorah (Figure 7.1). Summary statistics of the time-series data were

computed to provide information on mean monthly land erodibility conditions and the ranges of monthly variations. Trajectories of change in land erodibility within the four study area bioregions (Chapter 1, Section 1.6) were then examined by computing the proportional abundance (percentage cover) of each land erodibility class within the bioregions. The analysis was conducted at an annual time scale to provide an indicator of the response of the bioregions to inter-annual climate variability.

7.4.3 Inter-Annual Variability

The effects of drought on wind erosion activity have been examined from observational data on atmospheric aerosol concentrations and dust-storm frequencies in Africa, the Middle East, North America, China and Australia (McTainsh et al., 1989; Goudie and Middleton, 1992; Brooks and Legrand, 2000; Okin and Reheis, 2002; Prospero and Lamb, 2003; Zender and Kwon, 2005). Advances in modelling dust emissions at the global scale have enabled links to be established between atmospheric dust concentrations and climate forcing mechanisms at multiple temporal scales, including the North Atlantic Oscillation (NAO), El Niño/Southern Oscillation (ENSO), Pacific (inter-) Decadal Oscillation (PDO), and others (Moulin et al., 1997; Mahowald et al., 2003b; Ginoux et al., 2004; Hara et al., 2006). At regional scales the effects of these global teleconnections may be manifested through changes in precipitation and windiness, and have been shown to influence the onset, intensity and duration of drought and periods of enhanced rainfall (e.g. Allan, 1988). It is feasible to hypothesise that land erodibility in western Queensland will also be dynamic and respond to regional to global scale climate variability.

Pittock (1975) demonstrated that annual rainfall is correlated ($r^2 = \sim 0.4$) with the Troup Southern Oscillation Index (SOI) over eastern Australia (east of 138° longitude). The correlation between rainfall and the SOI varies seasonally, and is influenced by phase interactions of ENSO (3-7 year cycle) with the PDO (15-30 year cycle) (McBride and Nicholls, 1983; Power et al., 1999; Crimp and Day, 2003). Consequently, episodes of pasture degradation and recovery are linked to variations in Australian rainfall driven by ENSO-PDO interactions (McKeon et al., 2004). The final phase of the model output analysis sought to quantify the relationships between dynamic changes in land erodibility, rainfall over western Queensland, and these teleconnections.

A cross-correlation analysis was used to investigate the presence and strength of relationships between mean annual land erodibility and annual total rainfall averaged across the four study area bioregions, and two indicators of the condition of ENSO and the PDO. The Troup Southern Oscillation Index (SOI) was used in the analysis as a measure of ENSO fluctuations. The Troup SOI data were obtained from the Australian Government Bureau of Meteorology (BoM, 2007b) and represent the standardised anomaly of the Mean Sea Level Pressure difference between Tahiti and Darwin (Troup, 1965). Monthly PDO Index data (JISAO, 2007) were used to define PDO phases and represent the first principal component of monthly sea surface temperature variability in the North Pacific Ocean (Mantua et al., 1997). All data were converted to annual averages for the correlation analysis.

7.5 Results

Figure 7.2 presents mean annual land erodibility assessments for western Queensland from 1980 to 2006. The following sections analyse patterns in mean monthly model output and the data presented in Figure 7.2. First, spatial patterns in land erodibility are described. Temporal patterns in land erodibility dynamics are then resolved at seasonal and inter-annual time scales.

7.5.1 Spatial Patterns in Land Erodibility

Figure 7.3 summarises spatial patterns in land erodibility in western Queensland. Three regions of the study area have consistently high land erodibility. These include the Mulga Lands, Strzelecki Desert, and western side of the Channel Country (Figure 7.1). In the Mulga Lands high erodibility land occurs in the mulga (*Acacia aneura*) plains east of the Bulloo River between Quilpie and Thargomindah (Figure 1.3). This region extends into the flood-out country and dunefields to the south of the Bulloo River, and across into the Strzelecki Desert dunefields. A third region of high erodibility land lies between Birdsville and Urandangie along the outer-floodplains of the Georgina and Diamantina Rivers. The climates of the areas with consistently high land erodibility are similar. These areas receive <300 mm mean annual rainfall and experience annual evaporation in excess of 2400 mm (BoM, 2007). The conditions result in a moisture deficiency and sparse vegetation cover that is highly sensitive to grazing pressures (McKeon et al., 2004).

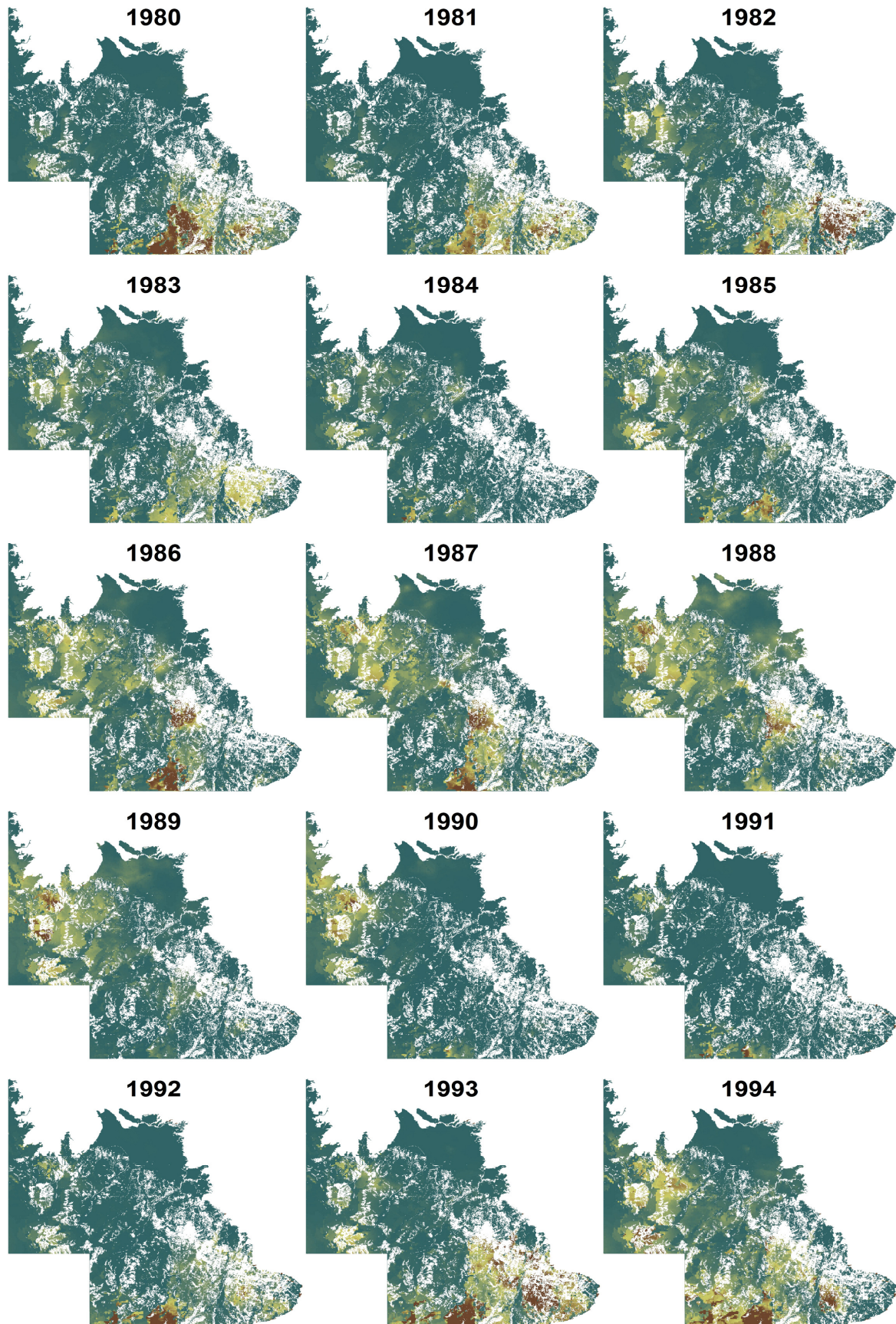


Figure 7.2 Mean annual land erodibility predictions from AUSLEM for the period 1980 to 1994. White areas are not erodible due to tree and stone cover being above the model thresholds.

