University of Alberta

Overwinter processes and wind erodibility of clay loam soils in southern Alberta

by

Murray S. Bullock ©

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Name of Author: **Murray S. Bullock**

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Murray Bullock

511 East 200 South
Burley, Idaho
83318

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"13. For it is expedient that I, the Lord, should make every man accountable, as a steward over earthly blessings, which I have made and prepared for my creatures.
17. For the earth is full and there is enough and to spare; yea, I prepared all things, and have given unto the children of men to be agents unto themselves.
4. And an account of this stewardship will I require of them in the day of judgement."

The Doctrine and Covenants 104:13,17 and 70:4

University of Alberta

Faculty of Graduate Studies and Research

The undersigned certify that they have read, and recommend to the faculty of Graduate Studies and Research for acceptance, a thesis entitled *Overwinter processes and wind erodibility of clay loam soils in southern Alberta* submitted by Murray S. Bullock in partial fulfillment of the requirements for the degree of Doctor of Philosophy in Soil Science.

Yongsheng Feng (Co-supervisor)

R. Cesar Izaurralde (Co-supervisor)

Francis J. Larney

G. Peter Kershaw

David S. Chanasyk

James A. Robertson

Donald J. Pluth for

Keith E. Saxton (external reader)

Date: April 17, 2001

ABSTRACT

Wind erosion is one of the main forms of soil degradation in the semiarid region of the Canadian prairies. Overwinter changes in aggregate size distribution generally increases wind erosion risk. Elucidation of overwinter processes (i.e., freeze-drying, freeze-thaw, blowing snow abrasion) would allow for wind erosion risk assessment and prediction. The objectives of this study were to: 1) determine the effect of freeze-drying on aggregate breakdown; 2) monitor changes in dry aggregate size distribution (DASD) and surface roughness and identify relationships to climatic variables and 3) determine the kinetic energy associated with blowing snow which may abrade soil and increase the erodible fraction.

The freeze-dry study had a laboratory and field component. Laboratory derived predictions estimated 31 freeze-dry cycles for the 0.1 kg kg^{-1} water content aggregates to reach 60% erodible fraction (EF), 9 cycles for the 0.2 kg kg^{-1} aggregates and only 2 for the 0.3 kg kg^{-1} aggregates. In a companion field study, freeze-dry cycles were estimated by calculating large vapour pressure (VPL) and small vapour pressure (VPS) gradients between the soil surface and the atmosphere. Predictions of number of freeze-dry cycles to reach the 60% EF were 60 for VPL and 37 for VPS.

DASD as quantified by EF and geometric mean diameter (GMD) was monitored for three winters on a different site each winter. Surface roughness was monitored at two of the sites. Regression analysis with time revealed a positive linear response for EF and an exponential decay for GMD for all three sites. Surface roughness regression response was an exponential decay over time for both sites. Timing and form of precipitation, which influence soil water content, were deemed the

major climatic factors in overwinter aggregate degradation leading to wind erosion risk.

Using a SENSIT™ meter to measure kinetic energy and particle counts of blowing snow, preliminary results indicated that there is a large potential for abrasive action on soil by blowing snow.

A schematic flowchart summarizing the various processes and possible pathways leading to overwinter changes of aggregate size distributions of bare clay loam soils in southern Alberta is presented.

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Chapter 1

General Introduction

Wind erosion continues to be one of the main forms of soil degradation in the semi arid region of the Canadian prairies (Wall et al., 1995). On-site losses of organic matter and nutrient rich topsoil are the main cause of this degradation resulting in a decrease in soil productivity. There are also off-site impacts of wind erosion which affect air and water quality (Saxton, 1995; Cihacek; 1993; Zobeck and Fryrear, 1986). The majority of these impacts can be attributed to agricultural management systems and climatic effects.

Prediction of potential wind erosion losses for different field and climatic scenarios has been an ongoing challenge. In 1965, the United States Department of Agriculture (USDA) released the wind erosion equation (WEQ; Woodruff and Siddoway, 1965). It is an empirically based equation including only five factors and giving only annual soil losses for wind erosion. Adaptation of WEQ to account for temporal and spatial variability of wind erosion losses has had limited success and WEQ has been deemed mature or outdated technology (Hagen, 1991). Consequently the USDA is making efforts to develop a process based daily time step model called the Wind Erosion Prediction System (WEPS). WEPS is an integration of several submodels including weather generator, erosion, hydrology, crop growth, decomposition, tillage and soil (Hagen, 1991).

Temporal changes in dry aggregate size distribution (DASD) of the surface soil will be included in the soil submodel of WEPS (Zobeck, 1991). The DASD influences the level of wind erosion risk. Put simply, the higher the proportion of small aggregates, the higher the wind erosion risk. DASD is determined by sieving dry soil in a rotary sieve (Chepil, 1952). By measuring the amount of soil passing specified sieve diameters (usually six sieve cuts) the geometric mean diameter (GMD), or diameter at which 50 % of the sample weight passes, can be calculated (Gardner, 1956). Another useful measure, which can be derived from rotary sieve data, is the erodible fraction (EF, % aggregates by weight < 0.84 mm diam.). GMD has a non-linear, inverse relationship with EF. Generally aggregates < 0.84 mm diameter are considered erodible and soils with an erodible fraction > 60% are considered high erosion risk (Chepil, 1942; Anderson and Wenhardt, 1966).

With the advent of the WEPS there is a need to model overwinter processes that might significantly contribute to soil losses by wind erosion. Prediction of temporal soil surface properties, such as DASD, from weather inputs is a major component of WEPS (Hagen, 1991).

Overwinter degradation of soil aggregates increases wind erosion risk, especially after conventional fallow, a common practice in the semiarid Canadian prairies (Larney et al., 1995). The process of soil freezing and thawing contributes to aggregate disruption and hence decreases GMD. In southern Alberta, overwinter changes in GMD for five fallow management systems on a clay loam soil were due to the climatic factors of cumulative snowfall, snow cover days, and degree of freeze-thaw activity (Larney et al., 1994). Increased snowfall and days of snow cover prevented aggregate breakdown while freeze-thaw cycles increased it. In North Dakota, Merrill et al. (1995) found significant relationships between the DASD and climatic factors such as number of snow cover days, number of freeze-thaw cycle days with no snow cover, and fall precipitation. From fall to spring, snow cover and fall precipitation increased GMD while the number of freeze-thaw cycles decreased GMD. For soils in Kansas and Texas, Chepil (1954) reported increases in erodibility from fall to spring for five soils ranging in texture from fine sandy loam to clay, with the greatest increases occurring on the finer-textured soils. Erodibility increases were attributed to the freezing of moist soil during winter, which caused the expansion of ice crystals within aggregates and subsequent shattering. The force exerted by freezing water in confined spaces (eg. soil pores) has been measured at 1465 Mg m^{-2} (Brady, 1990).

In addition to freeze-thaw effects, freeze-drying also plays an important role in aggregate breakdown. Overwinter studies on a clay loam soil in Saskatchewan showed that bare soil surfaces exposed to freeze-thaw and freeze-dry cycles had greater aggregate breakdown than those protected by snow cover (Anderson and Bisal, 1969). In addition, exposed soils that were wetted at the start of the study were more vulnerable than dry exposed soils. Hinman and Bisal (1968) also found that freeze-drying caused increased disruption of aggregates as soil wetness at time of freezing increased. However, if these frozen aggregates were allowed to thaw before drying (i.e. evaporative-drying *vs.* freeze-drying) there was substantially less or no change in erodibility. Staricka and Benoit (1995) found similar relationships between moisture content of aggregates at time of freezing and freeze-drying. In addition there was also a positive relationship between aggregate size and freeze-dry disruption but freeze-thaw frequency and soil type had less effect. De Jong and Kachanoski (1988) concluded, after a study on soil freeze-drying in Saskatchewan, that the climatic conditions in the chinook belt of southern Alberta are highly conducive to freeze drying and commented that the frequent occurrence of dust storms in this region is not surprising.

Abrasion by blowing snow is another overwinter process that can contribute to aggregate breakdown (Bullock et al., 1992). Pomeroy et al. (1993) in modeling the transport and sublimation of snow, considered the saltation component of blowing snow. An earlier snow sublimation model developer noted that snowflakes become spherical shapes similar to sand as they are moved by wind and abraded (Schmidt, 1972).

These overwinter climatic and snow layer factors are prevalent in the chinook belt region in southern Alberta. Air temperatures can change as much as 40°C within one hour of the onset or ending of chinook conditions. There can be as many as 100 freeze-thaw cycles during the winter season because of temperature fluctuations caused by chinooks. In addition, the strong, dry winds (100-120 km hr^{-1}) associated with chinook conditions will either sublimate snow or blow it into sheltered areas if temperatures are below 0°C. At temperatures above 0°C, chinook winds cause rapid evaporation and drying of the soil surface. The exposed soil is then vulnerable to freeze-thaw and freeze-dry processes. Chinook winds should initiate and promote a constant vapour pressure gradient for this evaporative- and freeze-drying to occur.

Regional cropping management systems also contribute to wind erosion risk. On dryland farms, the practice of mechanical fallowing for conservation of moisture leaves soil exposed to the elements as crop residues are buried and more rapidly decomposed. In addition, irrigation management systems for growing specialty row crops such as potatoes (*Solanum tuberosum* L.), sugar beets (*Beta vulgaris* L.) and pulses (*Faba* spp., *Pisum* spp) encourages little or no residue conservation. Acreages of these crops are increasing in the irrigation districts of southern Alberta. Fall plowing and land preparation for these crops is preferred to reduce spring work loads leaving fields vulnerable to overwinter erosion risk.

The objective of this study was to further elucidate, in a temporal manner, overwinter processes affecting changes in DASD and hence wind erodibility. This would provide a framework towards prediction and quantification of wind erosion risk in southern Alberta. This dissertation discusses the results from three separate but related studies conducted during the three winter seasons (approximately September to May) of 1992-93, 1993-94 and 1994-95 on sites near or on the Agriculture and Agri-Food Canada Research Centre, Lethbridge, Alberta.

Chapter 2, 'Freeze-drying processes and wind erodibility of a clay loam soil in southern Alberta', verifies DASD changes of freeze-dried soil in a laboratory study and then estimates freeze-drying conditions and their relationships to DASD changes in the field.

Chapter 3, 'Overwinter changes in wind erodibility of clay loam soils in southern Alberta', reports results from a three year study of overwinter temporal changes in DASD and surface roughness.

Chapter 4, 'Preliminary results of blowing snow and wind erodibility of clay loam soils in southern Alberta', presents results of blowing snow energy measurements and particle counts at a wind erosion monitoring site and compares them with similar data for blowing soil.

Chapter 5, 'Synthesis', reviews the pertinent findings of the study and uses them to present a conceptual flowchart of the major pathways for overwinter DASD change. It also lists some recommendations for future research directions.

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Chapter 2

Freeze-drying processes and wind erodibility of a clay loam soil in southern Alberta[1]

2.1 INTRODUCTION

Wind erosion is a major contributing factor to soil degradation on the Canadian prairies (Wall et al., 1995). Recently, the effects of wind erosion on offsite air quality have become a concern (Saxton, 1995). In northern climates, overwinter processes, such as wetting and drying, freezing and thawing, and freeze-drying influence soil aggregate size distribution and wind erosion risk (Larney et al., 1994). Overwinter studies on a clay loam soil in southwestern Saskatchewan showed greater increases in the <1 mm diameter soil fraction for exposed bare soil compared with surfaces insulated by snow or plastic (Anderson and Bisal, 1969). In addition, exposed soils wetted prior to freezing were more vulnerable to erosion than dry exposed soils.

Of the overwinter processes affecting soil aggregate breakdown and wind erodibility, freeze-drying has been documented in least detail. Freeze-drying occurs when the soil is exposed to a combination of the right climatic conditions: frozen soil surface, air temperatures of ~ 0°C, and high wind speeds (Aguirre-Puente and Sukwal, 1984; Lambrinos et al., 1987; de Jong and Kachanoski, 1988). High wind speeds reduce the boundary layer thickness and increase the drying potential (de Jong and Kachanoski, 1988).

Few field estimates of the impact of freeze-drying on aggregate size changes and wind erodibility have been reported (Anderson and Bisal, 1969). Most of the

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information on aggregate size changes due to freeze-drying have been derived from laboratory studies (Hinman and Bisal, 1968; Staricka and Benoit, 1995). These studies have shown that soil water content at the time of freezing is an important factor influencing soil disruption on freeze-drying. Hinman and Bisal (1968) found a positive relationship between soil wetness at freezing and aggregate disruption on freeze-drying. However, they noted that if frozen aggregates were allowed to thaw before freeze-drying there was substantially less, or no change in their susceptibility to wind erosion. Staricka and Benoit (1995) reported similar relationships between antecedent aggregate wetness and subsequent breakdown on freeze-drying. Additionally, they found that the amount of aggregate disruption was more related to aggregate size than freeze-thaw frequency and soil type. The method of drying after freezing, whether by sublimation or evaporation, is also important in increasing the erosive size fraction. This may explain the erratic changes in aggregate size distributions over monthly and yearly periods in southwestern Saskatchewan (Bisal and Ferguson, 1968; Bisal and Nielsen, 1964, 1967).

Climatic conditions in the chinook belt of southern Alberta are conducive to freeze-drying (Bisal and Nielsen, 1964). For freeze-drying to occur, at or near the soil surface, energy is needed to change ice to the vapour phase. In addition to the energy requirement, the vapour pressure above ice must be greater than that in air. A turbulent air flow over the surface helps mix the air and maintain a vapour pressure gradient (Oke, 1987).

Unfortunately, most research on freeze-drying of soil refers to temperature and relative humidity as separate entities rather than combining them to calculate the partial pressure of water vapour (Huang and Aughenbaugh, 1987; Lambrinos et al., 1987; McKenna Neumann, 1990). Relative humidity is the ratio of actual water vapour in air to the saturated water vapour at a specific temperature. Because soil surface temperatures and air temperatures often differ, the assumption that relative humidity reflects the maximum gradient in water vapour between the soil and air may be invalid. Therefore, using vapour pressure gradient as the driving force for freeze-drying is more appropriate than using relative humidity and temperature independently.

While the overwinter processes affecting wind erodibility (wetting and drying, freezing and thawing, freeze-drying) are inextricably linked, this paper attempts to isolate the freeze-drying effect on aggregate breakdown and wind erosion risk. A laboratory study examined the effect of soil water content on the production of wind erodible aggregates (<0.86 mm diameter) or erodible fraction (EF) after repeated freeze-dry cycles. In a companion overwinter field study on a clay loam soil in southern Alberta, the presence of a vapour pressure gradient was used to indicate the occurrence of freeze-drying which in turn was related to EF changes.

2.2 MATERIALS AND METHODS

Laboratory study

Soil samples for the laboratory study were obtained on 26 October 1993 and 11 January and 29 April 1994 from a Dark Brown Chernozemic clay loam soil (organic carbon concentration of 17.2 g kg^{-1} at 0-2.5 cm), 15 km south of the Agriculture and Agri-Food Research Centre at Lethbridge, Alberta (49° 43'N, 112° 48'W). The soil was conventionally fallowed in 1993 with 4 passes of a heavy-duty cultivator (40-cm sweeps) but had been in a zero-tilled spring wheat (*Triticum aestivum* L.)-canola (*Brassica napus* L.)-winter wheat rotation for 8 yr previously. The 26 October 1993 samples were taken immediately after the last fallow tillage operation. On each sampling date, 5 kg of soil was collected with a flat shovel at approximately 10 m intervals along the same 200 m transect for a total of 18 samples. After air-drying, all samples were passed through an improved rotary sieve (Chepil, 1952). The 1.2-2.0 mm aggregate fraction was retained for laboratory analysis which involved three distinct steps: aggregate pre-wetting, freezing for 24 h and freeze-drying for 3 h.