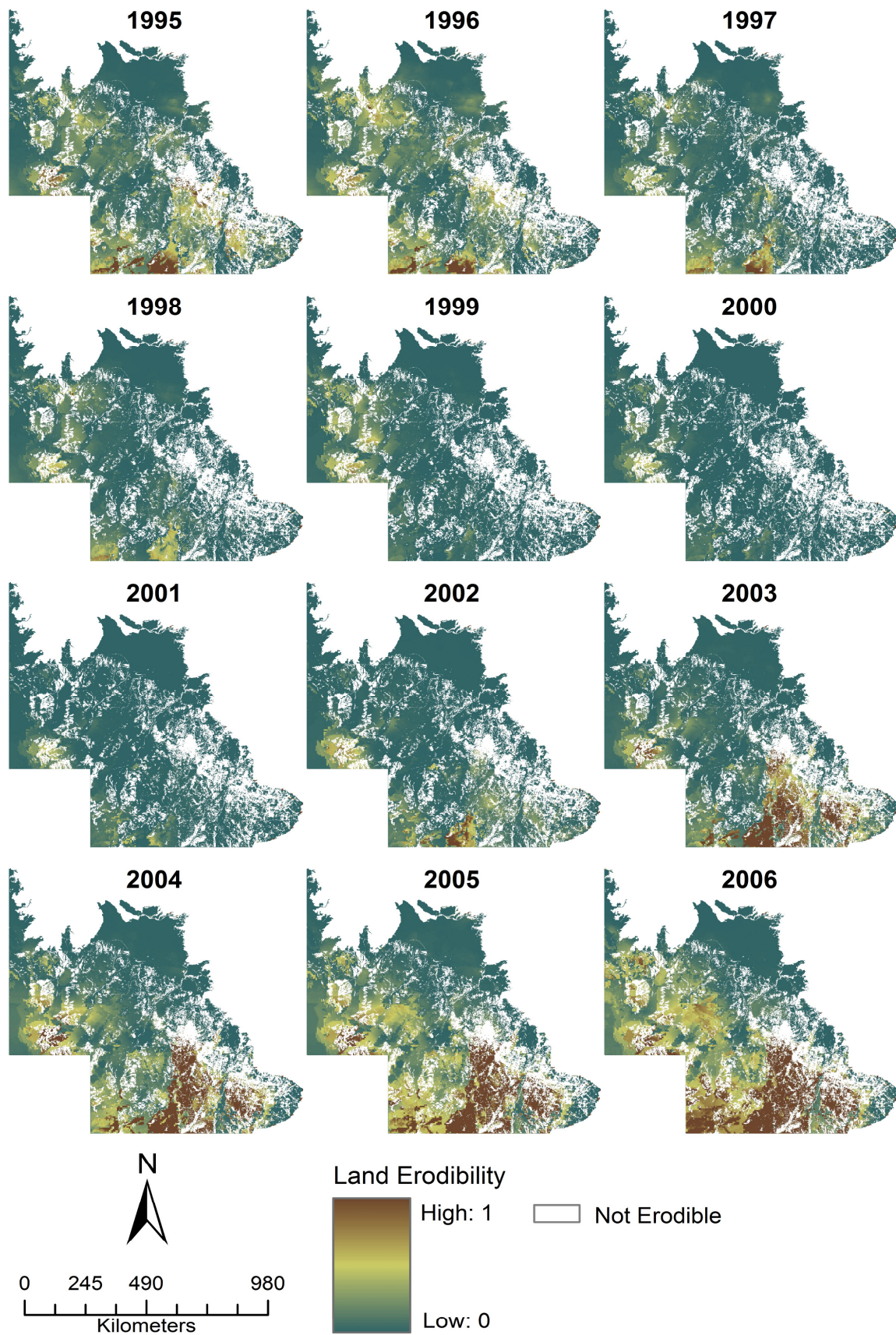


Figure 7.2 continued, showing mean annual land erodibility predictions from AUSLEM for the period 1995 to 2006.

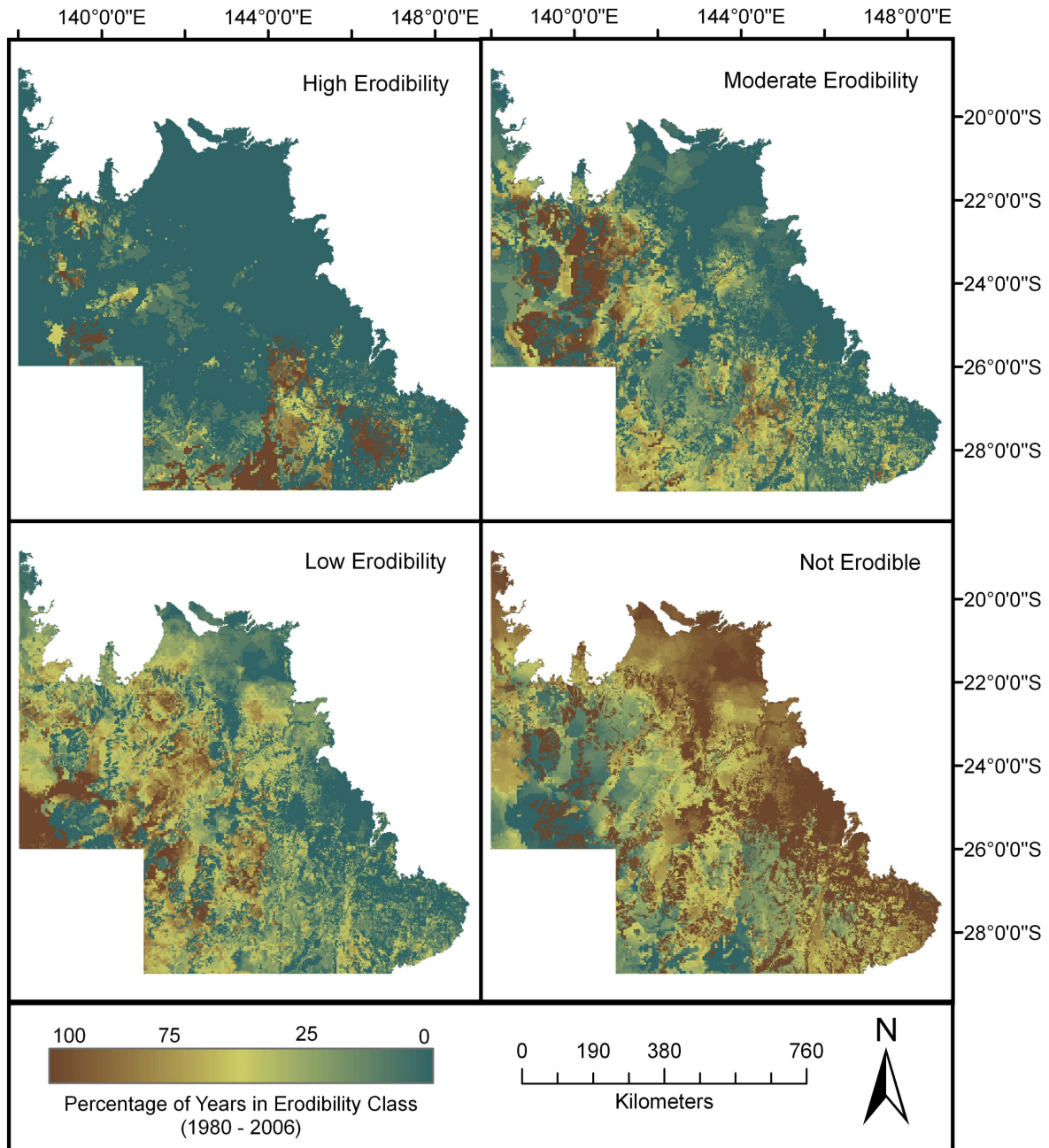


Figure 7.3 Map showing the percentage of years in the period 1980 to 2006 in which land in the study area was modelled as having high, moderate, low and no susceptibility to wind erosion.

Areas that frequently have a moderate land erodibility ranking extend from the high erodibility regions. These areas expand across the driest parts of the study area, covering the western and southern portions of the Channel Country, the Strzelecki dunefields and the western half of the Mulga Lands. A belt of moderately erodible land frequently extends north-east from the Channel Country into the Mitchell Grass Downs near Longreach (see Figure 7.1; Chapter 1, Section 1.6).

Land areas with consistently low and no erodibility occur across the northern and eastern boundaries of the study area in the Mitchell Grass Downs and Mulga Lands (Figure 7.2). This is consistent with the regions receiving on average the highest annual rainfall across the study area (McTainsh and Leys, 1993). The Channel Country frequently has an elevated land erodibility ranking. The only areas of the bioregion that consistently have no erodibility are the gibber plains (stony country) and inner-floodplains of the Cooper, Diamantina and Georgina river systems. The land area extending between Coopers Creek and the Bulloo River had a low erodibility for much of the period between 1980 and 2006 (Figure 7.2). This is interesting given that surrounding regions generally had higher erodibility rankings. In the south-eastern corner of the study area the mulga plains around Charleville generally have a low susceptibility to wind erosion (Figure 7.3). Extensive surficial stone cover (gibbers) and high tree cover levels reduces the erodibility of these areas. The model identified the Simpson Desert bioregion as having consistently low land erodibility.

Together the areas identified as having high to moderate land erodibility match the spatial pattern of areas recording the highest dust-storm frequencies in the north-eastern half of the Lake Eyre Basin (Middleton, 1984; Burgess et al., 1989; McTainsh et al., 1990). Dust-storm frequencies decline to the east of Longreach and Charleville (Figure 7.1) and the model assessments of land erodibility are consistent with this pattern. Miles and McTainsh (1994) and McTainsh et al. (1999) reported high spatial variability in wind erosion activity in the Channel Country and Mulga Lands. This is reflected in the heterogeneity of land erodibility in these regions (Figure 7.2).