Sub-samples (4 g) of the air-dry (0.02-0.045 kg kg^{-1}) 1.2-2.0 mm aggregates were placed into purpose-built cylindrical sieves (3.8 x 3.8 cm) that had 0.86-mm-diameter stainless screen soldered to the bottom. Chepil (1941, 1942) suggested that aggregates <0.84 mm were wind erodible, hence the sieve size was as close as possible to this value. The aggregates were pre-wet to 0.1, 0.2, or 0.3 kg kg^{-1} water contents with a mist humidifier and subjected to repeated freeze-dry cycles. These water contents were within the range of permanent wilting point and field capacity for the soil (Chang et al., 1990). Wetted aggregates were allowed to equilibrate at room temperature for 18 h.

Before freezing, the sieves were suspended upright in aluminum cans by means of cardboard collars, so that disrupted aggregates passed through the sieves and into the cans. The cans and sieves were placed in an insulated box containing a frozen ice-pack, the box was placed in a large plastic bag to prevent moisture loss, and the bag placed in a -40°C freezer chamber for 24 h.

After removal from the freezer, the samples were kept in the insulated box prior to placement in freeze-dryer flasks. The samples were then freeze-dried at a coil temperature of -45°C using a Freezone 4.5 freeze-dryer (Labconco Inc., Kansas City, MO) with the flasks evacuated to a pressure of at least 13.3 Pa for approximately 3 h.

After freeze-drying, the samples were gently sieved by hand until all disrupted aggregates <0.86 mm passed through to the cans. An erodible fraction (EF) was calculated by dividing the weight of aggregates passing through the sieve by the initial oven-dry weight of the sample. The aggregates retained on the 0.86 mm sieve were subjected to repeated cycles of pre-wetting, freezing and freeze-drying as outlined above. The erodible fraction was calculated after each cycle as the cumulative weight passing through the sieves over successive freeze-dry cycles.

Field Study

The field study was conducted at the Agriculture and Agri-Food Canada Research Centre at Lethbridge, Alberta during the winter of 1994-95. The soil was a Dark Brown Chernozemic clay loam that had been in a spring wheat-fallow rotation since 1955 (Johnston et al., 1995). The selected plots had been fallowed in 1994 with three passes (24 May, 8 July, 30 August) of a one-way disk.

Air temperature and relative humidity sensors were installed on the site at 1 m height on 30 August 1994. A data logger algorithm (Campbell Scientific Ltd., Logan, UT) calculated saturated and actual vapour pressure using 5 s air temperature and relative humidity data, which were subsequently averaged for each hour. Since the actual vapour pressure at the soil surface was not measured, a maximum and a minimum value was estimated. Soil surface temperatures measured with an infrared thermometer (Model 4000A, Everest Interscience Inc., Fullerton, CA) were used to calculate the saturated surface vapour pressure which allowed an estimate of maximum vapour pressure. This maximum surface vapour pressure was based on the assumption that the soil matrix air is saturated under normal field conditions (Hillel, 1982). A minimum surface vapour pressure estimate was made using soil surface temperature and the 1-m ambient air relative humidity. In this latter calculation, the relative humidity in air is assumed to be equal to that at the surface even though the temperatures may not be, thus this represents a minimum surface vapour pressure.

With the assumption that the vapour pressure gradient is the main driving force behind the freeze-drying process, the vapour pressure gradient was determined by subtracting the soil surface vapour pressure from the ambient air vapour pressure. The largest vapour pressure gradient (VPL), was calculated by subtracting the maximum surface vapour pressure estimate from the ambient air vapour pressure. However, since an unknown boundary layer thickness and interstitial ice conditions could exist at the soil surface, the minimum vapour pressure estimate was subtracted from the ambient air vapour pressure allowing the smallest estimate of a vapour pressure gradient to be made (VPS). Realistically, the actual vapour pressure gradient would probably fall within the VPL and VPS extremes but was probably closer to the VPL (dependent on the aggregate size distribution at the surface, and wind conditions, which may increase or decrease the thickness of the boundary layer).

The VPL and VPS boundaries allowed us to test the sensitivity of calculating freeze-dry cycles under various field conditions. A freeze-dry cycle was defined by a negative vapour pressure gradient period followed by a period of non-freeze-drying conditions characterized by snow cover, non-frozen soil or positive vapour pressure gradients. With our logging equipment freeze-drying cycles >1 h duration could be distinguished.

Snow depth was recorded daily at the Lethbridge Research Centre weather station, about 1 km from the study site. Gravimetric soil water content and bulk density (0-2.5 cm depth) were measured twice weekly (depending on snow cover) using a 7.5-cm diameter core. Soil gravimetric water contents were measured in 15-cm increments to a depth of 1.5 m with a truck mounted corer on 22 September 1994.

During the 1994-95 winter, dry aggregate size distribution (DASD) samples were taken with a flat shovel (0-2.5-cm depth, ~5 kg). Sampling interval varied, being dictated by the absence of snow cover. Two 6 x 20-m main plots (located 6 m apart) were each divided into twenty 6 x 1-m sub-plots. At each sampling time, five soil samples were taken from each of the two plots on randomly chosen sub-plots for a total of 10 DASD samples per sampling time. There were a total of 14 sampling times (i.e., 6 subplots not sampled) between 30 August 1994 and 24 May 1995.

All DASD samples were air-dried and passed through an improved rotary sieve (Chepil, 1952) with 38, 12.6, 7.1, 1.9, 1.2, and 0.47 mm diameter sieves. Since a 0.84 mm sieve was not used, EF (% of aggregates <0.84 mm diameter) was estimated from the regression equation of log of sieve size *vs.* cumulative fraction oversize as described by Larney et al. (1994).

Statistical Analyses

For the laboratory study, mean EF and standard errors were calculated for each freeze-dry cycle at each sample date and soil water content (n = 18). The general linear models (GLM) procedure (SAS Institute Inc., 1989) was used to determine differences in EF within the same water content and date ($P < 0.05$). Sample positions (n =18) in the field were considered replicates for this analysis. Least significance difference analysis was performed on the means. Mean EF and standard errors for combined dates at the same soil water content (n = 54) were also calculated. Regression analysis was performed for EF (combined dates at each water content) *vs.* the number of cycles ($P < 0.05$). An "analysis of residual variation" or analysis of variation of residuals (Mead and Curnow, 1983) was performed on EF data from each date for the 0.2 and 0.3 kg kg^{-1} water content aggregates to justify combining data from all dates for regression analysis.

For the field study, mean EF and standard errors were calculated for each date (n=10) and the means regressed against the cumulative freeze-drying cycles ($P < 0.05$). Bartlett's test (Steel and Torrie, 1980) for homogeneity of EF variances was non-significant.

2.3 RESULTS AND DISCUSSION

Laboratory study

The relationships between the number of freeze-drying cycles and EF for the three water contents and sampling dates are shown in Fig. 2-1. Anderson and Wenhardt (1966) reported that soils with >60% EF were prone to wind erosion and hence this value has been illustrated as the erosion threshold (Fig. 2-1). For all three sampling dates, aggregates wetted to 0.3 kg kg^{-1} water content crossed the erosion threshold (60% EF) after 1-2 freeze-dry cycles (Figs. 2-1a-c). The 0.2 kg kg^{-1} water content aggregates passed the erosion threshold after 8 freeze-dry cycles for all three dates (Figs. 2-1a-c). However, the 0.1 kg kg^{-1} water content aggregates were much slower in reaching the erosion threshold, with the 11 January samples eventually approaching the erosion threshold after 30 cycles (Fig. 2-1b). Due to time constraints the 0.1 kg kg^{-1} samples from October were terminated after only 8 freeze-dry cycles and the April samples after only 7 cycles.

This positive relationship between water content at time of freezing and aggregate disruption has also been reported by Hinman and Bisal (1968) in Saskatchewan. They found that when a clay was wet to -1.5 MPa water potential prior to freezing and freeze-drying, only 4.6% of aggregates >4 mm diameter were reduced to <1 mm in diameter. However, when the soil was wet to -0.03 MPa water potential, up to 28.5% of the aggregates were <1 mm diameter after freeze-drying. In comparison, the mean EF for all sampling dates after one freeze-dry cycle was 5.3% for aggregates pre-wet to 0.1 kg kg^{-1} water content and 43.5% for those pre-wet to 0.3 kg kg^{-1} water content for our clay loam soil. However, our aggregates were initially 1.2-2.0 mm diameter rather than >4 mm while degree of disruption was measured as aggregates <0.86 mm diameter rather than <1 mm in the case of Hinman and Bisal (1968).

Aggregates sampled on 26 October 1993 and pre-wet to 0.2 and 0.3 kg kg^{-1} water contents were more resilient to the disruptive forces of the freeze-drying compared with the 11 January and 29 April 1994 samples (Table 2-1). For example, October EF values were significantly lower (P <0.05) than January values up to the seventh freeze-drying cycle, for aggregates pre-wet to 0.2 kg kg^{-1}. However, October aggregates pre-wet to 0.3 kg kg^{-1} were more resilient than the January and April aggregates after the first freeze-drying cycle only. Bullock et al. (1988) showed that wet aggregate stability of silt loam soils in Idaho and Utah decreased from fall to spring and suggested that this was related to freeze-thaw cycles. Layton et al. (1993) noted an overwinter decrease in dry aggregate stability of a silt loam soil on clean tilled fields (residue buried) in Kansas. Stability decreases were attributed to substantial precipitation and frequent exposure to freeze-drying, freezing and thawing and wetting and drying processes. Other researchers have reported that freeze-drying alone does not alter inherent soil properties and that the surface of the soil particle is unchanged according to heat of immersion measurements (Bisal and Pelton, 1971).

In order to model the relationships between the number of freeze-dry cycles and EF for the different pre-wetting water contents, data from all three sampling dates were pooled (Fig. 2-2), except for the 0.1 kg kg^{-1} water content aggregates where only January samples were available. An exponential model described the relationship for the 0.1 kg kg^{-1} water content aggregates (Fig. 2-2a). From this equation, we calculated that 31.4 cycles were required to reach the erosion threshold of 60% EF.

An analysis of residual variation was performed on the 0.2 kg kg^{-1} water content aggregates and this substantiated that there was no bias in combining the data from all dates for regression analysis (P <0.05; Mead and Curnow, 1983). A linear model proved the best fit for the 0.2 kg kg^{-1} aggregates and we calculated that 9.3 cycles were necessary to reach the erosion threshold of 60% EF (Fig. 2-2b).

As with the 0.1 kg kg^{-1} aggregates, the relationship between the number of freeze-drying cycles and EF for the 0.3 kg kg^{-1} water content aggregates was exponential (Fig. 2-2c). An analysis of residual variation on these data indicated that the three sampling dates should not be combined (P <0.05). However, after observing differences of approximately 1% EF between observed and predicted values and an R^2 value of 0.99, we decided that this model was acceptable. The equation suggested that only 2 freeze-drying cycles were required to reach the erosion threshold of 60% EF.

Field Study

To quantify the occurrence of freeze-drying in the winter 1994-95 field study, not only was it necessary to estimate favourable vapour pressure gradients from meteorological data, we also had to assume that freeze-drying occurred only when favourable vapour pressure gradients were accompanied by a soil surface bare of snow cover and with a temperature <0 °C.

Using these criteria and the VPL gradient estimate, we calculated that under the specified boundary conditions, freeze-drying occurred in 133 distinct cycles ranging in duration from 1 to 104 h (Fig. 2-3a) with a cumulative total of 1142 h. All cycles occurred between 8 October 1994 and 13 May 1995 (a period of 5208 h). Therefore freeze-drying processes occurred for 21.9% of the elapsed time between these dates. The VPL values ranged from -1 to -427 Pa with a mean of -97 Pa (Fig. 2-3a).

Using the smaller vapour pressure gradient estimate (VPS), freeze-drying occurred in 59 cycles ranging in length from 1 to 76 h (Fig. 2-3b) with a cumulative total of 434 h. Freeze-drying occurred between 31 October 1994 and 8 April 1995 (3816 h) or 11.4% of the time between these dates. Values ranged from -1 to -331 Pa with a mean of -37 Pa (Fig. 2-3b). These values are substantially less compared to the values estimated using VPL criteria.

Previous investigators (Hinman and Bisal, 1968; Staricka and Benoit, 1995) as well as our laboratory results have shown that soil water content at freezing is an important factor influencing soil disruption by freeze-drying. Volumetric soil water contents for the 0-2.5-cm depth, taken 38 times over the VPL period (Fig. 2-3c), ranged from 0.18 m m^{-3} on 14 October 1994 to 0.05 m m^{-3} on 13 April 1995. The mean gravimetric water content over the VPL period was 0.10 kg kg^{-1} with a mean bulk density of 0.98 Mg m^{-3}. This soil water content is very close to the permanent wilting point of this soil (0.11 kg kg^{-1}, Chang et al., 1990) and identical to the 0.10 kg kg^{-1} pre-wetting treatment we used for aggregates in the laboratory study. Our laboratory study showed that >30 freeze-drying cycles would be required to reach the erosion threshold of 60% EF at these low field water contents (Fig. 2-2a).

Volumetric soil water contents for the 0- to 2.5-cm depth were taken 27 times over the VPS period (Fig. 2-3c). Contents ranged from 0.15 m m^{-3} on 20 January 1995 to 0.06 m m^{-3} on 23 December 1994 and 31 March 1995. The mean gravimetric water content over the VPS period was 0.09 kg kg^{-1} with a mean bulk density of 1.0 Mg m^{-3} which was similar to that estimated for the VPL period as well as to the permanent wilting point water content of this soil.

Soil profile water contents taken to 1.5 m depth on 22 September 1994 averaged only 0.14 kg kg^{-1}, with the 15- to 30-cm depth having the highest water content (0.19 kg kg^{-1})., Anderson and Bisal (1969) found that fall-irrigated soils are more prone to aggregate breakdown by freeze-drying than non-irrigated soils. Subsurface water can move upward toward a freezing front at the soil surface due to pressure and temperature gradients (Cary et al., 1979; Pikul and Allmaras, 1985). Accumulation of water, at or near the soil surface, accompanied by freezing at the freezing front could influence soil aggregation on freeze-drying. However, because of the dry profile conditions of the study soil, the amount of water available for migration to the surface as the soil cooled in the fall was minimal. The mean surface water content (0-2.5-cm depth) over the study period (0.10 kg kg^{-1}) was only about 0.04 kg kg^{-1} less than the mean profile water content measured in the fall. This would indicate that surface soil water contents would not be conducive to extensive aggregate breakdown by freeze-drying.

A quadratic model best described the relationship between EF and cumulative freeze-drying cycles for both VPL (Fig. 2-4) and VPS (Fig. 2-5) scenarios. This was in contrast to the exponential relationship for the 0.1 kg kg^{-1} laboratory treatment and the linear relationship for the 0.2 kg kg^{-1} laboratory treatment (Fig. 2-2). The flattening and possible downward trend of the quadratic prediction as spring approaches is probably due to processes other than freeze-drying. During this overwinter study we calculated 50 freeze-thaw cycles (based on –2°C air temperature for freeze and +2°C for thaw) (Bullock and Larney, 1997). There was also a positive relationship with EF and cumulative freeze-thaw cycles. Freeze-thaw process can be quite detrimental to wet aggregate stability (Bullock et al., 1988) and dry aggregate stability (Layton et al., 1993). In the spring, we observed that snowmelt and precipitation in the form of rainfall caused consolidation and crusting of low-stability surface aggregates, which would decrease the percent of erodible aggregates toward the end of the study.