7.5.2 Temporal Dynamics in Land Erodibility

Seasonal Variations

Figure 7.4 presents graphs of mean monthly land erodibility for eight meteorological stations in western Queensland. Land erodibility has weak seasonality. The strength of seasonal variations changes across the study area and between land types. The largest seasonal variations occur at Quilpie (range 0.027), Thargomindah (range 0.018) and Urandangie (0.01). The smallest variations occur at Boulia (range 0.004), Charleville (range 0.003) and Windorah (range 0.001).

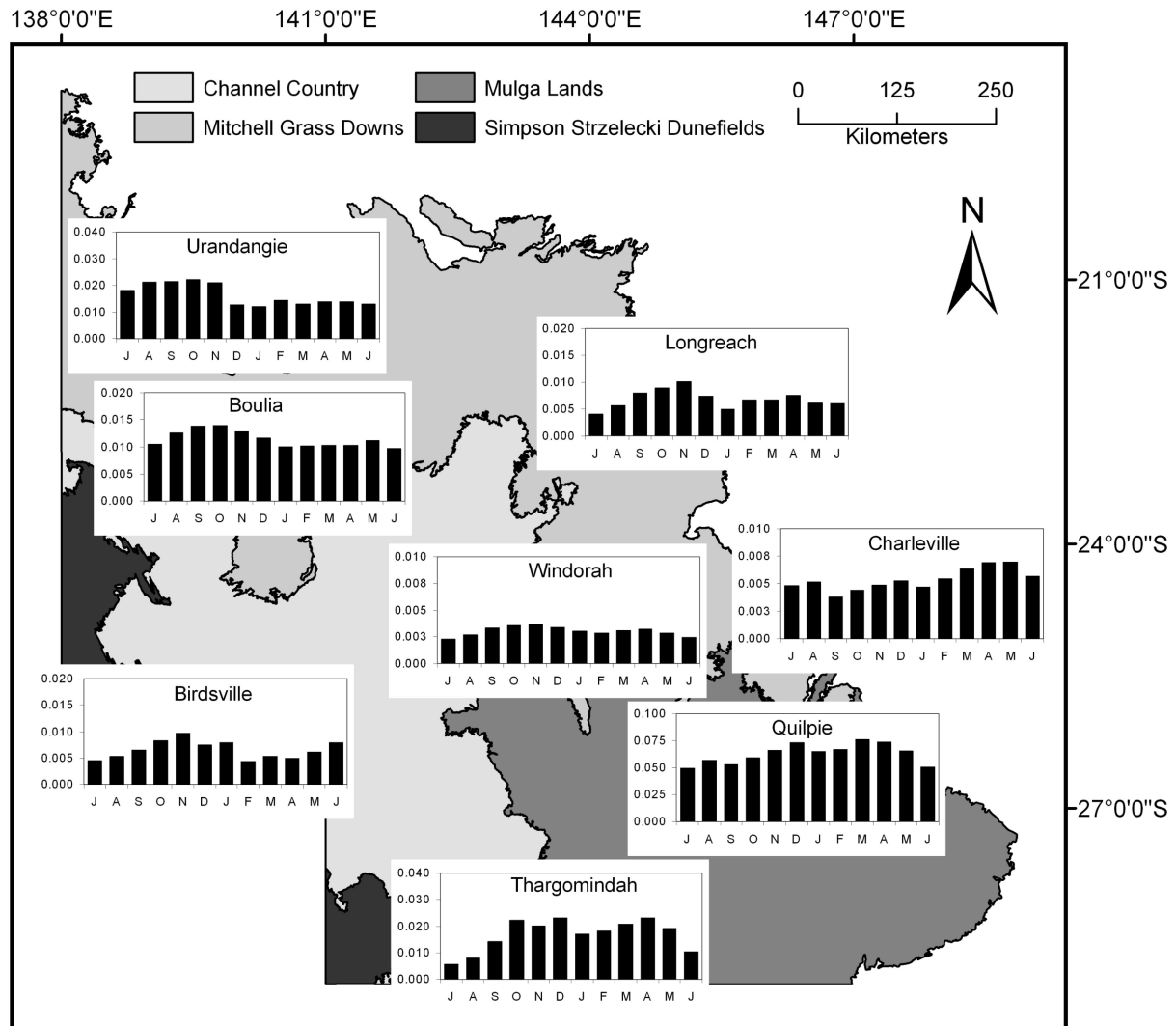


Figure 7.4 Graphs of mean monthly land erodibility (from 0: not erodible, to 1: high erodibility) for eight stations across the study area, based on modelled daily land erodibility data (1980-2006) extracted from areas with 50 km radius around the stations.

There is a north-south spatial pattern in the seasonality of land erodibility across western Queensland (Figure 7.4). Land erodibility reaches a maximum in spring (SON) and early summer (ND) in the Channel Country and Mitchell Grass Downs, and a minimum in late summer (JFM) through to winter (JJA). In the Mulga Lands to the south of the study area land erodibility peaks in spring (SON) and autumn (MAM), and reaches a minimum over winter (JJA). These dynamics are consistent with seasonal patterns in dust-storm frequencies reported by Eckström et al. (2004). The northern pattern reflects a dominance of summer rainfall associated with thunderstorm activity, and a reduction in grass cover through winter and spring associated with lower rainfall. The southern pattern reflects a tendency of the

Mulga Lands to receive a less pronounced peak in rainfall over summer and a response to higher winter rainfall (McTainsh et al., 1998).

Inter-Annual Variability

Regional changes in erodible land (Figure 7.2), and weak seasonal variability (Figure 7.4) suggest that land erodibility dynamics in western Queensland are driven by forcing mechanisms operating at longer (> monthly) time scales. Figure 7.5 shows the annual percentage area of each bioregion covered by land with modelled high, moderate, low and no susceptibility to wind erosion.

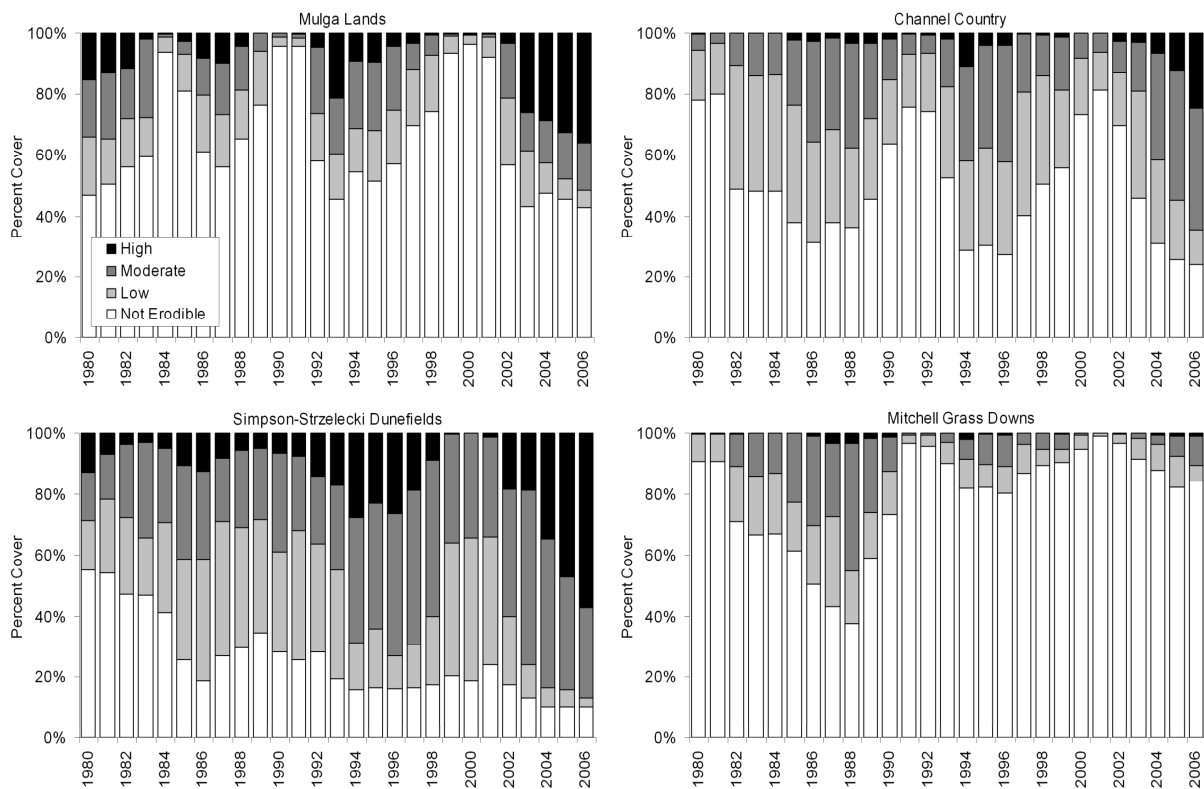


Figure 7.5 Graphs of annual proportional abundance (percentage cover) of land in the four study area bioregions classified into four land erodibility groups: high, moderate, low, and not erodible.

Land erodibility dynamics differ markedly between the four bioregions in western Queensland. In the Mulga Lands, land erodibility displays cyclic behaviour, with alternating periods of high and low susceptibility to wind erosion. In years of reduced erodibility (e.g. 1984, 1990-91, 1999-2001) the Mulga Lands appear to 'shut down' in terms of the area

assessed as being susceptible to wind erosion, with >90% of the bioregion indicated as being not erodible.

In the Channel Country, peaks in land erodibility occurred in similar periods to the Mulga Lands. However, for every year in the period 1980-2006 >20% of the bioregion had at least a low to moderate susceptibility to wind erosion. In no years did the model indicate a near-complete (>90%) reduction in land erodibility. This consistently erodible condition is a result of the aridity of the bioregion and its inherently low vegetation cover levels.

Land erodibility in the Mitchell Grass Downs was generally low for the analysis period. Temporal changes in land erodibility in this bioregion are distinct in comparison to the other regions. Peaks in erodible land occur at similar times to those in the Channel Country; however, the magnitudes of the peaks are not consistent. The portion of the Mitchell Grass Downs modelled as not being susceptible to wind erosion did not drop below 75% between 1993 and 2006.

Extensive areas (>40%) of the Simpson-Strzelecki bioregion had some susceptibility to wind erosion in each year from 1980-2006. This high fractional cover is a result of the Strzelecki Desert frequently having moderate to high land erodibility. Figure 7.2 shows that the Simpson Desert dunefields did not experience regular high magnitude changes in erodibility between 1980 and 2006. These dynamics are evident in the Strzelecki Desert and are indicative of higher grazing pressures, vegetation characteristics (dunefields populated with short-lived grasses (*Aristidia* spp.) and forbs rather than drought-resistant hummock grasses (*Triodia* spp.)), and the higher sensitivity of the region to climate variability.

Land Erodibility, Climate Oscillations and Rainfall

Table 7.1 summarises the correlations between annual rainfall, the SOI, PDO and modelled erodible land cover. There is a negative correlation between rainfall and land erodibility ($r^2 = -0.48$, $p < 0.05$) in the Mulga Lands. The relationship weakens in the Mitchell Grass Downs and Channel Country, and is very weak in the Simpson-Strzelecki bioregion. There are no significant or consistent correlations between the SOI and land erodibility. However, there is a weak negative correlation in the Mulga Lands. This is supported by the rainfall-erodibility relationship and a positive correlation ($r^2 = 0.56$, $p < 0.05$) between the PDO and land erodibility. There are weak correlations between land erodibility and the PDO in the Channel

Country and Mitchell Grass Downs, but there is no correlation between the PDO and erodible land cover in the Simpson-Strzelecki Dunefields.

Table 7.1 Correlation (r^2) between mean annual rainfall, Troup SOI, PDO and modelled land erodibility for the four study area bioregions, based on the 27 year simulation. Significant correlations ($p < 0.05$) are **boldfaced**

Bioregion	Rainfall	SOI	PDO
Channel Country	-0.19	-0.05	0.28
Mitchell Grass Downs	-0.23	0.09	0.36
Mulga Lands	-0.48	-0.38	0.56
Simpson-Strzelecki Dunefields	-0.09	0.02	-0.16

Figure 7.6 presents a plot of mean annual rainfall and the Troup SOI for the four study area bioregions. The data are used to interpret mechanisms driving the temporal patterns and correlations between land erodibility, rainfall and the climate indices (Table 7.1).

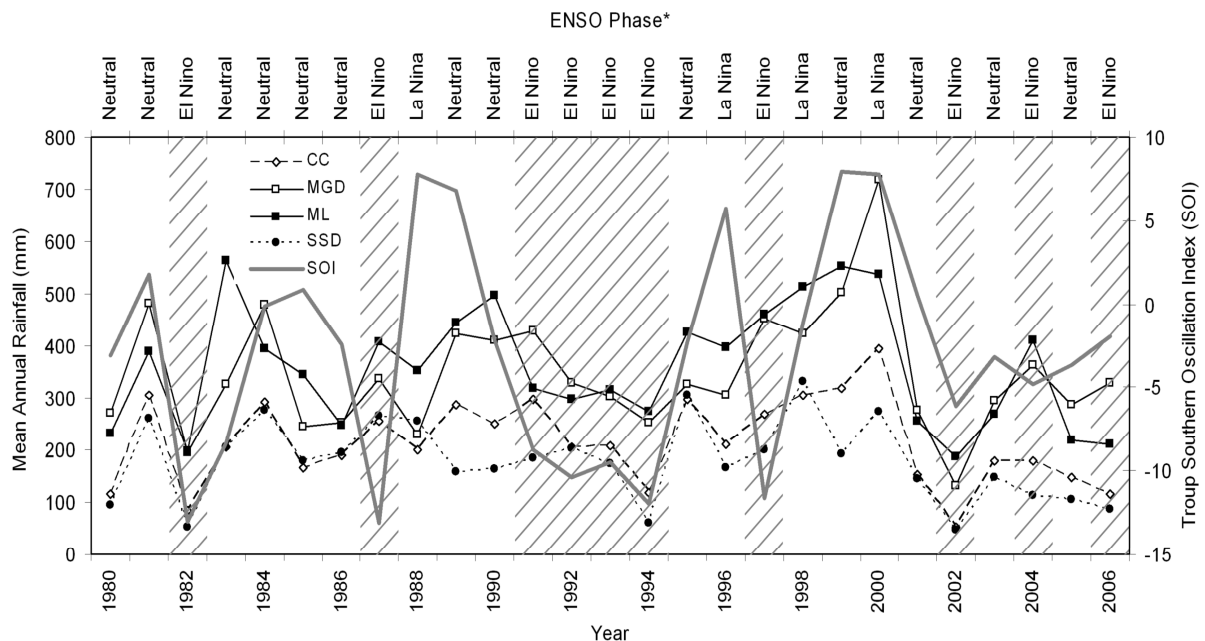


Figure 7.6 Graph of the Troup SOI and mean annual rainfall for the four study area bioregions: Channel Country (CC); Mitchell Grass Downs (MGD); Mulga Lands (ML); and Simpson-Strzelecki Dunefields (SSD). *Years are classified into ENSO phases after McKeon et al. (2004)

In the Mulga Lands peaks in land erodibility coincide with negative SOI phases in 1982, 1986, 1992-94 and from 2002-06 (Figure 7.6). The peaks are consistent with those years receiving low annual rainfall (<300 mm). Reductions in erodible land in the remaining years

are consistent with those years receiving annual rainfall >400 mm. However, the correlations between rainfall, SOI and land erodibility in the Mulga Lands are affected by a lag response in land erodibility change relative to shifts in the SOI phase (positive to negative). This is evident in the transition from the wet La Niña conditions from 1988-90 to the dry El Niño event from 1991-94. A mechanism behind the lag is the steady rather than rapid increase in annual rainfall from 1988-90, and sustained vegetation cover over the bioregion until the drought in 1992. Short (~1 year) El Niño events, for example in 1997, do not necessarily cause a reduction in rainfall. Land erodibility may therefore continue to increase or decrease through these periods in line with trends determined by antecedent rainfall conditions.

The Mitchell Grass Downs, Channel Country and Simpson-Strzelecki Desert bioregions are less sensitive than the Mulga Lands to inter-annual rainfall and ENSO variability (Table 7.1). These bioregions appear to be more sensitive to intense and persistent (multi-year) drought. This explains the stronger positive correlations of land erodibility with the PDO in the Channel Country and Mitchell Grass Downs. The lower sensitivity of these bioregions to inter-annual climate variability stems from the nature of climate variability in these regions, and their vegetation and soil characteristics.

While the Mitchell Grass Downs experiences high inter-annual rainfall variability (Figure 7.6), rainfall across this bioregion is adequate to sustain regionally high vegetation cover levels and low land erodibility (Figures 7.2 and 7.5). Peaks in land erodibility in the Mitchell Grass Downs occur in the north-western half of the bioregion. This area, between Boulia and Urandangie, receives lower annual rainfall than the eastern half of the bioregion and is subsequently more sensitive to drought (e.g. 1986-1989). While rainfall has a strong association with El Niño events in the Mitchell Grass Downs (McKeon et al., 2004), variability over the eastern half of the bioregion is not sufficient to induce regular regional scale (10^4 km^2) changes in land erodibility that would provide strong correlations with either rainfall or the SOI.

Like the Mitchell Grass Downs, land erodibility dynamics in the Channel Country reflect rainfall variability and the PDO more than ENSO phase changes (Table 7.1). The gradual increase in land erodibility in the Channel Country between 1982 and 1988 is consistent with the sustained 'neutral' ENSO conditions and intensification of the PDO (cooling of Sea Surface Temperatures in the tropical West Pacific). The bioregion received <200 mm annual

rainfall in that period (Figure 7.6). Subsequent peaks in land erodibility in 1994-96 and 2004-06 follow similar extended (>4 year) periods of low rainfall. The Channel Country land erodibility response to rainfall displays a lag similar to that in the Mulga Lands. This means that short (1 year) periods of increased annual rainfall, for example 200 mm in 1995, may not induce an immediate regional scale decline in land erodibility.