Although the quadratic models fit our data best, the equations did not allow predictions for the erosion threshold of 60% EF. This was probably due to the small range of EF values available for development of the models (i.e., VPS and VPL EF ranged from 32.4 to 54.9%). Linear regressions were performed in order to allow comparison to the laboratory study and to make a prediction of the number of freeze-dry cycles that would be necessary to reach the wind erosion threshold of 60% EF (Figs. 2-4 and 2-5).

We estimated that 201 freeze-dry cycles were required to reach erosion threshold for VPL conditions and 89 for VPS conditions. These values are higher than the 31 estimated for the 0.1 kg kg^{-1} water content aggregates in the laboratory study (Fig. 2-2a). This discrepancy may be due to additional sources of energy influencing the freeze-drying process.

Under laboratory conditions, energy for freeze-drying was provided as the freeze-dry flasks that contained the frozen soil were exposed to room temperatures (about 21°C). Under field conditions during the VPL period, air temperatures were >0°C for 25.3% of the time while the soil surface was <0°C. Energy was supplied for freeze-drying under these conditions and using this percentage value of 25.3, the predicted number of cycles required to reach erosion threshold dropped from 201 to 51, which is closer to the laboratory study prediction of 31. Since air temperatures remained <0°C during freeze-drying cycles estimated using VPS, a similar adjustment could not be made.

Solar radiation is another energy source for freeze-drying. For the VPL estimate, freeze-drying occurred during daylight (0700-1700 h) for 29.7% of the time while it occurred during daylight for 41.0% of the time using the VPS estimate. The number of cycles required to reach erosion threshold dropped from 201 to 60 for the VPL estimate and from 89 to 37 using the VPS estimate if predictions are adjusted for the percentage of time freeze-drying was accompanied by solar radiation. The solar radiation adjustment for VPS conditions resulted in a cycle count of 37 which was very close to the 31 cycles predicted in the laboratory study (Fig. 2-2a).

2.4 CONCLUSIONS

A laboratory study conducted on clay loam soils in southern Alberta showed that water content of soil at time of freezing was a more important factor in soil aggregate disruption than the sampling date. However, fall-sampled aggregates were slightly more resilient to disruption when exposed to freeze-drying, than those sampled in winter or spring. Exponential relationships between cumulative freeze-dry cycles and EF were found for aggregates that were cyclically wet to 0.1 and 0.3 kg kg^{-1}, frozen and freeze-dried. A linear relationship was found for aggregates wet to 0.2 kg kg^{-1} water content. The number of freeze-drying cycles required for aggregates to reach the erosion threshold of 60% EF was 31.2 for aggregates pre-wet to 0.1 kg kg^{-1} water content, 9.3 cycles for those pre-wet to 0.2 kg kg^{-1}, and 1.7 cycles for those pre-wet to 0.3 kg kg^{-1} water contents. Dry soils (0.1 kg kg^{-1}) subjected to sufficient freeze-drying cycles (>30) eventually reached an erosion threshold. This finding has implications for wind erosion risk of arid and semi-arid soils of the Canadian prairie and northern Great Plains.

In a companion field study, we established quadratic relationships between EF and cumulative freeze-dry cycles. However, it was necessary to use linear relationships to make comparisons with laboratory findings for the number of cycles needed to reach the erosion threshold. By distinguishing periods where energy inputs from air temperatures >0°C or solar radiation accompanied favourable vapour pressure gradients, the estimates of freeze-dry cycles were close to laboratory predictions. Our study demonstrates the importance of freeze-drying as a mechanism in the aggregate breakdown and wind erosion risk of soils on the Canadian Prairies and the northern Great Plains. It also confirms that freeze-drying should be included as a factor in future wind erosion prediction models.

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Table 2-1. Effect of sampling date on erodible fraction (EF) of aggregates subjected to successive freeze-drying cycles at three water contents [means within soil water contents and cycles followed by the same letter are not significantly different from each other (P <0.05, n=18)]

	Erodible fraction, EF (%)								
	Water content								
	0.1 kg kg^{-1}			0.2 kg kg^{-1}			0.3 kg kg^{-1}		
Cycle	Oct. 26/93	Jan. 11/94	Apr. 29/94	Oct. 26/93	Jan. 11/94	Apr. 29/94	Oct. 26/93	Jan. 11/94	Apr. 29/94
1	2.8c	5.2b	8.1a	3.8c	7.7a	5.8b	40.2b	44.3a	46.1a
2	5.2c	8.5b	12.2a	6.3b	18.7a	16.5a	67.0a	65.9a	66.2a
3	6.9c	10.5b	15.2a	8.7b	21.2a	19.2a	79.3a	77.3a	78.0a
4	11.5b	12.5b	18.6a	15.7b	26.6a	23.8a	86.6a	83.2b	82.4b
5	14.5b	14.4b	20.6a	23.6b	32.3a	27.7ab	91.3a	87.2b	86.0b
6	17.7b	16.2b	24.0a	29.6b	36.8a	34.9a	-	-	-
7	20.8b	19.4b	29.1a	39.6b	42.3ab	46.7a		-	-
8	-	-	-	53.7a	53.3a	57.5a		-	-
9	-	-	-	64.1a	63.3a	63.2a		-	-

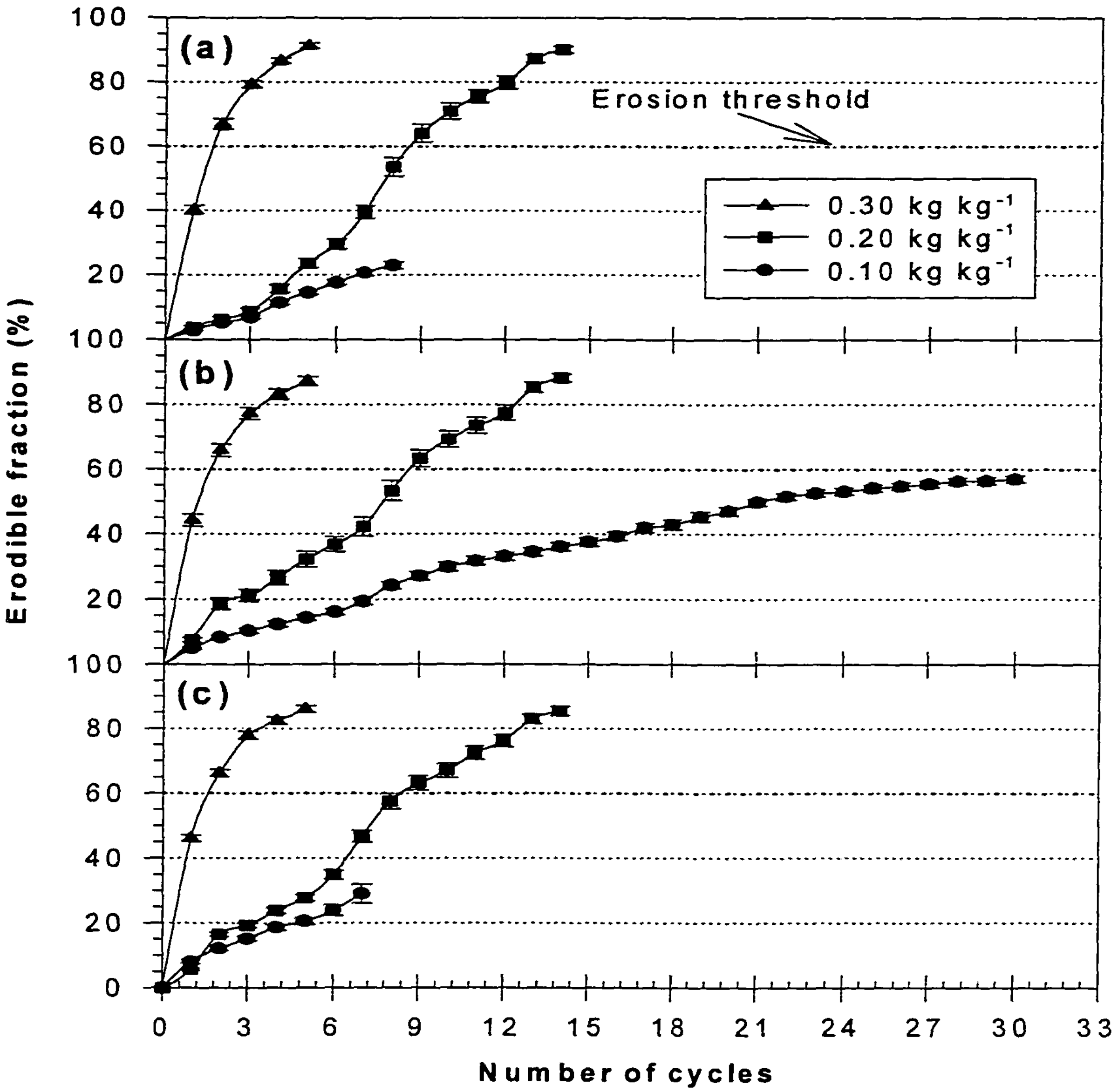

Fig. 2-1. Effect of number of freeze-drying cycles on cumulative EF (aggregates <0.86 mm diameter) for 1.2- to 2.0-mm aggregates pre-wet to three water contents sampled on (a) 26 October 1993; (b) 11 January 1994; and (c) 29 April 1994 (error bars = ±1 std. error).

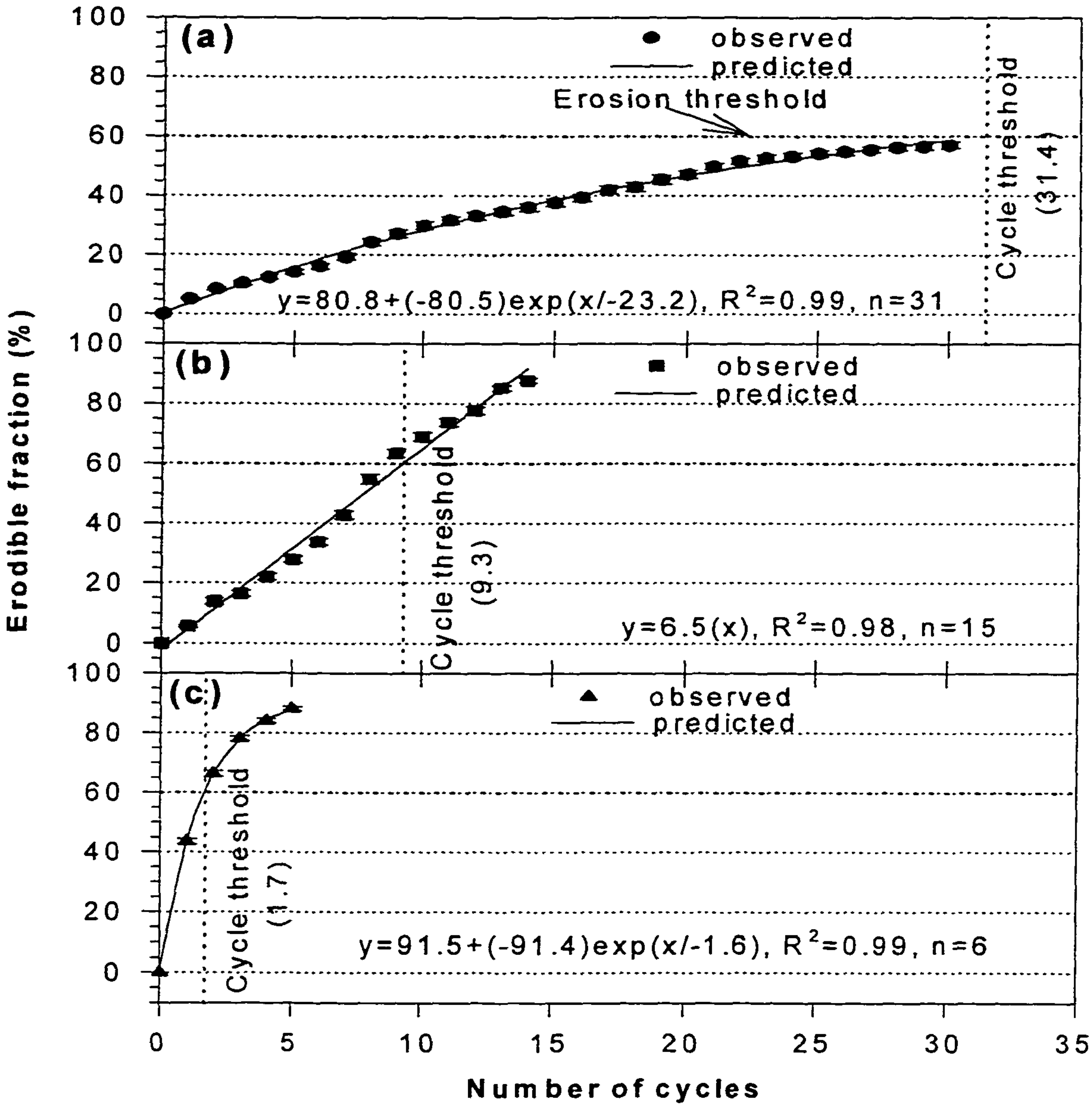

Fig. 2-2. Relationships between number of freeze-dry cycles and erodible fraction (26 October 1993, 11 January 1994 and 29 April 1994 combined) for aggregates pre-wet to (a) 0.1 kg kg^{-1}; (b) 0.2 kg kg^{-1}; and (c) 0.3 kg kg^{-1} water contents (error bars = ±1 std. error).

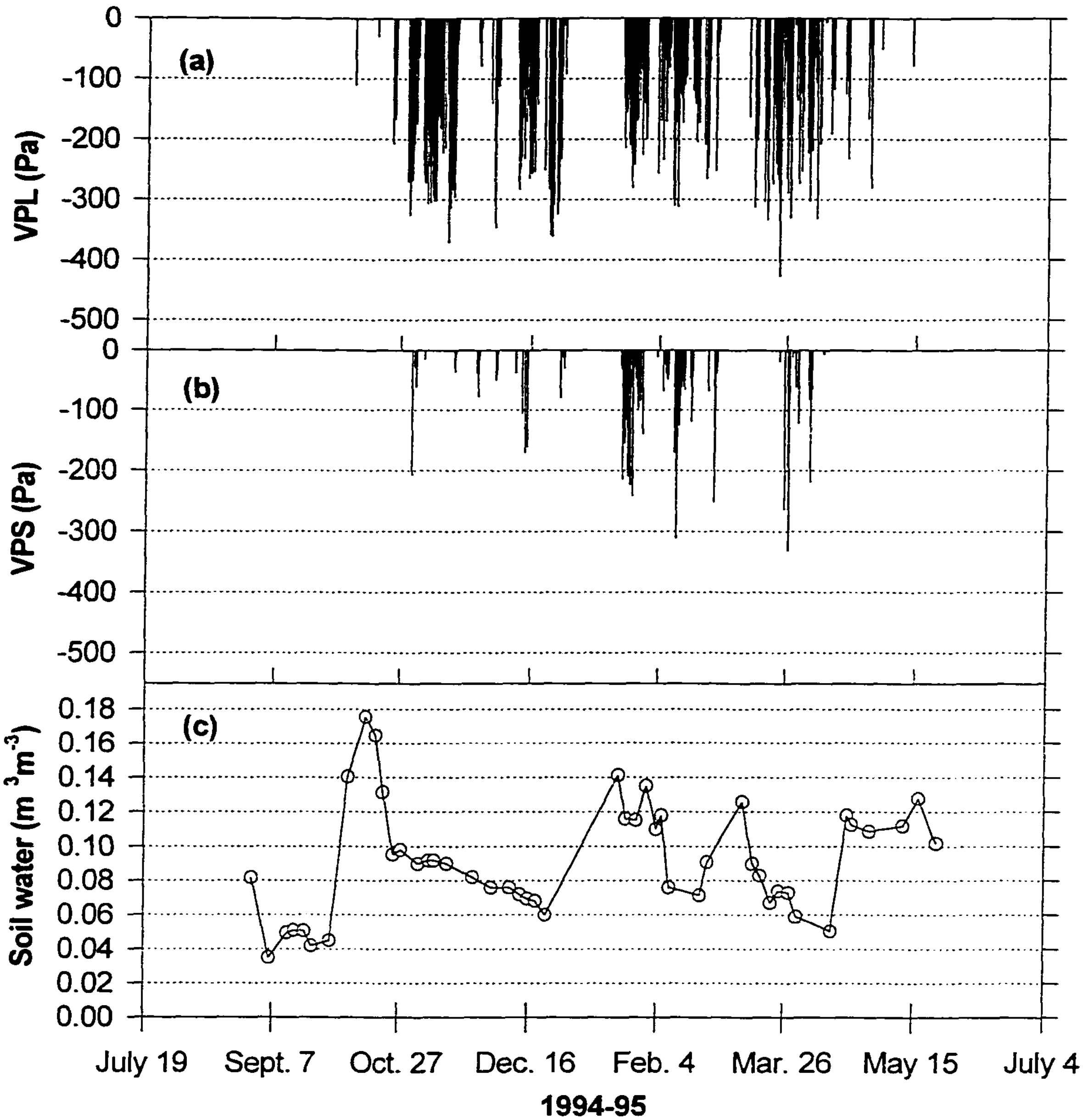

Fig. 2-3. Hourly vapour pressure gradient estimates using the (a) large vapour pressure gradient (VPL) criteria; and (b) small vapour pressure gradient (VPS) criteria, and (c) volumetric water contents for the duration of the field study (0- to 2.5-cm depth).

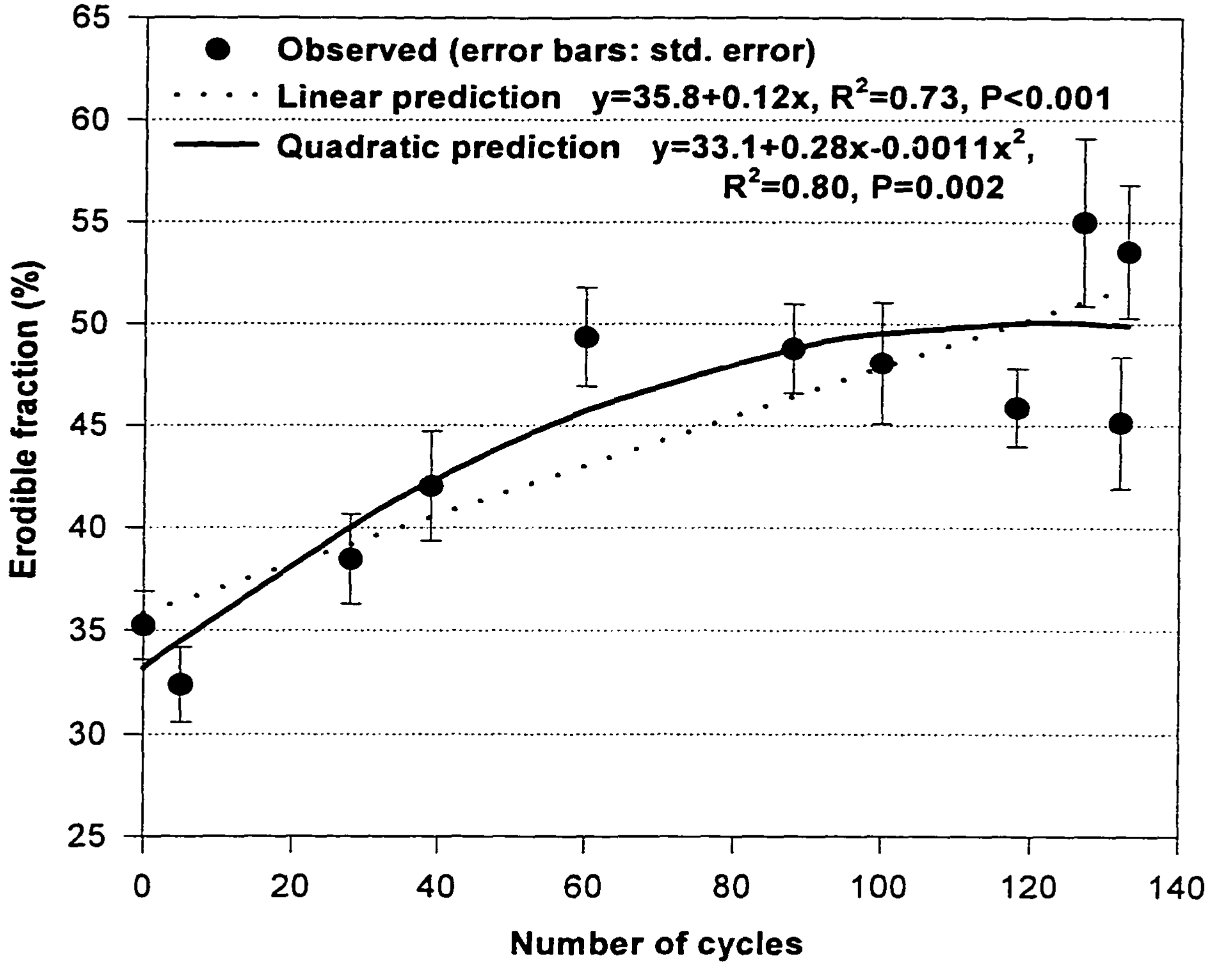

Fig. 2-4. Relationship between erodible fraction and the cumulative number of freeze-dry cycles estimated by the large vapour pressure gradient (VPL) criteria.

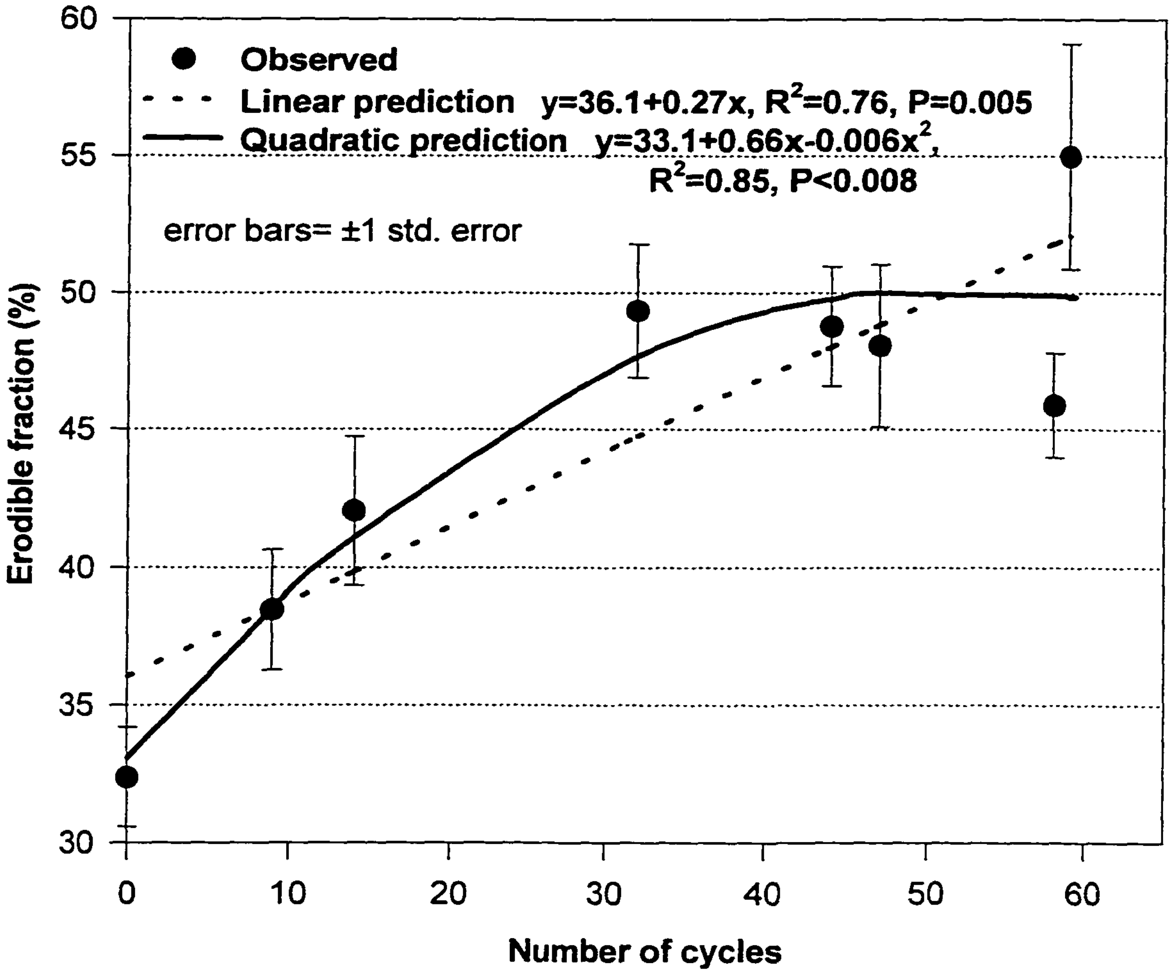

Fig. 2-5. Relationship between erodible fraction and the cumulative number of freeze-dry cycles as estimated by the small vapour pressure gradient (VPS) criteria.

Chapter 3

Overwinter changes in wind erodibility of clay loam soils in southern Alberta[1]

3.1 INTRODUCTION

Wind erosion is one of the main forms of soil degradation in the semiarid region of the Canadian prairies (Wall et al., 1995). The associated loss of organic matter and nutrient-rich topsoil results in a decrease in soil productivity (Larney et al., 1998). Additionally, off-site effects of wind erosion may impact air and water quality (Cihacek et al., 1993; Larney et al., 1999; Saxton, 1995).

Merrill et al. (1999) defined soil wind erodibility as the tendency of surface soil to resist or be vulnerable to being transported by wind. Both soil and plant factors contribute in the estimation of wind erodibility risk for any particular environment. The soil factors include inherent wind erodibility, soil surface roughness, and soil wetness while the plant factors include surface cover of the soil by living plants and their residues as well as the vertical effect of standing plants or residues (Merrill et al., 1999).

Dry aggregate size distribution (DASD), and soil surface roughness are important indicators of wind erodibility (Zobeck, 1991) and are influenced by management (tillage, cropping) and climatic factors (freeze-thaw, wetting-drying, freeze-drying). Overwinter breakdown of soil aggregates increases wind erosion risk, especially after conventionally-tilled fallow (heavy-duty cultivator, disk harrow), a

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common practice that leads to low crop residue levels or bare soils on the semi-arid Canadian prairies (Larney et al., 1995).

In southern Alberta, overwinter changes in DASD on five fallow management systems of a clay loam soil were related to climatic factors such as cumulative snowfall, snow cover days, and degree of freeze-thaw activity (Larney et al., 1994). Increased snowfall and snow cover days reduced aggregate breakdown while increased freeze-thaw cycles stimulated aggregate breakdown. In North Dakota, Merrill et al. (1995) found that climatic factors influenced DASD more than tillage treatments in a spring wheat-fallow cropping system. They found significant relationships between DASD and climatic factors such as number of snow cover days, number of freeze-thaw cycle days with no snow cover and fall precipitation. From fall to spring, snow cover and fall precipitation increased aggregate GMD while the number of freeze-thaw cycles decreased GMD. Furthermore, Merrill et al. (1999) reported that crop rotation, and sampling date were significant sources of variation in DASD while tillage effects were non-significant.

In Kansas and Texas, Chepil (1954) reported increases in erodibility from fall to spring for five soils ranging in texture from fine sandy loam to clay, with greatest increases on the finer textured soils. Erodibility increases were attributed to the freezing of moist soil during winter which caused expansion of ice crystals within aggregates and subsequent shattering.

Layton et al. (1993) discussed overwinter climatic processes and DASD changes for three tillage systems with different amounts of residue cover in Kansas.

Precipitation was the driving force behind DASD changes. However, precipitation effects were also influenced by residue amounts. A zero till treatment showed the most consistent fall- spring decrease in GMD, even though the high residue cover was thought to have decreased freeze-thaw and freeze-dry effects. Conversely, overwinter studies on a clay loam soil in Saskatchewan showed that bare soil surfaces exposed to freeze-thaw and freeze-drying cycles were more vulnerable to breakdown than those protected by snow cover (Anderson and Bisal, 1969).

Surface roughness is closely associated with DASD and was identified as a governing principle of erosion risk by Chepil (1950), who concluded that the degree of surface roughness is related to the size, shape, distribution and proportion of surface projections present as non-erodible aggregates. Römkens and Wang (1986) scaled field roughness into 4 classes including that due to individual particles or aggregates of 0-2 mm diameter; surface variations or random roughness on the order of 10 cm diameter; tillage or oriented roughness in the range of 10-30 cm diameter and roughness on a field topography scale. Large clods and tillage ridges perpendicular to the prevailing wind are effective for erosion control (Fryrear, 1984).

Zobeck and Onstad (1987) reviewed the many devices for measuring soil roughness along with the various indices used to summarize the data. Potter and Zobeck (1990) and Potter et al. (1990) developed a microrelief or surface roughness index that estimated microrelief effects on soil susceptibility to wind erosion. Most methods for measuring soil roughness have used pin meters but recently a roller chain method was developed by Saleh (1993). The advantages and disadvantages of using the roller chain method have been discussed (Merrill, 1998; Saleh, 1997; Skidmore, 1997).

Previous research on overwinter effects on DASD has consisted of only fall and spring sampling (Larney et al., 1994; Layton et al., 1993; Merrill et al., 1995). This type of two-point sampling precludes the development of equations to predict relationships between aggregate size changes and climatic variables. Since overwinter change in DASD and surface roughness are dynamic phenomena, this study was initiated to monitor overwinter changes in these parameters and determine relationships between these changes and climatic variables for three clay loam soils in southern Alberta.

3.2 MATERIALS AND METHODS

Study Sites

The study was conducted over three consecutive winters (1992-93, 1993-94, 1994-95) near Lethbridge, Alberta (49° 43' N, 112° 48' W). The mean annual precipitation (1902-98) at Lethbridge is 402 mm and the mean January temperature is -8.6°C. A different site was selected each winter. All study soils were Dark Brown Chernozemic clay loams (fine-loamy, mixed, Typic Haploborolls).