The peak in land with low erodibility in the Simpson-Strzelecki Dunefields in 1986 reflects the pattern of drying experienced in the Channel Country and Mitchell Grass Downs at that time (Figure 7.5). Changes in the area covered by moderate to highly erodible land occurred in the El Niño drought years of 1994-96 and 2001-06 (Figure 7.6). A reduction (down to <1% of the bioregion) in highly erodible land was only experienced from 1999 to 2000 following three successive years with annual rainfall >300 mm (Figure 7.6). The net change in erodible land in the Simpson-Strzelecki Dunefields during this period was <10% (Figure 7.5). This suggests that the susceptibility of the bioregion to wind erosion may remain consistent for a number of years unless a significant change in local climate takes place. This is also illustrated by the large increase in highly erodible land following six years of annual rainfall <150 mm from 2001-06.

7.6 Discussion

7.6.1 Spatial Patterns in Land Erodibility

Spatial patterns in land erodibility identified by AUSLEM are in agreement with observational records of wind erosion activity across western Queensland (Middleton, 1984; Burgess et al., 1989; McTainsh et al., 1990; Chapter 1, Section 1.2). The landforms associated with erodible lands, i.e. river floodplains and dunefields, are consistent with those in other global dust source areas (Washington et al., 2003). The model assessments of low land erodibility over the Simpson Desert are not, however, consistent with records of wind erosion activity in that area (McTainsh et al., 1990). The assessments are contrary to the presence of the longitudinal dunefields that traverse the bioregion. This dunefield becomes an active dust source area following the removal of vegetation by fire (McGowan and Clark, 2008).

A number of factors affect model performance in the Simpson Desert. Firstly, sub-grid scale averaging of vegetation cover over the model 5 x 5 km grid cells resulted in an inability to detect narrow (10-15 m) erodible dune crests that have an average spacing of ~500 m (Purdie, 1984). Secondly, the Aussie GRASS simulations of herbaceous vegetation cover, used as input to AUSLEM, are driven by a plant growth-death-consumption scheme (Carter et al., 1996). The Simpson Desert is delineated as a National Park, so plant consumption rates in the Aussie GRASS model are kept low (low effective stocking rates), resulting in consistently high predictions of grass cover. Finally, fire burn scars were not delineated in the model input data. In November 2001 and January 2002 extensive burning of the Simpson Desert resulted in the denudation of ~4000 km² (10%) of the bioregion. These events resulted in a proportional increase in the area of the bioregion susceptible to wind erosion. However, the burn scars had not been included in the model input data at the time of acquisition so the model has indeed underestimated the extent of erodible land between 2001 and 2006. These issues highlight the importance of data resolution and currency in the performance of spatially explicit wind erosion models. Updating the model inputs to account for fire burn scars and running the model at a higher spatial resolution, e.g. 30 m, would be required to improve model performance in the Simpson Desert bioregion.

7.6.2 Temporal Patterns in Land Erodibility

Factors affecting temporal patterns in land erodibility include: climate, in particular rainfall quantities and distribution; land management practices, manifested here through stocking rates that affect the model grass cover inputs, and finally; the sensitivity of the bioregions to these external forcing mechanisms as governed by geomorphology, soil characteristics and vegetation structure and resilience. Seasonal variations in land erodibility are weak but follow a north-south spatial pattern in response to differences in rainfall seasonality across western Queensland (Figure 7.4). The lack of robust correlations between annual rainfall, the SOI, PDO and land erodibility can be attributed to the complex interaction of these factors and lags in the landscape response to climate variability. The length of the data time-series (27 years) also affected the results. In particular, interpretations of the influence of inter-decadal variations in the PDO should be treated with caution.

Results show a weak correlation between land erodibility and annual rainfall across the study area (Table 7.1). Importantly, land erodibility is responsive to multi-year (>2 years) rainfall deficiencies (drought) and periods of above average rainfall. Similar responses have been reported in dust source areas in the African Sahel and North America (e.g. Brooks and Legrand, 2000; Prospero and Lamb, 2003; Reheis, 2006). Significant increases in the areas of land susceptible to wind erosion occur in drought years (Figures 7.5 and 7.6). The landscape response to drought varied between bioregions and is dependent on antecedent rainfall and vegetation conditions, which generates the lag responses in land erodibility change (also Peters and Eve, 1995). The modelled increases in land erodibility with drought are consistent with reports of increased wind erosion activity over the study area, e.g. McTainsh et al. (1989), and a global dependence of temporal variations in wind erosion on episodic droughts (Middleton, 1985; Goudie and Middleton, 1992; Gao et al., 2003).

The spatial extent of drought in western Queensland is dependent on the rainfall relationship with ENSO (Crimp and Day, 2003). Temporal patterns in areas affected by drought are not, however, consistent and may vary considerably from year-to-year (McKeon et al., 2004). The poor correlation between modelled land erodibility and the SOI (Table 7.1) is a result of this phenomenon. The El Niño/Southern Oscillation, represented in this study by the SOI, is a fluctuation in the intensity and position of the Walker circulation (Troup, 1965). Sustained negative SOI phases (El Niño) are associated with extended periods of warm sea-surface temperatures (SST) in the equatorial eastern Pacific Ocean, a weakening of the Walker circulation and reduced convection over the Australian continent. Positive SOI phases (La Niña) are associated with a strengthening of the Walker circulation and easterly trade winds and may result in enhanced convection and rainfall over parts of eastern Australia (Sturman and Tapper, 2001). The association of peaks in land erodibility over the study area during El Niño driven drought events suggests that despite the poor correlation ENSO plays an important role in modulating land erodibility dynamics in western Queensland.

The interaction of ENSO with the PDO adds complexity to understanding rainfall and drought variability in western Queensland. Power et al. (1999) reported on the inter-decadal modulation of ENSO and its effects on rainfall in Australia. Their results showed that warm (positive) and cool (negative) phases of the PDO may enhance or suppress positive and negative SOI phases and the probability of eastern Australia receiving above or below average rainfall. McKeon et al. (2004) reported that less than 10% of years (1890-1991) with

a combined negative SOI and cool PDO exceeded median annual rainfall in the Mulga Lands and Strzelecki Desert regions of southwest Queensland. Conversely, during positive SOI-cool PDO phases >70% of years in the analysis period received above average rainfall over the entire study area. For the period of the current study (1980-2006) the PDO was consistently in a warm phase. Based on the results of McKeon et al. (2004), the impact of this on rainfall may have been in enhancing drought over the study area during El Niño events and increasing rainfall in the Mulga Lands during La Niña events. The implications of this in terms of land erodibility change are difficult to surmise. Table 6.1 suggests that in the Mulga Lands at least, increases in land erodibility are related to periods of decreasing rainfall, negative SOI and warm PDO. The weaker correlation between rainfall and the SOI in southern and western Australia (Pittock, 1975) suggests that the relationship between land erodibility and ENSO is likely to be weaker outside the current study area. Extending the length and coverage of the model simulation back to the 1950s would provide an opportunity to assess a larger combination of land erodibility, rainfall variability, ENSO and PDO conditions across Australia and would improve our ability to quantify the nature of their interactions in other bioregions.

It is conceivable that additional climate oscillations that influence rainfall over western Queensland will affect the erodibility of the landscape. Such oscillations include the Madden Julian Oscillation (MJO) that has been shown to affect rainfall over northern Australia on a 30-60 day cycle (Donald et al., 2006), and at inter-annual time scales teleconnections like the Indian Ocean Dipole (IOD) (Saji et al., 1999). The IOD has been shown to operate independently of ENSO with a 2 year periodicity (Ashok et al., 2003a; Behera and Yamagata, 2003), and has been found to have a significant impact on rainfall in western and southern Australia (Ashok et al., 2003b). Determining the influence of teleconnections like the MJO and IOD on rainfall and land erodibility in western Queensland requires that their effects can be separated from those of ENSO and the PDO. Globally, the significance of these and other teleconnections is likely to vary considerably between continents (Ginoux et al., 2004). Analysis of land erodibility-climate interactions at higher temporal resolutions, e.g. monthly, is necessary to achieve this but was beyond the scope of the current research. This is because model simulation accuracy at short (monthly) time scales is affected by the lack of a robust scheme to predict temporal changes in soil erodibility (Webb et al., 2009; Chapter 5).

Finally, the sensitivity of the bioregions to land erodibility change can be described. The erodibility of the Mulga Lands is most sensitive to climate variability (Figure 7.5). This is supported by the correlation of land erodibility with rainfall, the SOI and the PDO, and historical reports of land degradation in the bioregion in response to drought and overgrazing in the 1960s, 1970s and early 1980s (McKeon et al., 2004). The sensitivity of the bioregion to degradation can be attributed to the region's arid climate, high stocking rates and the fragmented nature of the landscape (Stokes et al., 2008). These factors contribute to low vegetation resilience to short-term climate variability and a response of significant reductions in understory grass cover during periods of low rainfall. This sensitivity exists in other similarly fragmented semi-arid rangelands around the world (Galvin et al., 2008), suggesting that other dust source areas will be as responsive to global teleconnections.