The Fairfield site (1992-93) was located 6 km east of Lethbridge. The 0-2.5 cm soil layer contained 39% sand, 28% silt, 33% clay and 17 g kg^{-1} organic carbon. Silage barley (*Hordeum vulgare* L.) was grown on the site from 1989-1991 using conventional tillage (heavy-duty cultivator, disk harrow). In 1992, the site was fallowed with herbicides and one pass of a wide-blade cultivator. Stubble residue was pulverized by a severe hail storm in August 1992. On 18 Sep. 1992, a 20 x 20 m plot was chisel-ploughed to a 12-cm depth. This resulted in a ridged, roughened surface with negligible residue cover. The Wilson site (1993-94) was located 15 km south of Lethbridge. The soil contained 35% sand, 31% silt, 34% clay, and 17.2 g kg^{-1} organic carbon in the 0-2.5 cm layer. The site had been continuously cropped under no-till since 1984. Spring wheat (*Triticum aestivum* L.) was grown in 1992. From May-August 1993, a 200 m diameter circle (3.14 ha) was conventionally fallowed with three passes (11 May, 28 June, 2 August) of a heavy-duty cultivator equipped with a hydraulically-driven rod weeder. A final pass with a heavy-duty cultivator, rod weeder and tine harrows to a depth of 8 cm was conducted on 26 Oct. 1993 resulting in minimal surface residue.

The Lethbridge site (1994-95) was located 1 km north of Lethbridge and the soil contained 41% sand, 27% silt, 32% clay, and 15 g kg^{-1} organic carbon in the surface 7.5 cm layer. The site had been in a spring wheat-fallow rotation since 1955 (Johnston et al., 1995). It was cropped in 1993 and fallowed in 1994 with three passes (24 May, 8 July, 30 August) of a one-way disk.

Dry Aggregate Size Distribution

At all three sites, 5 kg of soil was sampled from the 0-2.5 cm depth with a flat shovel for DASD analysis. Sampling interval varied, being dictated by the absence of snow cover. At the Fairfield site, the 20 x 20 m chisel-plowed plot was divided into twenty 1-m-wide x 20-m-long sub-plots for DASD sampling. A different sub-plot was randomly chosen on each of 15 sampling dates between 18 Sep. 1992 and 12 May 1993. Ten sub-samples were taken at each sampling date.

At the Wilson site, samples for DASD were taken on 9 dates between 26 Oct. 1993 and 29 Apr. 1994. Locations were at 10 m intervals along a 200 m west-east transect on the circular plot for a total of 20 sub-samples per sampling date.

At the Lethbridge site, two 6 x 20 m plots were each divided into twenty 1-m-wide x 6-m-long sub-plots. A different sub-plot on each of the two plots was randomly chosen on each of 14 sampling dates between 30 Aug. 1994 and 24 May 1995. Five DASD samples were taken from each of the sub-plot for a total of 10 per sampling date.

All DASD samples were air-dried and passed through an improved rotary sieve (Chepil, 1952) with 38, 12.6, 7.1, 1.9, 1.2, and 0.47 mm diameter sieves. DASD was expressed as geometric mean diameter (GMD), the aggregate diameter at which 50% of the sample weight passed (Gardner, 1956), and erodible fraction (EF), the percent of aggregates <0.84 mm diameter. Since a 0.84 mm sieve was not used, EF was calculated by substituting 0.84 into the regression equation for cumulative percent passing *vs.* log of sieve size. GMD has a non-linear, inverse relationship to EF.

Generally, aggregates <0.84 mm diameter are erodible by wind, and soils with an EF >60% are considered at relatively high erosion risk (Chepil, 1941, 1942).

Soil Surface Roughness

Soil surface roughness parallel to the direction of tillage or random roughness (C_{par}), which is roughness caused by aggregates only, was measured at the Wilson and Lethbridge sites, by the chain method of Saleh (1993) using a 1-m roller chain with 0.95-cm links. This method expresses roughness as a percentage value (Cr) as calculated in Eq. [1]:

$$Cr = (1 - L1/L2) \times 100 \qquad [1]$$

where L1 is the horizontal distance (m) between the ends of the chain lying on the soil surface and L2 is the length of the chain (Saleh, 1993). Zero percent is considered a perfectly flat surface. Roughness perpendicular to tillage (C_{per}), which includes roughness associated with both tillage operations (oriented roughness) and aggregates, was also measured. However, C_{per} data are not reported as values were similar to C_{par} values, since there were no obvious tillage ridges at the three sites.

In fall, 10 positions were randomly chosen for soil surface roughness measurements at the Wilson site, and 12 at the Lethbridge site. These positions were permanently marked and visited on each measurement date. C_{par} was measured 11 times at the Wilson site between 26 Oct. 1993 and 10 May 1994 and 14 times at the Lethbridge site between 30 Aug. 1994 and 24 May 1995. Soil roughness was measured when there was no snow cover.

Climatic Variables

Mean hourly air temperature (1 m height) was measured on each site, commencing on 23 Oct. 1992 at Fairfield, 1 Nov. 1993 at Wilson, and 30 Aug. 1994 at Lethbridge. From air temperature data, the number of freeze-thaw cycles was determined for all days and for days without snow cover. Cutoffs of –2°C for freezing and +2°C for thawing were used to estimate the number of soil freeze-thaw cycles. Rainfall, snowfall, snow depth and total precipitation were measured at the Agriculture and Agri-Food Canada Research Centre weather station, located approximately 5 km from the Fairfield site, 15 km from the Wilson site and 1 km from the Lethbridge site. A trace of snow constituted a snow cover day.

Statistical Analysis

Mean GMD, EF and C_{par} values and standard errors were calculated for each sampling date. For all three sites, regression analysis (SAS Institute Inc., 1989) was performed between the mean values of the dependent variables GMD and EF and cumulative days from initiation of the study. For the Wilson and Lethbridge sites a similar regression was performed with C_{par} as the dependent variable.

3.3 RESULTS AND DISCUSSION

Geometric Mean Diameter

All three sites showed a general decline in geometric mean diameter (GMD) with cumulative days from initial fall sampling (Figs. 3-1a-c). An exponential model described the relationship at all sites with R^2 values ($P = 0.05$) of 0.57 for Fairfield, 0.97 for Wilson and 0.79 for Lethbridge. The Fairfield site (1992-1993) recorded the lowest initial GMD value post fall-tillage (1.88 mm) and also the lowest GMD of the study (0.08 mm, 22 Apr. 1993).

An increase in GMD occurred at the Fairfield site midway through the study (Fig. 3-1a). Snowmelt (about 8 cm of snow depth) followed by evaporative drying allowed the saturated soil surface to crust by the 1 Feb. 1993 sampling date. DASD sampling of a crusted soil surface created clods of considerable size, strong enough to resist breakdown during rotary sieving, which increased GMD. Soils in northern climatic zones are prone to crusting during winter, due to a decline in wet aggregate stability from fall to spring caused by freeze-thaw cycles (Bullock et al., 1988). Merrill et al. (1995) also reported increases in GMD after snowmelt in spring. However, climatic conditions during February and March 1993 allowed GMD to return to pre-crusting values (Fig. 3-1a). The decline in GMD after surface crusting followed a similar pattern to the post-tillage decline in the fall (Fig. 3-1a). If the 1 February value is treated as an outlier and deleted, the R^2 value of the exponential relationship increased to 0.73 ($P = 0.05$).

At the Wilson site, GMD ranged from a maximum of 9.05 mm after fall tillage (26 Oct. 1993) to a minimum of 1.17 mm (14 Mar. 1994). Although the maximum and minimum GMD values were greater than those of the other study sites, the decline in GMD over time was greater than at the other sites, as denoted by the different form of exponential equation (Fig. 3-1b).

At the Lethbridge site, GMD ranged from 4.71 mm after fall tillage (30 Aug. 1994) to 0.80 mm (17 May 1995). Although the maximum fall GMD value at the Lethbridge site was much lower than that at the Wilson site (4.71 *vs.* 9.05 mm), minimum GMD values in spring were quite similar (0.80 *vs.* 1.17 mm). This demonstrates that similar levels of erosion risk may exist in spring on two soils that had widely different DASD values in the previous fall.

Erodible Fraction

All sites showed a general increase in erodible fraction (EF) from post-tillage values in fall to final values in spring (Figs. 3-2a-c). Linear regressions of EF data with cumulative days from initial fall sampling resulted in R^2 values ($P = 0.05$) of 0.57 for the Fairfield site, 0.91 for the Wilson site and 0.62 for the Lethbridge site. EF ranged from a minimum value of 12.6% (26 Oct. 1993, Wilson site) to a maximum of 74% (22 Apr. 1992, Fairfield site). Anderson and Wenhardt (1966) determined that soils with EF values >60% were generally erodible by wind. This threshold level was exceeded at the Fairfield site only from 4 Dec. 1992-1 Feb. 1993 and from 3 March-12 May 1993. Like the GMD data, EF data from the Fairfield site reflected the effects of surface crusting with a large decrease in the EF on 1 Feb. 1993. However, EF values rapidly returned to pre-crusting values (Fig. 3-2a). If the EF value for 1 February is deleted, the R^2 value for the linear regression increased from 0.57 to 0.68 ($P = 0.001$).

Linear regression intercepts were 15.4 for Wilson, 33.6 for Lethbridge, and 42.2 for Fairfield indicating that the Wilson site was least erodible at initial fall sampling. Regression slopes ranged from 0.06 for Lethbridge, to 0.12 for Fairfield and 0.18 for Wilson. These values indicate that the rate of increase in EF over time varied from 0.06-0.18% d^{-1}. The Wilson soil, with the lowest initial EF in fall, showed the fastest overwinter increase in EF. Also, the Fairfield site, which had the highest EF in fall did not have the slowest overwinter increase in EF.

Soil Surface Roughness

Soil surface roughness (C_{par}) values for the Wilson site (Fig. 3-3a) ranged from 15.1 (26 Oct. 1993) to 3.7% (30 Mar. 1994). C_{par} values for the Lethbridge site (Fig. 3-3b) ranged from 14.4 (30 Aug. 1994) to 3.3% (19 Apr. 1995). Both sites exhibited an overwinter smoothing trend following the rough conditions created by fall tillage. Merrill et al. (1999) reported a range of C_{par} from 12.5-3.1% for a low residue till (<10% residue cover) spring wheat-fallow study in North Dakota which was very similar to the ranges found at our sites.

The relationships between C_{par} and cumulative time since the start of each overwinter study were described by exponential equations which had R^2 values (P = 0.05) of 0.98 for Wilson and 0.91 for Lethbridge (Figs. 3a-b). Although the initial fall values of C_{par} were similar at each site, the dates of last fall tillage were quite different (26 October for Wilson, 30 August for Lethbridge). However, roughness at both sites declined at similar rates before the rate of smoothing slowed. At the Wilson site, roughness declined from 15.1 to 5.1% between 26 Oct. 1993 and 15 Feb. 1994, a period of 112 d. Roughness at the Lethbridge site declined from 14.4 to 4.3% between 30 August and 19 Dec. 1994, a period of 111 d.

Climatic Variables and Wind Erodibility Changes

The total number of freeze-thaw cycles, using the air temperature cutoffs of +2°C for thawing and -2°C for freezing, was 54 for Fairfield in 1992-93, 65 for Wilson in 1993-94 and 50 for Lethbridge in 1994-95 (Table 3-1). The number of cycles on days with no snow cover was 45 for Fairfield, 58 for Wilson, and 39 for Lethbridge. Therefore, soils were exposed to 78-89% of the total freeze thaw cycles occurring over the study periods. The freeze-thaw and EF data were grouped into three periods based on the timing and form of precipitation (Table 3-1). The first or "fall rain/snow" period started at the beginning of the study and ended after the last rainfall event in fall although precipitation may have been in the form of rain or snow during this period. The greatest increase in erodibility during the "fall rain/snow" period was at the Fairfield site (+7.4% change in EF, Table 3-1). This site also had the most freeze-thaw cycles (7) and 60.4 mm of total precipitation (Table 3-1). The Wilson site had an intermediate increase in EF (+4.1%), less freeze-thaw cycles (5) than Fairfield, but the least cumulative precipitation at only 7.0 mm. The Lethbridge site had the smallest increase in EF (+1.1%) with zero freeze-thaw cycles, but had the greatest cumulative precipitation (83.0 mm). These findings demonstrate that erodibility increases during this "fall rain/snow" period were greatest when a combination of high precipitation and a high number of freeze-thaw cycles existed. Ice expansion in inter- and intraaggregate pores helped aggregate breakdown (Chepil, 1954), and as air and soil temperatures fell below freezing and the "winter snow" period approached, the reversal of this disruptive process may not have been possible (Bullock et al., 1988).

The second or "winter snow" period represented the time from the last rainfall event in fall to the first rainfall event in spring when all precipitation fell as snow (Table 3-1). For all three sites the "winter snow" period produced the largest increases in EF (Table 3-1). The Wilson site showed the largest increase in erodibility (+25.1%) and had over twice the number of freeze-thaw cycles without snow cover (40 *vs.* 18) as the other two sites. Previous studies have shown a positive correlation between freeze-thaw cycles and EF increases (Larney et al., 1994; Merrill et al., 1995). Additionally, the higher snowfall amount at the Wilson site (82.4 mm) compared to the other sites (42-43 mm) may have allowed greater wetting of the soil surface during the thaw periods, leading to greater soil disruption upon subsequent freezing.

The Fairfield and Lethbridge sites had similar snowfall amounts and numbers of freeze-thaw cycles (Table 3-1), but the Lethbridge site had a larger increase in EF compared to Fairfield (+17.0 *vs.* +11.2). Other processes such as freeze-drying of the soil surface or soil abrasion by blowing snow (Bullock et al., 1992, 1999; de Jong and Kachanoski, 1988) may have contributed to this difference.

The third and final period was the "spring snow/rain" period which represented the time from the first rainfall event in spring and included precipitation as snow and rain until the end of each study year (Table 3-1). During the "spring snow/rain" period, only the Fairfield and Wilson sites showed increases in EF and these increases were the smallest of all three overwinter periods.

The Lethbridge site showed a decrease (-9.2%) in EF for this period which may be related to the high amount of total precipitation (155.7 mm) and a daily maximum rainfall of 24.4 mm. Most of the precipitation (78%) during this period at the Lethbridge site was in the form of spring rain near the end of the study (May) when the frequency of freeze-thaw cycles was declining. We believe this precipitation caused saturation of the soil surface, slaking of aggregates and soil crusting as reported by Kemper et al. (1987) and Uehara and Jones (1974). In addition, Anderson and Wenhardt (1966) found that a clay loam soil had a lower EF in spring than the previous fall in 5 out of 7 yr and attributed this to a positive correlation with precipitation. The reduced frequency of freeze-thaw cycles at this time of year also decreased soil disruption. Similar decreases in EF did not occur at Fairfield and Wilson as total precipitation amounts were lower than at Lethbridge for the "spring snow/rain" period (Table 3-1).

At the Lethbridge site, even though the "winter snow" and the "spring snow/rain" periods had similar numbers of freeze-thaw cycles (26 *vs.* 24, Table 3-1), total precipitation in the "winter snow" period was only 23% of that in the "spring snow/rain" period. Trends in erodibility were quite different with the "winter snow" period showing a large increase (+17.0% change in EF) and the "spring snow/rain" period showing a decrease (-9.2% change in EF). This illustrates that the type and amount of precipitation, rather than the number of freeze thaw cycles, is an important factor in the magnitude and direction of erodibility changes.

Climatic factors affecting overwinter changes in EF also played a role in overwinter changes in soil surface roughness (C_{par}). The overwinter decline in C_{par} (Fig. 3-3) was related to the breakdown of large aggregates by the same climatic forces that induced declines in EF. However, unlike EF, the impact of rainfall may account for a larger component of the decline in C_{par}, due to its smoothing effect (Römkens and Wang, 1987; Zobeck and Popham, 1990). Even though the Wilson and Lethbridge sites differed in the amounts and form of precipitation, the number of days with snow cover, and the number of freeze-thaw cycles (Table 3-1), final C_{par} values were similar at both sites (3.7% at Wilson, 3.3% at Lethbridge, Fig. 3-3). This illustrates that similar levels of soil roughness can exist in spring, despite being influenced by a different set of climatic factors over the previous winter.

Management Effects on DASD

Although the sites in this study had quite similar surface textures and organic carbon levels, their management histories varied. The Fairfield site had been continuously cropped with conventional tillage and all surface residue was removed for three years prior to the study. This site had the highest initial EF value (38.9%, Fig. 3-2a) and was the only site where EF increased to >60% (wind erosion threshold; Anderson and Wenhardt, 1966) during the winter period. Physical disruption from a history of tillage and the removal of crop residue seems to have increased erosion risk.

The Wilson site had been in zero tillage for 7 yr prior to the fallow season in 1993. Although this site had the lowest initial EF value (12.6%, Fig. 3-2b) of all three sites, it showed the fastest overwinter increase in EF (Fig. 3-2b). This agrees with the findings of Larney et al. (1994) and Layton et al. (1993) who indicated that zero till soils may have higher erosion risk than tilled soils if residue cover is jeopardized. In addition, the Wilson site was the only one where earthworm casts were observed on the soil surface. Clapperton et al. (1997) reported significantly higher populations of earthworms (*Aporrectodea calignosa*, Savigny) under zero tillage than under conventional tillage in southern Alberta. Earthworm casts are known to have lower bulk densities and have a propensity to absorb more water than surface soil, making them more susceptible to freeze-thaw disruption (Lal and Akinremi, 1983).

The Lethbridge site, which had been under long term wheat-fallow management with conventional tillage, had an initial EF value (30.4%, Fig. 3-2c) that fell between the values for the Fairfield (38.9%, Fig. 3-2a) and Wilson (12.6%, Fig. 3-2b) sites but had the lowest organic carbon levels of all sites (15 g kg^{-1}). The EF values at the Lethbridge site did not increase above the threshold value of 60%, unlike the Fairfield site, and increased very slowly over the winter period (Fig. 3-2c), unlike the Wilson site.

Our results suggest that management systems and cropping histories may play a significant role in overwinter changes in wind erodibility risk. They influence the starting value of EF in fall as well as the slope of the overwinter relationship between EF and time.

3.4 CONCLUSIONS

This 3 yr study elucidated the effect of overwinter climatic factors on wind erodibility of clay loam soils in southern Alberta. A decrease in GMD of soil aggregates, an increase in EF, and a decrease in soil surface roughness led to increased erodibility of these soils from fall to spring.

Dividing the overwinter period into three segments, distinguished by the timing and form of precipitation, provided useful insights into erodibility changes. In the "fall rain/snow" period, freeze-thaw cycles were detrimental to soil structure, especially if accompanied by appreciable precipitation which increased soil water content and provided a mechanism for soil disruption (*i.e.* ice expansion in confined pore space). The largest changes in erodibility occurred during the "winter snow" period when intermittent snowmelt likely increased soil water content and allowed freeze-thaw cycles to be more effective in aggregate breakdown. Freeze-drying and aggregate abrasion by blowing snow may have also contributed to de-aggregation during this period. In the "spring snow/rain" period, while freeze-thaw cycles and precipitation were still important in aggregate breakdown, heavy rains in late spring were the dominant factor in decreasing the erodibility of the Lethbridge soil in the form of a surface crust.

The study also demonstrated the effect of cropping history on wind erodibility and pointed to the fragility of zero tillage soils once the protective layer of surface residue is removed. A soil that was cultivated after seven years of zero tillage, although non-erodible in fall, was not resilient to overwinter climatic forces experienced in southern Alberta, and had approached erosion risk in spring.

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Table 3-1. Climatic variables and erodible fraction (EF) changes during "fall rain/snow", "winter snow" and "spring snow/rain" overwinter periods at Fairfield (1992-93), Wilson (1993-94) and Lethbridge (1994-95) sites.

	Fall rain/snow			Winter snow			Spring snow/rain		
	Fairfield	Wilson	Lethbridge	Fairfield	Wilson	Lethbridge	Fairfield	Wilson	Lethbridge
Start date	18 Sep.	26 Oct.	30 Aug.	13 Nov.	11 Nov.	28 Oct.	8 Feb.	29 Mar.	26 Jan.
End date	13 Nov.	11 Nov.	28 Oct.	8 Feb.	29 Mar.	26 Jan.	12 May	29 Apr.	24 May
No. of days	56	16	59	87	138	90	93	31	118
Total rainfall, mm	39.6	3.6	47.6	0	0	0	44.6	7.3	121.1
Max. daily rainfall, mm	10.0	2.4	13.2	0	0	0	12.6	5.0	24.4
Snowfall, cm	21.2	2.8	21.6	43.2	82.4	42.1	36.1	8.0	45.7
Total precip., mm	60.4	7.0	83.0	41.0	64.4	35.9	84.2	14.1	155.7
Snow cover, d	9	2	3	67	60	60	31	2	30
Mean snow depth, cm	2.5	0.8	2.7	3.8	4.4	3.1	4.6	6.3	2.7
Total FT[†] cycles	7	5	0	24	46	26	23	14	24
FT[†] cycles, no snow cover	7	5	0	18	40	18	20	13	21
EF change, %	+7.4	+4.1	+1.1	+11.2	+25.1	+17.0	+5.7	+1.3	-9.2

[†]FT cycles = freeze-thaw cycles

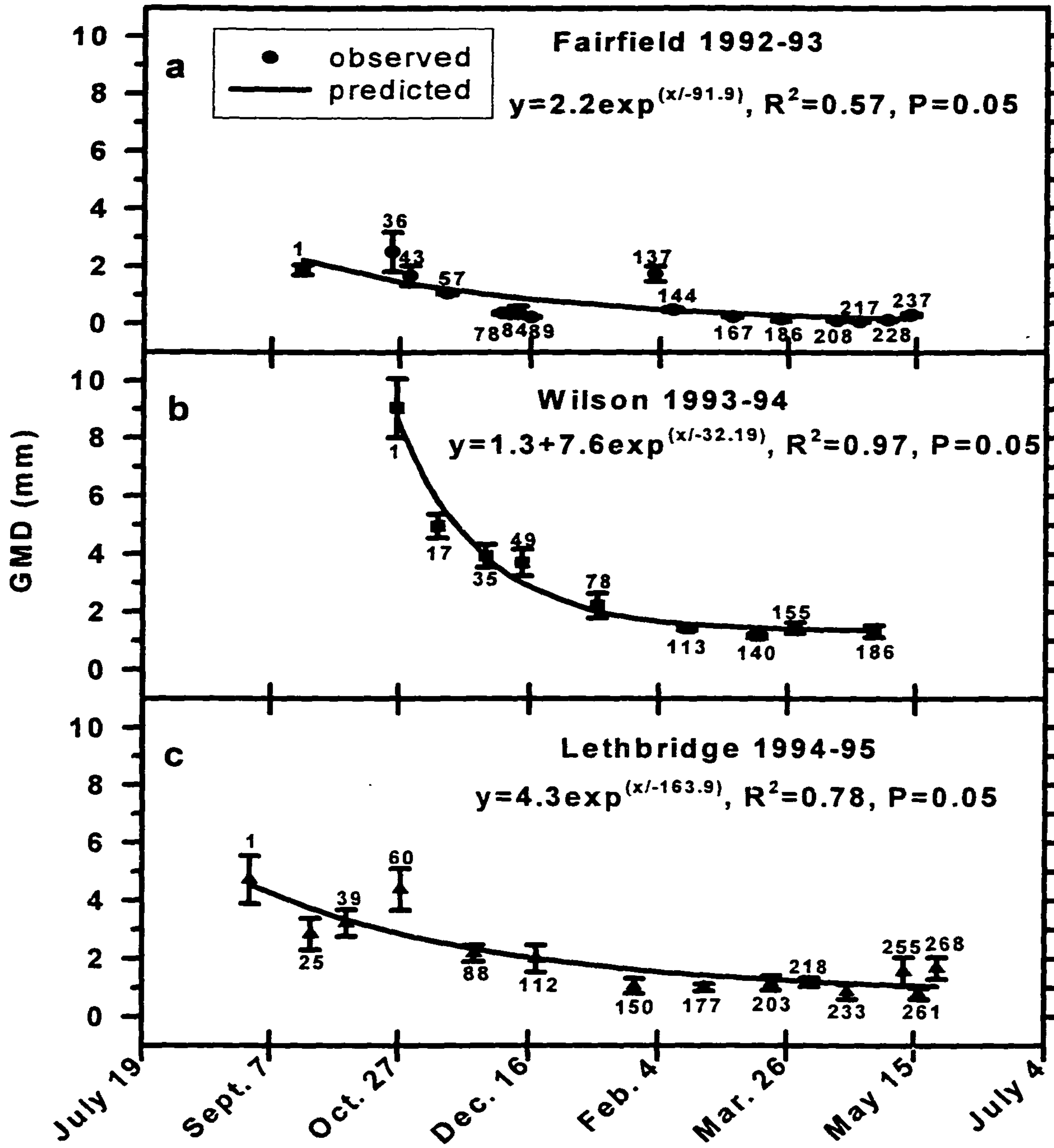

Fig. 3-1. Relationship between cumulative days from start of study in fall and geometric mean diameter (GMD) for (a) Fairfield (1992-93); (b) Wilson (1993-94) and (c) Lethbridge (1994-95) sites. For ease of interpretation the x-axis label is shown as date rather than cumulative days. Cumulative days are shown above or below data points. Day 1 = 18 Sept. 1992 at Fairfield; 26 Oct. 1993 at Wilson and 30 Aug. 1994 at Lethbridge. Error bars = ±1 std. error.

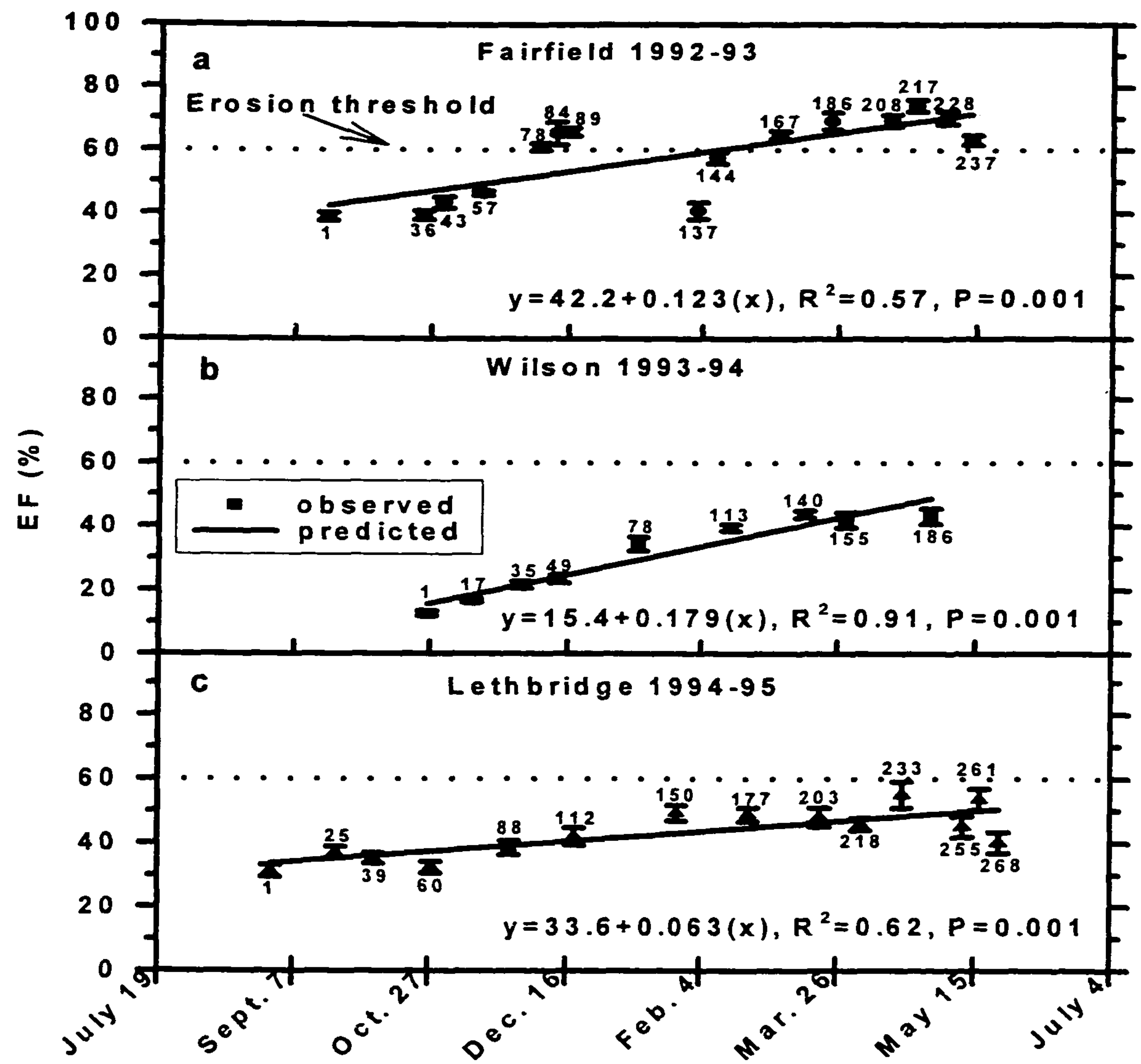

Fig. 3-2. Relationship between cumulative days from start of study in fall and erodible fraction (EF) for (a) Fairfield (1992-93); (b) Wilson (1993-94) and (c) Lethbridge (1994-95) sites. For ease of interpretation the x-axis label is shown as date rather than cumulative days. Cumulative days are shown above or below data points. Day 1 = 18 Sept. 1992 at Fairfield; 26 Oct. 1993 at Wilson and 30 Aug. 1994 at Lethbridge. Error bars = ±1 std. error.