Figure 7.5 indicates that land erodibility dynamics in the Channel Country operates at a similar time scale to the Mulga Lands, but that rapid (<1 year) increases in high erodibility land do not occur as frequently. Land erodibility in the Mitchell Grass Downs and Simpson-Strzelecki bioregions has a lower sensitivity to inter-annual rainfall variability. This is indicated by the low net fluctuations in erodible land over the bioregions for much of the period from 1980-2006. Interestingly, these bioregions do appear to be sensitive to high magnitude changes in regional climate affecting rainfall. In the Mitchell Grass Downs this sensitivity is indicated by the high cover of erodible land (up to 60%) during the intense 1982 and 1984 droughts, and the near complete reduction in erodible land cover change (maximum 20%) between 1990 and 2006. In the Simpson-Strzelecki bioregion this sensitivity is indicated by the reduction (to <1%) in highly erodible land under the regionally wet 1998 and 2000 La Niña events. A genetic ability of the dominant Spinifex vegetation (*Triodia* spp.) to survive under an arid climate likely contributes to the low net variation in the erodibility in the Simpson Desert except under extreme environmental change (Specht and Specht, 1999).

Regional scale field surveys to monitor temporal changes in the susceptibility of these bioregions to wind erosion and extended (>30 years) simulations of land erodibility are required to elucidate more specific characteristics of the sensitivity of western Queensland to climate variability and management.

7.7 Conclusions

This research has examined spatial and temporal patterns in land erodibility in western Queensland, Australia. The distribution of erodible land areas has been mapped, and temporal changes in land erodibility have been analysed in the context of regional and global scale climate variability. Consistently erodible land areas were found to be located in dunefields and the outer floodplains of the regions' major river systems. These spatial patterns in land erodibility are consistent with observational records of wind erosion activity. Temporal trends in land erodibility were found to be distinct across the four study area bioregions. Regional scale land erodibility dynamics were found to be influenced by vegetation cover sensitivity to rainfall, ENSO and PDO interactions. In a global context the research has highlighted the complex and dynamic nature of land erodibility.

Model predictions of climate change indicate increasing variability in precipitation across the world's deserts (Meehl et al., 2007). Concurrently, global measurements of outgoing long-wave radiation indicate a poleward expansion of the Hadley circulation (Hu and Fu, 2007). These may lead to increased frequencies of drought and the expansion of mid-latitude deserts in the northern and southern hemispheres. The results presented in this chapter demonstrate that regional scale changes in land erodibility are sensitive to rainfall variability driven by global scale climate teleconnections. This emphasises the importance of research to map and monitor land erodibility so that we can better understand the effects of land management and future climatic changes on wind erosion processes.

Further research into land erodibility dynamics in Australia requires an extension of that presented here. Firstly, research should focus on extending the length of the model simulations. This would allow variations in land erodibility to be analysed under a greater range of ENSO and PDO conditions. Secondly, effort should be directed toward developing schemes to simulate temporal changes in soil erodibility in rangeland environments. Land erodibility dynamics over small (sub-regional) spatial and monthly temporal scales are highly dependent on soil erodibility, particularly in landscapes with sparse vegetation cover. The addition of a robust soil erodibility scheme to AUSLEM would improve model skill in assessing land erodibility. It would also allow for analyses of model output at higher temporal resolutions and an ability to assess the impacts of short- and long-term land management practices on potential wind erosion activity. Finally, analyses should be extended to include

the southern and western regions of Australia over both rangelands and cultivated environments as these regions are likely to experience significant change in response to future climate change (Meehl et al., 2007).

Chapter 8

Conclusions

This chapter summarises the outcomes of the research and demonstrates how each of the research aims and objectives have been addressed. The main research findings are presented and discussed in the context of how the thesis has contributed to our knowledge of the controls of wind erosion processes. The chapter concludes with a summary of the limitations of the research and a statement of future research priorities.

8.1 Problem Statement and Research Aims

This thesis has presented new information that successfully addresses some major deficiencies in our knowledge of wind erosion processes in Australia. The research has also developed new methods for monitoring and modelling spatial and temporal variations in land susceptibility to wind erosion that have global application. Chapter 1 outlined the rationale for conducting the research in this thesis. This included a statement of research problems relevant to developing our understanding of land erodibility dynamics:

- There is a poor knowledge of which areas of western Queensland, Australia, are susceptible to wind erosion. This is a significant problem considering that Australia contains the dominant dust source area in the southern hemisphere – the Lake Eyre Basin.
- There is a lack of research into spatial and temporal patterns in land erodibility at the landscape scale. Research into land erodibility at this scale is essential if we are to better link field scale wind erosion processes to regional dust emission and transport processes.
- We have a poor knowledge of how soil and land erodibility respond to climate variability and land management, particularly in rangeland environments which cover ~45% of the world's land surface.
- There is a lack of quantitative models to predict temporal changes in soil erodibility to wind. Soil erodibility is a fundamental control on wind erosion and so this issue affects any research that seeks to model wind erosion processes.

- There is a growing requirement to learn more about the sensitivity of rangelands to climate variability and land management pressures in light of uncertain future climate change. Assessing the landscape susceptibility to land degradation processes like wind erosion is an essential component of this research.

Five research aims were set to address these deficiencies. The research aims focus on developing monitoring and modelling methods that can be used to assess spatial and temporal patterns in land erodibility. They were:

1. To develop a framework for modelling temporal changes in soil erodibility in response to climate variability and land management pressures.
2. To develop AUSLEM into a functional model to assess land susceptibility to wind erosion, i.e. land erodibility, across western Queensland, Australia.
3. To validate the performance of the land erodibility model.
4. To map the spatial extent of areas susceptible to wind erosion in western Queensland.
5. To identify the role of climate variability in determining spatial and temporal patterns in land erodibility dynamics in western Queensland.

The following section describes outcomes of the research that address these aims and the research objectives.

8.2 Research Findings

The first aim of this thesis was to develop a framework for modelling temporal changes in soil erodibility. This aim was achieved by addressing three research objectives. The first objective was to provide a systems analysis of factors controlling soil and land susceptibility to wind erosion. This was presented as a literature review in Chapter 2. The systems analysis concluded with the presentation of a conceptual model of the land erodibility continuum. The conceptual model was used as the basis for the second research objective: to present a review of methods for modelling soil and land susceptibility to wind erosion as incorporated within current wind erosion modelling systems. This review was presented as Chapter 3 of this thesis. The review provided a summary of approaches for modelling land erodibility from the field to regional and global scales. It identified approaches for integrating wind erosion

controls reviewed in Chapter 2, and provided a synthesis of the limitations to the models. The reviews indicated a significant lack of research into quantitative modelling of spatial and temporal variations in soil erodibility, and application of models to assess spatial and temporal patterns in controls on wind erosion at the landscape scale.

The third research objective was to address these deficiencies by developing a framework for modelling temporal changes in soil susceptibility to wind erosion. The model framework was presented in Chapter 4. First, a conceptual model of the soil erodibility continuum was developed relating soil textural characteristics, aggregation and surface crust conditions to wind erosion rates. The conceptual model is based on empirical research from Australia and the United States reviewed in Chapter 2. A framework for modelling temporal changes in soil erodibility within the continuum was then presented. The model framework sought to characterise the nature and timing of changes in soil erodibility in response to climate variability and land management pressures. Parameterisation and integration of the model into AUSLEM were restricted by a lack of quantitative research into soil aggregation and surface crust responses to climate variability and land management. This meant that an alternate approach to accounting for soil erodibility was required in modelling land erodibility in Chapters 5 to 7. To conclude the research, a synthesis of published work examining controls on soil erodibility was presented. The synthesis highlighted gaps in our understanding of factors driving soil erodibility dynamics and identified future research priorities to parameterise the model and advance soil erodibility modelling.

The second aim of this thesis was to develop AUSLEM into a functional model to assess land erodibility across western Queensland, Australia. This aim was achieved through the fourth research objective. Development of the model was presented in Chapter 5 and published in Webb et al. (2009). The model was developed from process relationships identified in the systems analysis in Chapter 2, and built upon a land erodibility model, AUSLEM, developed by Webb et al. (2006). The original rule-based structure of AUSLEM was replaced with an empirical model that addressed the requirements for modelling land erodibility through a continuum, and at the landscape scale. AUSLEM was designed in a GIS environment and can be run to assess spatial patterns in land erodibility at a 5 x 5 km spatial resolution on a daily time-step across western Queensland. Temporal changes in soil erodibility were accommodated in the model by restricting the output analyses to monthly temporal resolutions at landscape to regional spatial scales.

The third aim of this thesis was to validate the performance of the land erodibility model. This was achieved through the fifth and sixth research objectives: to develop a method for visually assessing land erodibility at the landscape scale; and to employ the visual assessments of land erodibility and observational records of wind erosion activity to validate the model.

The first approach for model validation drew on a comparison of time-series model assessments of land erodibility with observational records of wind erosion activity across western Queensland (Chapter 5). The observational data included records of locally blowing dust, dust storms, dust hazes, and dust whirls. The validation was conducted at eight meteorological stations located across the four bioregions covering the study area. A cross-correlation approach was used to examine the synchronisation of trends in modelled land erodibility, annual total dust-event frequencies, and mean annual 3 pm wind speeds at four spatial length scales (from 25 to 150 km) between 1980 and 1990. The model output had strong correlations with dust-event frequencies at half of the validation stations. The model agreement with trends in dust-event frequencies varied across spatial scales and was highly dependent on land type variability around the reference stations. Poor correlations at the other stations were linked to limitations of the model, i.e. the lack of a soil erodibility scheme, and the coarse spatial resolution of the model input data. The validation was also affected by the types of dust events used in the validation process. Dust source areas for dust storms and dust hazes cannot be detected from the observational data and so these event types did not always provide good records of wind erosion activity by which model performance could be assessed.