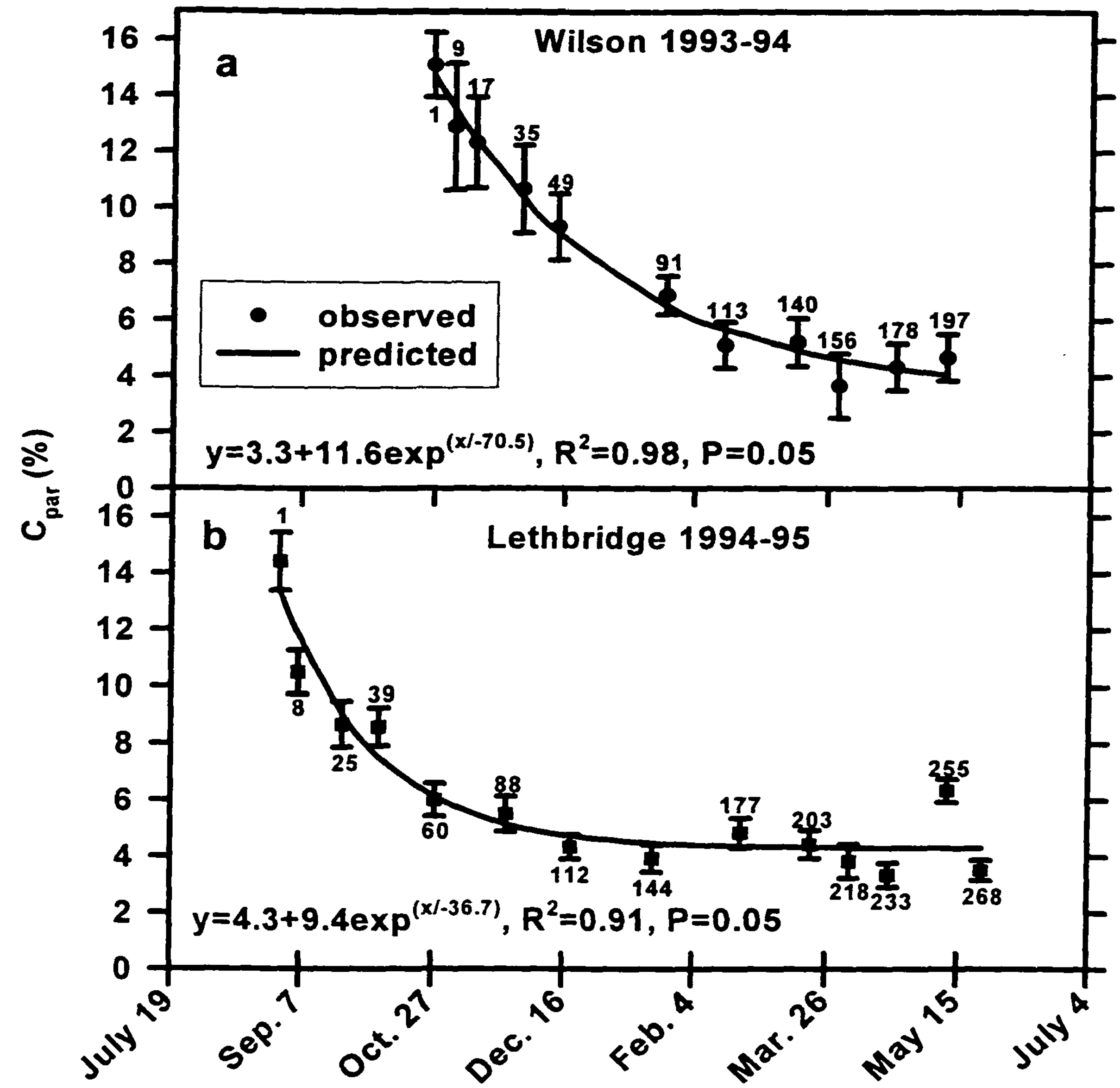

Fig. 3-3. Relationship between cumulative days from start of study in fall on soil surface roughness (C_{par}) for (a) Wilson (1993-94) and (b) Lethbridge (1994-95) sites. For ease of interpretation the x-axis label is shown as date rather than cumulative days. Cumulative days are shown above or below data points. Day 1 = 26 Oct. 1993 at Wilson and 30 Aug. 1994 at Lethbridge. Error bars = ±1 std. error.

Chapter 4

Preliminary results on the role of blowing snow on wind erodibility of clay loam soils in southern Alberta

4.1 INTRODUCTION

Wind erosion is a major contributor to soil degradation in the semi-arid region of the Canadian prairies (Wall et al., 1995). Overwinter breakdown of soil aggregates can increase soil erodibility and lead to wind erosion especially under tillage fallow systems like those in southern Alberta (Larney et al., 1994, 1995). An improved understanding of this process is needed if we are to design state-of-the-art prediction systems (i.e. Wind Erosion Prediction System- WEPS) to help plan effective wind erosion control measures (Hagen, 1991).

Southern Alberta is renowned for the winter winds called "chinooks" (Grace, 1987). These are warm dry winds that descend the leeward or eastern side of the Rocky Mountains. Wind speeds during these events generally range from 60 to 100 km hr^{-1} while temperature changes may exceed 20°C. Blowing snow abrasion of soil is an overwinter process that has been overlooked as a contributor to soil aggregate breakdown. Pomeroy et al. (1993) considered the saltation component in modeling snow transport and sublimation and found that 59-83% of transport occurred within 0.05 m of the snow surface at two sites in Saskatchewan.

Teichert (1939) noted that "snow at low temperatures possesses the physical properties required for corrasion effects even in hard rocks". Dietrich (1977) used silt sized ice crystals as abrader in a wind tunnel (7.8 to 8.6 m s^{-1}) for the mineral blocks of calcite, synthetic flourite and synthetic periclase. After only ten minutes the edges and corners of the blocks were “sporadically broken”. It was concluded that “wind erosion may involve much softer or smaller missiles, such as snow or dust”. Saltation of blowing soil is a major contributor to increasing the erodible mass (Hagen, 1984). However, the abrasive properties of blowing snow have not been related to wind erodibility.

Li and Pomeroy (1997a) defined the threshold wind speed for snow transport as the minimum wind speed initiating or sustaining saltation and found that it was related to properties of the surface snowpack (Li and Pomeroy, 1997a). Aspects such as snow particle bonding, cohesion, and kinetic friction must be taken into account. Snow particle bonding can be affected by the process of wind deposition which packs particles together allowing more bonding points. Metamorphism is another process affecting snow bonding. It occurs as snow ages and is related to water vapor movement due to vertical temperature gradients or, under an equi-temperature condition, a vapor pressure gradient may occur because of snow crystal shapes (Li and Pomeroy, 1997a).

Snow cohesion “is related to viscous forces associated with thin layers of water on snow crystals under warm conditions” (Li and Pomeroy, 1997a). Cohesion can increase dramatically as temperatures approach 0°C (Schmidt, 1980; Hosler et al., 1957; Oura et al., 1967; Conklin and Bales, 1993). The water layer thickness relationship to temperature is non-linear and increases rapidly above –8°C. Snow cohesion is a significant factor in southern Alberta as chinook winds cause temperatures to rise above 0°C. These processes of snow aging can be extremely complex to quantify but are closely related to air temperature (Li and Pomeroy, 1997a).

To simplify predictions of the probability of blowing snow three factors, air temperature, snow age, and wind speed, have effectively allowed modeling of the process (Li and Pomeroy, 1997b). The probability of blowing snow decreases with increasing air temperature and snow age, and increases with increasing wind speeds. Probabilities of dry snow movement are more accurately and consistently predicted than for wet snow movement (Li and Pomeroy, 1997b).

There is little information regarding particle counts and kinetic energy generated as snow or soil blows across a field. A SENSIT™ wind erosion sensor measures kinetic energy of moving soil and snow particles as well as particle counts (Gillette and Stockton, 1986). Bullock et al. (1992) correlated climatic variables and visual observations of soil abrasion by blowing snow with measurements of kinetic energy taken with a SENSIT™ meter.

The objectives of this study were to compare (1) the threshold wind speeds for blowing snow and blowing soil; and (2) the abrasive capacity or kinetic energy of blowing snow with that of blowing soil.

4.2 MATERIALS AND METHODS

The study site was located 15 km south of Lethbridge, Alberta. Soil within the top 2.5 cm layer contained 35% sand, 31% silt, 34% clay, and 17.2 g kg^{-1} organic carbon. The site had been continuously cropped under no-till since 1984. Spring wheat (*Triticum aestivum* L.) was grown in 1992. From May-August 1993, a 200 m diameter circle (3.14 ha) was conventionally fallowed with three passes (11 May, 28 June, 2 August) of a heavy-duty cultivator equipped with a hydraulically-driven rod weeder. A final pass with a heavy-duty cultivator, rod weeder and tine harrows to a depth of 8 cm was conducted on 26 Oct. 1993 resulting in minimal surface residue. Surrounding the site was 25 cm tall wheat stubble for at least 0.5 km in all directions.

Meteorological equipment (Campbell Scientific Ltd., Logan, UT) was erected immediately after fall tillage to measure wind speeds at the 2 m height. A SENSIT™ wind erosion sensor was installed at 5 cm above the soil surface at the center of the site. Snowfall, snow depth, and air temperatures were recorded at the Lethbridge Research Centre approximately 15 km away.

4.3 RESULTS AND DISCUSSION

Sixteen blowing snow events were recorded between 3 Nov. 1993 and 23 Feb. 1994. The 8-9 Feb. 1994 event was selected for special focus for two reasons: (1) it was the largest in terms of kinetic energy and particle counts and (2) it immediately preceded the initial soil erosion event at the site on 9 Feb. 1994 (Table 4-1, Fig. 4-1a) allowing a comparison of blowing snow and blowing soil events.

Approximately 7.0 cm of snow fell on 5 Feb. 1994 at the beginning of a 4-day period when maximum daily air temperatures did not exceed –5°C. At these temperatures, the likelihood occurrence of snow cohesion would be low (Schmidt, 1980; Hosler et al., 1957; Oura et al., 1967; Conklin and Bales, 1993).

A non-linear segmented regression model was used to determine the threshold wind velocities for blowing snow and soil on 8-9 Feb. 1994 from kinetic energy data provided by the SENSIT™ (Fig. 4-2). The threshold wind velocity was 7.9 m s^{-1} at the 2 m height for snow. In Saskatchewan, Li and Pomeroy (1997a) reported threshold wind speeds for dry snow transport ranging from 2.49-6.83 m s^{-1} with an average of 4.78 m s^{-1} while wet snow transport thresholds ranged from 4.35-8.7 m s^{-1} with an average of 6.15 m s^{-1}. It should be noted that the Li and Pomeroy (1997a) wind speeds were converted from 10 m height to 2 m height by the formula:

$u_2 = u_{10}\,[\ln(z_{10}\,z_2^{-1})]^{-1}$, where

u_2 = 2 m wind speed in m s^{-1}, z_2 = 2 m height in m
u_{10} = 10 m wind speed in m s^{-1}, z_{10} = 10 m height in m

The blowing snow threshold wind speed for our event fell within the wet snow transport threshold range reported by Li and Pomeroy (1997a). The maximum air temperature during blowing snow approached –5°C on 9 February, a temperature conducive to snow cohesion hence necessitating higher wind speeds for snow movement (Fig. 4-1b; Li and Pomeroy, 1997a, b).

The threshold wind velocity was 13.8 m s^{-1} for soil (Fig. 4-2). This compares favorably with minimum 2 m wind velocities (12.5-13.0 m s^{-1}) associated with initial soil erosion events on a similar soil type (Larney et al., 1995).

Total kinetic energy pulses were >35,000 for snow compared to 614 for soil. Soil erosion losses associated with the event were 1 Mg ha^{-1}. If we use the snow-derived kinetic energy, the proportional soil loss would have been 58 Mg ha^{-1}. The kinetic energy for blowing snow and soil closely paralleled wind velocity (Fig. 4-1a).

Snow particle counts summed for 10 min periods during the 8-9 Feb. 1994 event reached a maximum of 149,677 at 0650 hr on 9 February and revealed a positive relationship with wind speed (Fig. 4-3a). Snow particle counts were lower at initiation of blowing snow, which may indicate that larger particles were more difficult to move at that time. Soil particle counts were much lower than snow reaching a maximum of 24,274 at 1440 hr on 9 Feb. 1994.

A kinetic energy particle count ratio (KE PC^{-1}) was calculated for snow and soil by dividing kinetic energy by particle count (Fig. 4-3b). Apart from the initial snow movement period there was not a close relationship between KE PC^{-1} and wind velocity. The largest KE PC^{-1} values (40 to 94 KE PC^{-1}) were at the beginning of the blowing snow period (0 h, 9 February). At 0100, h the KE PC^{-1} fell below 1. Although the wind velocities and kinetic energy were higher at 0700 h on 9 February (Fig. 4-3b, 4-1a) than those at 0 h on 9 February, the KE PC^{-1} for blowing snow was lower (0.013 KE PC^{-1}). This indicates movement of low mass particles during the latter stages of the blowing snow event. Earlier researchers noted that snowflakes become spherical shapes similar to sand as they are moved by wind and abraded (Schmidt, 1972).

The soil erosion maximum KE PC^{-1} value was 0.27 at 1400 h and was much less than for the snow (Fig. 4-3b). This is an indication of soil resistance to movement by wind even though wind velocities were higher during the blowing soil event than during the blowing snow event (Fig 4-3b).

From this preliminary study it is evident that the threshold wind speed for blowing snow was within the range reported in the literature. The kinetic energy associated with blowing snow was much larger than that associated with blowing soil for the 8 and 9 February event. This study has provided further evidence that the role of blowing snow in wind erodibility, mainly as an abrader of exposed soil aggregates, should not be overlooked and should therefore be included in any kind of wind erosion prediction modeling effort.

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Table 4-1. Dates and duration of blowing snow and recorded kinetic energy and particle counts as measured by the SENSIT^{TM} in 1993-94 at a 3.14 ha bare field site near the Lethbridge Research Centre.

Date	Duration (Hr)	Kinetic Energy (Pulses)	Particle Counts
Nov. 3-4, 1993	25	4,778	103,689
Nov. 13, 1993	6	2,255	11,596
Nov. 21-22, 1993	32	2,464	3,929
Dec. 4, 1993	10	6,436	349,188
Dec. 18, 1993	10	80	563
Dec. 22-23, 1993	27	2,743	105,813
Dec. 26, 1993	9	122	760
Jan. 5, 1994	16	2,632	4,141
Jan. 6-7, 1994	7	12,934	750,094
Jan. 16, 1994	5	404	234
Jan. 29, 1994	8	247	538
Jan. 31, 1994	6	1,886	71,866
Feb. 1-2, 1994	18	437	2,733
Feb. 4-6, 1994	45	30,466	1,994,030
Feb. 8-9, 1994*	14	35,796	2,017,924
Feb. 18-23, 1994	115	13,364	124,879
TOTAL	**353**	**117,044**	**5,541,977**

*Date soil erosion began

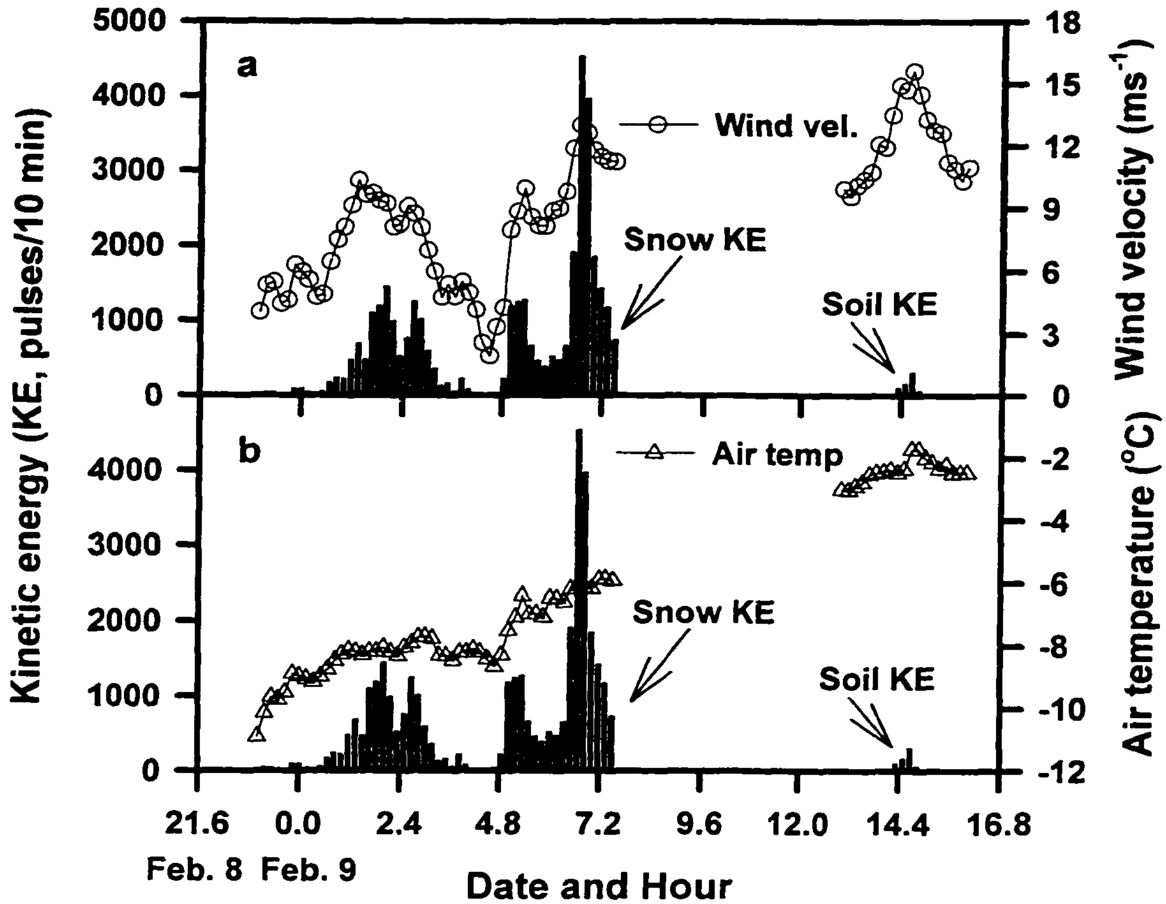

Fig. 4-1. Relationships between kinetic energy (pulses) for blowing snow and soil and (a) wind velocity and (b) air temperature on February 8 and 9, 1994.