The second approach for model validation relates to the fifth objective of this thesis. That was, to develop a method for visually assessing land erodibility that can be used to monitor rangeland conditions at the landscape scale (Chapter 6). This objective was addressed by establishing criteria for evaluating land erodibility based on empirical relationships between soil texture, vegetation cover, land type characteristics, and wind erosion. The criteria were then used to visually assess land erodibility over long distances (10^3 km) on vehicle-based transects run through the western Queensland rangelands. To validate the model, data collected on the transects were compared with model assessments of land erodibility at a 200 x 200 m spatial resolution in five sub-sections of the study area. The comparison indicated that AUSLEM performs better in the open grasslands of the western Mitchell Grass Downs

and Channel Country than in the fragmented woodlands of the Mulga Lands. The result highlighted the importance of wind erosion models to be able to account for the spatial distribution of vegetation in environments with heterogeneous cover.

The fourth aim of this thesis was to map the extent of areas susceptible to wind erosion in western Queensland. This aim was achieved by addressing the seventh research objective. That was, to apply the model developed in Chapter 5 to assess spatial patterns in land erodibility across western Queensland. Spatial patterns in land erodibility were reported in Chapter 7 from a 27-year simulation from 1980 to 2006. Three regions of the study area were found to have consistently high land erodibility. These include the south-western Mulga Lands, Strzelecki Desert, and western side of the Channel Country. Areas that frequently have moderate land erodibility extend from the high erodibility regions. These areas expand across the driest parts of the study area, covering the western and southern portions of the Channel Country, the Strzelecki dunefields and the western half of the Mulga Lands. A belt of moderately erodible land also frequently extends north-east from the Channel Country into the Mitchell Grass Downs near Longreach. Land areas with consistently low and no erodibility occur across the northern and eastern boundaries of the study area in the Mitchell Grass Downs and Mulga Lands. The landforms associated with erodible land, i.e. river floodplains and dunefields, are consistent with those in other dust source areas in Africa, the Middle East, and China. The spatial patterns in land erodibility were found to match the spatial pattern of dust-storm frequencies in western Queensland. The coarse model input data resolution meant that AUSLEM could not accurately represent erodible land in the Simpson Desert bioregion.

The final aim of this thesis was to identify the role of climate variability in determining spatial and temporal patterns in land erodibility in western Queensland. This aim was achieved by addressing the eighth research objective. That was, to analyse spatial and temporal patterns in modelled land erodibility in the context of regional scale rainfall and teleconnections driving climate variability across western Queensland. Results were presented in Chapter 7 and are based on the 27-year simulation of land erodibility (1980-2006).

First, seasonal patterns in land erodibility were examined. The research found that land erodibility in western Queensland has weak seasonality. The strength of seasonal variations

was shown to change across the study area and between land types. There is a north-south spatial pattern in the seasonality of land erodibility across western Queensland. Land erodibility reaches a maximum in spring (SON) and early summer (ND) in the Channel Country and Mitchell Grass Downs, and a minimum in late summer (JFM) through to winter (JJA). In the Mulga Lands, land erodibility peaks in spring (SON) and autumn (MAM), and reaches a minimum over winter (JJA). These dynamics are consistent with seasonal patterns in dust-storm frequencies in eastern Australia.

At inter-annual time scales, land erodibility in western Queensland had a weak correlation with rainfall, ENSO and the PDO. The research found that land erodibility is responsive to multi-year (>2 years) rainfall deficiencies (drought) and periods of above average rainfall. The lack of robust correlations between annual rainfall, ENSO, the PDO and land erodibility can be attributed to the complex interaction of these factors and lags in the landscape response to climate variability. The strongest correlation between land erodibility, rainfall and the teleconnections was found in the Mulga Lands. This is the most sensitive bioregion in western Queensland to climate variability due to frequent drought, high stocking rates, and fragmentation of the landscape due to small property sizes. Temporal changes in land erodibility in the Channel Country and Mulga Lands were found to occur at similar time scales. In the Mitchell Grass Downs, however, rapid (<1 year) increases in high erodibility land at the regional scale do not occur frequently. Land erodibility in the Mitchell Grass Downs and Simpson-Strzelecki bioregions was found to have the lowest sensitivity to inter-annual rainfall variability. The research has shown that these bioregions are more sensitive to disturbance by fire and high magnitude changes in regional climate affecting rainfall.

8.3 Contribution of the Research

The contribution of this research and implications of the research outcomes can be described in terms of the problem statements established in Chapter 1 and repeated in Section 8.1.

1. There is a poor knowledge of which areas of Australia are susceptible to wind erosion. This is a significant problem considering that Australia contains the dominant dust source area in the southern hemisphere – the Lake Eyre Basin.

The research presented in this thesis has provided a quantitative analysis of spatial and temporal patterns in land erodibility in the western Queensland rangelands (Chapter 7). Output from AUSLEM was used to define the frequency at which land in the four bioregions of western Queensland is susceptible to wind erosion. The analysis of a 27-year output time-series provided quantitative information on landscape to regional scale land erodibility dynamics at monthly and inter-annual time scales. A methodology was also presented for conducting field assessments of land erodibility in remote rangeland environments (Chapter 6). The model development, field assessment methodology and subsequent analyses have made a significant contribution to our understanding of spatio-temporal patterns in land erodibility in western Queensland.

2. There is a lack of research into spatial and temporal patterns in land erodibility at the landscape scale. Research into land erodibility at this scale is essential if we are to better link field scale wind erosion processes to regional dust emission and transport processes.

This thesis has presented a new model specifically designed to assess spatial and temporal patterns in land erodibility at the landscape scale. The thesis has also presented a new methodology for assessing land erodibility in the field that can be used to monitor temporal changes in land erodibility. Both the model framework and field assessment methodology have potential applications outside Australia. Results from the model application (Chapter 7) highlight the role of global scale climate teleconnections in controlling land erodibility dynamics. This outcome emphasises and strengthens the need for further research of this type to be conducted in other dryland environments.

3. We have a poor knowledge of how soil and land erodibility respond to climate variability and land management, particularly in rangeland environments which cover ~45% of the world's land surface.

The research presented in this thesis has provided a quantitative analysis of the relationships between land erodibility, rainfall and teleconnections including the El Niño/Southern Oscillation and Pacific (inter-) Decadal Oscillation (Chapter 7). The research has also presented a framework for modelling relationships between soil erodibility, climate variability and land management pressures (Chapter 4). The research was conducted in the western Queensland rangelands, and this area has a similar climate, vegetation structure and

land use characteristics to the rangelands in India, Africa and North America. The complex interactions found between climate and land erodibility dynamics in the four bioregions of western Queensland highlight the importance of pursuing global research into land erodibility in other rangelands environments. The research did not seek to quantify relationships between land management and model assessments of land erodibility. This was because research is first required to better quantify climate-land erodibility interactions, which is dependent on the ability to model temporal changes in soil erodibility. Parameterisation of soil erodibility models must therefore be a priority for future research.

4. There is a lack of quantitative models to predict temporal changes in soil erodibility to wind. Soil erodibility is a fundamental control on wind erosion and so this issue affects any research that seeks to model wind erosion processes.

This thesis has presented a framework for modelling temporal changes in soil erodibility. The framework draws on relationships between soil aggregation, surface crusting, climate variability and land management pressures (Chapter 4). Application of the model was restricted by the lack of quantitative research into soil erodibility-climate-management interactions. The thesis has provided a review of current research in this area and has presented a summary of future research priorities that will enable parameterisation of the model and its integration into existing models to assess wind erosion processes.

5. There is a growing requirement to learn more about the sensitivity of rangelands to climate variability and land management pressures in light of uncertain future climate change. Assessing the landscape susceptibility to land degradation processes like wind erosion is an essential component of this research.

This thesis has made a significant contribution to the body of knowledge about wind erosion in rangeland environments. This contribution includes the development of models and methods to assess soil and land erodibility, and the application of these to generate new information on land erodibility dynamics in western Queensland. The research has provided new information on the sensitivity of the western Queensland rangelands to potential land degradation as influenced by climate variability. The information generated from the research relates to processes at the landscape to regional scales. Outcomes of the research therefore

have direct relevance to policy and management decision making which are implemented across these scales in rangeland environments.

The research presented in this thesis contributes toward modelling and understanding regional scale wind erosion processes both in Australia, and internationally. In the context of the wind erosion models reviewed in Chapter 3, this research has developed a new tool specifically aimed at assessing land susceptibility to wind erosion at the landscape (10^3 km^2) to regional ($>10^4 \text{ km}^2$) scales. This capacity fills a gap in wind erosion modelling between these scales, and a lack of research to assess spatial and temporal variations in land erodibility in rangeland environments. In terms of model functionality and scale, AUSLEM is now positioned between the highly empirical field-scale models like RWEQ, WEPS and TEAM, and the deterministic regional to continental scale DPM and WEAM models (Figure 3.1).

The AUSLEM framework differs from other regional scale wind erosion models such as DPM (Marticorena and Bergametti, 1995) and IWEMS (Lu and Shao, 2001) in that it operates at a higher spatial resolution ($5 \times 5 \text{ km}$), and using Aussie GRASS inputs could be used to assess land erodibility back to January 1890. As shown in Chapter 7, this application potential provides a powerful tool for examining land erodibility responses to climate variability and land management factors. AUSLEM suffers a similar limitation to other wind erosion models in that it does not account for short-term temporal variations in soil erodibility. Nevertheless, this research has produced a framework (Chapter 4) for modelling dynamic changes in soil erodibility that may be applicable not just to AUSLEM, but through other wind erosion modelling systems. Pursuing the development of soil erodibility models must be a priority in wind erosion research as few models currently contain such schemes, and those that do have a high dependence on field-measured input data (e.g. Hagen, 1991; Fryrear et al., 1998; Gregory et al., 2004).

Application of AUSLEM has demonstrated how land erodibility in western Queensland is dynamic and responsive to climate variability at the landscape to regional scales. These research outcomes complement the modelling results of Shao and Leslie (1997), Lu (1999) and Lu and Shao (2001), who have shown the importance of land erodibility dynamics in modelling dust emission processes in Australia. While AUSLEM and IWEMS have different model schemes to account for vegetation, soil moisture and soil textural effects on the erodibility of the land surface (Chapter 3), the underlying conceptual frameworks of the

models are not dissimilar. Maps of u_{*t} and land erodibility modelled by Lu (1999) and AUSLEM for western Queensland show similarities across the Channel Country. It would be constructive to compare the assessments in cross-model validation exercises. The methodology for collecting field surveys of land erodibility described in Chapter 6 would support these comparisons and provide additional data for validating models like IWEMS.

Finally, this research has clearly demonstrated the requirement for further studies to assess spatial and temporal patterns in land susceptibility to wind erosion. Dynamic changes in land erodibility like those identified by AUSLEM must be considered in the context of modelling regional to global scale dust emission and transport (e.g. Mahowald et al., 2003a). Grini et al. (2005) sought to identify suitable methods for characterising erodibility patterns in global dust source areas. The frameworks for modelling soil and land erodibility presented in this thesis support that research, and there is significant potential for integrating the methods and findings of this research into such studies.

8.4 Research Limitations

The major limitations of this research relate to three main areas. They are: 1) the availability of robust models to predict temporal changes in soil erodibility, and the lack of quantitative data with which to parameterise such models; 2) spatial scale effects on model performance in heterogeneous landscapes; and 3) the availability of data suitable for model validation.

1. The accuracy of AUSLEM at spatial scales less than $\sim 50 \text{ km}^2$ was severely affected by the absence of a robust scheme to predict temporal changes in soil erodibility. This issue became evident when analysing trends in the model output at locations with similar vegetation cover but different soil textural attributes (Chapter 5). Analysis of the model output was therefore restricted to the landscape (10^3 km^2) to regional (10^4 km^2) scales. While AUSLEM output accuracy at smaller scales is affected by the accuracy of its Aussie GRASS inputs, the addition of a soil erodibility scheme is likely to significantly improve model performance. It would also allow for the analysis of model output at time scales less than one month, at which variations in soil erodibility are an important control on land erodibility dynamics.

2. Spatial scaling issues were found to adversely affect both model performance and efforts to validate the model output using field assessments of land erodibility. This issue was found to be particularly relevant to the performance of AUSLEM in the Mulga Lands, to the east of the study area. In this bioregion the distribution of tree cover is a dominant control on land erodibility. Neither the model input data nor model framework accommodated sub-grid scale variations in vegetation cover. This meant that model predictions of land erodibility were largely driven by the model tree cover threshold (of 20%) and accuracy of the 30 x 30 m resolution foliage projective cover (FPC) data in representing the distribution and cover of woody vegetation. Field observations suggest that AUSLEM was unable to detect local variations in land erodibility in this bioregion. This became a significant problem when comparing field assessments of land erodibility to AUSLEM output for model validation. Because neither the model nor the field assessments specifically accounted for vegetation distribution effects on erodibility, a good measure of model performance could not be obtained in the Mulga Lands. Further research is required to incorporate vegetation distribution effects into the model.
3. The lack of quantitative estimates of land susceptibility to wind erosion affected efforts to validate AUSLEM. Model validation was therefore dependent on a comparison of time-series trends in model output with observational records of dust events. While the comparison demonstrated that the model works well in the western portion of the study region, it did so because the dust event records at the meteorological stations in that area were representative of local dust entrainment. The records at the western stations therefore provided a good indicator of temporal changes in land erodibility around the stations. Poor agreement between the model output and dust event records at the eastern stations was attributed to the fact that the dust events there did not provide a good indicator of local conditions. Thus, model performance to the east of the study area could not reasonably be assessed using that methodology. While an approach was developed to collect field assessments of land erodibility across the study area, application of the data in validating AUSLEM was affected by the availability of data that could be used as input to the model at an appropriate spatial resolution for comparison with the field assessments. It was also affected by the fact that the field observations were made on a categorical basis and then compared to a continuous and non-linear range of model output values. This research limitation can be addressed by: refining the field assessment methodology so that visual observations are recorded on a continuous scale; and making

multiple assessments of the model performance across scales and under a range of climatic conditions (i.e. during drought and periods of above average rainfall).

8.5 Future Research Priorities

This thesis has identified several areas that require future research. These relate to the limitations of this research, understanding land erodibility in Australia, and to the development of models to assess spatial and temporal changes in soil and land erodibility in general. They include:

1. More studies are required to quantify relationships between factors controlling soil erodibility (aggregation and surface crusting), soil physical properties, climatic factors and land management. In particular, research is required at high temporal resolutions (e.g. days) to elucidate the response of soils to individual rainfall events and antecedent climate and management conditions. This research is essential for developing functional soil erodibility models, which are lacking from most process-based wind erosion models today. A component of this research should focus on how to integrate new soil erodibility models into current wind erosion and land erodibility modelling systems. For example, the output of soil erodibility models should be physical parameters that can be input into schemes that compute the threshold friction velocity for sediment mobilisation.
2. More attention needs to be given to how the effects of heterogeneous surface roughness can be incorporated into wind erosion and land erodibility models. While statistical methods have been developed to account for sub-grid scale heterogeneity in vegetation cover (Raupach and Lu, 2004), further research is required to quantify surface roughness effects at scales relevant to the field distribution of roughness elements (e.g. Okin, 2008). This will be facilitated by developing methods that utilise remote sensing technologies to map the distribution of vegetation cover at moderate to high spatial resolutions (e.g. pixels < 30 x 30 m). While this may not currently be feasible for dynamic herbaceous cover, estimates of the distribution of tree cover in rangeland environments (which varies over longer time scales) would significantly improve model assessments of potential wind erosion.

3. Following the development of the first two research priorities, the spatial application of AUSLEM should be extended to assess land erodibility in central, western and southern Australia. This would allow for a continent-wide analysis of spatial and temporal patterns in land erodibility. An analysis of this type would significantly increase our understanding of wind erosion processes in Australia. Research outcomes could be integrated into further wind erosion modelling research, enhancing land management policy, and be used to identify regions that require more intensive field monitoring or remediation.
4. The temporal range of AUSLEM simulations should be extended to better resolve the effects of inter-decadal climate oscillations (e.g. the Pacific inter-Decadal Oscillation) on land erodibility dynamics. Integrating a soil erodibility scheme into the model would also allow for higher spatial and temporal resolution studies of land erodibility to be conducted. This would allow for the effects of short-term climate oscillations (e.g. the Madden-Julian Oscillation) on land erodibility to be examined. These studies would provide a basis from which to analyse land management effects on spatial and temporal patterns in land erodibility. It would also provide valuable information on the sensitivity and susceptibility of the rangelands to natural and anthropogenic disturbance and erosion.
5. Additional research is required to refine methods for making field assessments of land erodibility. Modelling studies should not be developed independently of rigorous field experimentation at appropriate spatial and temporal scales. Developing approaches for assessing land erodibility at the landscape scale and over large geographic areas is essential for calibrating and validating model predictions of land erodibility at broad spatial scales. Particular attention should be given to collecting field data at scales that are comparable with output from existing models, and which are based on quantitative and repeatable methodologies.
6. Finally, there is scope for applying existing wind erosion models to assess land erodibility dynamics. Wind erosion models have been developed to operate at a range of spatial and temporal scales, and across a range of application environments, i.e. cultivated lands and rangelands. Application of the models to assess land erodibility would allow for the analysis of land erodibility dynamics outside Australia, and across spatial and temporal scales. Examining output scenarios from different models within a region would provide

the opportunity to build evidence about land erodibility-climate-management interactions. Information generated through this process could also be used as a feedback for refining existing modelling systems and to generate research questions that can be addressed through field scale experimentation.

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