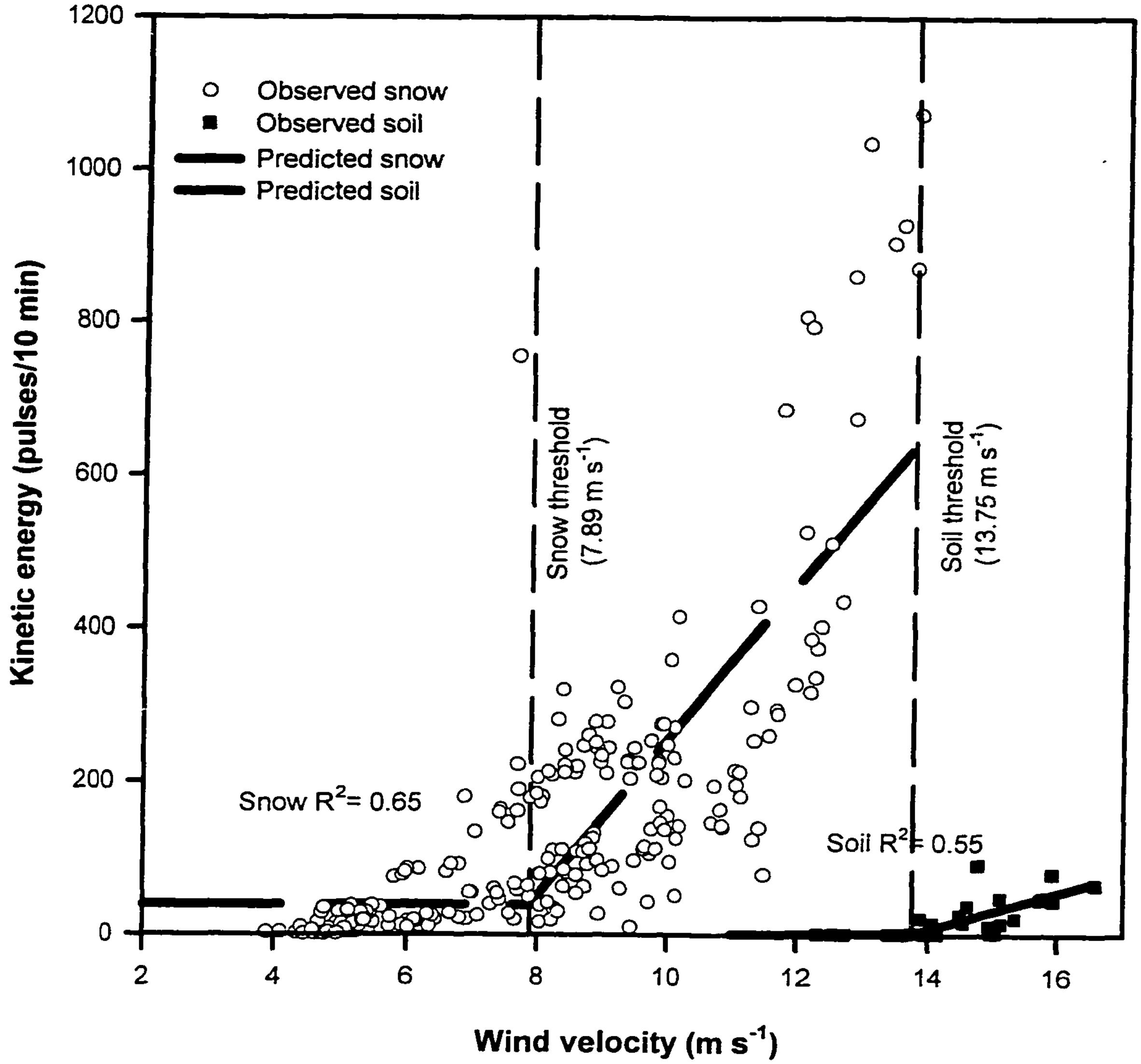

Fig. 4-2. Observed and predicted kinetic energy pulses for blowing snow and soil with observed wind velocity for February 8 and 9, 1994.

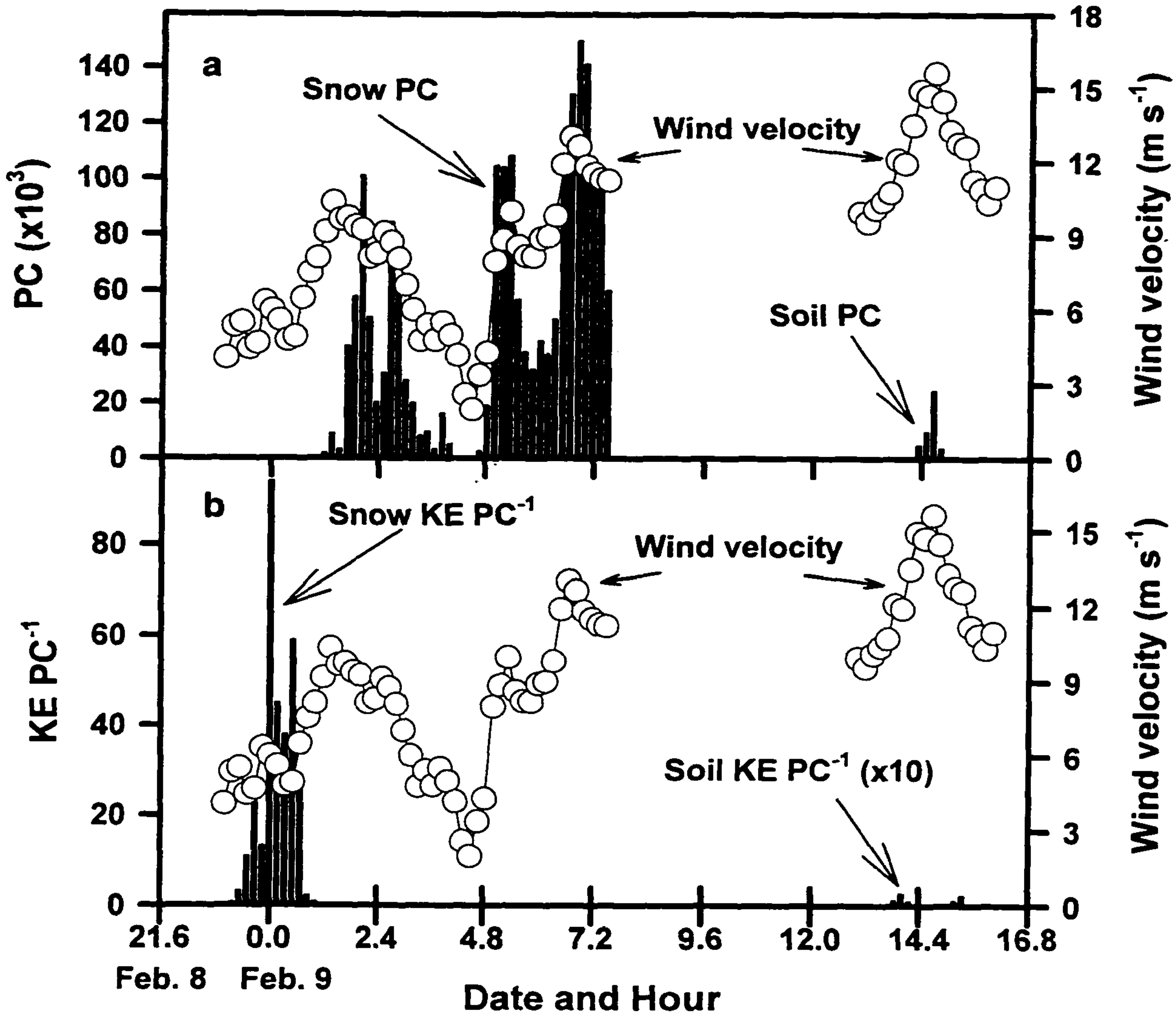

Fig 4-3. Relationships between (a) particle count (PC) and (b) kinetic energy per particle count (KE PC^{-1}) and wind velocity for blowing snow and soil on February 8 and 9, 1994.

Chapter 5

Synthesis

5.1 CONCLUSIONS

There were three objectives in this study to elucidate overwinter processes that affect temporal aggregate size distribution and wind erodibility. I will recap these objectives along with pertinent findings and conclusions. Based on my findings, I will outline soil management alternatives to deter overwinter breakdown and will propose possible directions for future wind erosion research.

The first objective was to isolate the overwinter process of freeze-drying (in laboratory and field studies) and determine how it affects soil aggregate size distribution. It was concluded that there was a positive correlation between water content at time of freezing and soil aggregate disruption by freeze-drying. This finding agreed with those of previous investigators (Anderson and Bisal, 1969; Hinman and Bisal, 1968) and suggested that fall irrigation (a common management practice in southern Alberta aimed at increasing moisture reserves for spring seeding) may render soils more erodible in spring.

The freeze-dry study also revealed that fall sampled aggregates are slightly more resilient to freeze-dry processes than those sampled in winter or spring. This agrees with other investigators (Bullock et al., 1988; Layton et al., 1993) and suggests that a freeze-dry cycle in the fall may not be as detrimental to soil structure as one in late winter or early spring. It also indicates that soils enter winter with considerable bonding strength that subsequently weakens.

The reaction of aggregates to repeated cycles of wetting, freezing and freeze-drying was also examined in the laboratory. Using the 60% erodible fraction (EF) as a threshold for erosion, it was found that only 1.7 cycles were needed to reach this threshold for aggregates pre-wet to 0.3 kg kg^{-1}, 9.3 cycles for those pre-wet to 0.2 kg kg^{-1} and 31.2 cycles for those pre-wet to 0.1 kg kg^{-1}. This shows once again that water content prior to freezing and freeze-drying influences soil structural integrity and supports the management decision of leaving exposed cloddy soils as dry as possible in the fall. Probably most important of all, and perhaps often overlooked, is that even dry soils (e.g. 0.1 kg kg^{-1} which is close to permanent wilting point for these clay loams) can reach the erosion threshold if subjected to sufficient freeze-dry cycles. This finding demonstrates the role of freeze drying in increasing wind erosion risk and has implications for soils in arid and semi-arid regions of the Canadian prairies and northern Great Plains. It should be noted that the laboratory study was performed on aggregates of 1.2-2.0 mm diam. A much wider range of aggregate diameters would be found in field situations.

The field component of the freeze-dry study estimated the number of freeze-drying periods by calculating vapor pressure gradients and suggested that freeze-drying occurs when field soils are frozen and have no snow cover. If freeze-drying periods were also accompanied by an energy source, such as air temperatures >0°C or sunlight, estimates of the number of freeze-dry cycles required to reach erosion thresholds (>60% EF) were similar to those predicted by the laboratory study.

The second objective was to intensively monitor overwinter changes in soil aggregate size distribution and surface roughness and determine as well, the relationships of these changes with time and climatic variables. Previous overwinter research used only two-point sampling (i.e. fall and spring samples) to measure aggregate size distribution changes (Larney et al., 1994; Merrill et al. 1999). With an increased sampling regime, more accurate relationships between aggregate size distribution and time could be modelled. Exponential decay models described the GMD and soil roughness data and a positive linear response was found between EF and time for all three clay loam soils in the study. However, even though the soils were similar in texture and organic matter, initial aggregation in fall and overwinter responses were quite different. Historical soil management practices (e.g. no till) influenced temporal responses in DASD. It should also be stressed that time is not responsible for changing the aggregate size distributions but rather time-related variables or processes (eg. freeze-drying, freeze-thaw).

By dividing the three study winters into three periods designated as: 1) "fall rain/snow", 2) "winter snow" and 3) "spring snow/rain", it was possible to draw some relationships between aggregate size distribution changes and climatic variables. For the "fall rain/snow" period, freeze-thaw activity was only effective in disrupting aggregates if accompanied by appreciable precipitation. As previously suggested, leaving these clay loam soils as dry as possible in the fall (e.g. by omitting fall irrigation) would greatly reduce aggregate degradation.

The "winter snow" period was the most disruptive for aggregates in all three winters, which was attributed to the large number of freeze-thaw cycles during this period. The site with the greatest increase in erodibility, not only had the greatest number of freeze-thaw cycles, but also had the largest amount of precipitation.

Additionally, other processes such as freeze-drying and blowing snow abrasion of soil aggregates could also have been contributors to aggregate breakdown during this period (Bullock et al., 1992, 1999; de Jong and Kachanoski, 1988).

During the "spring snow/rain" period, two of the three soils had the smallest increases in EF compared to the other overwinter periods and the third had a decrease in EF which was attributed to soil crusting due to heavy rains in late spring. Low wet and dry aggregate stability of soils in spring has been previously documented by other researchers (Bullock et al., 1988; Layton et al., 1993). Exposure of these soils to rain energy and/or snowmelt leading to wetting, and perhaps saturation, could allow dispersion into individual soil particles and upon evaporative drying the substantial clay content would promote crust formation (Kemper et al., 1987; Uehara and Jones, 1974).

The third objective was to determine the abrasive capacity of blowing snow as well as threshold wind speeds for blowing snow initiation. Sixteen blowing snow periods were isolated between November 3, 1993 and February 23, 1994 with a SENSIT™ wind erosion sensor. One blowing snow event (8-9 Feb. 1994) was examined closely and pointed to the potential abrasive capacity of blowing snow by quantifying higher levels of kinetic energy and particle counts than associated with blowing soil. The study also reported the lower threshold wind speed necessary for blowing snow initiation compared with that for blowing soil initiation. A protective stubble surface on these soils would enhance snow trapping and inhibit soil abrasion by blowing snow as well as provide insulation from temperature fluctuations.

Just as their parent materials were mechanically broken apart by frost action, the resultant soils, in their aggregate forms, are also susceptible to frost effects (Brady, 1990). Precipitation or some form of wetting of soil aggregates in conjunction with freezing temperatures proved a driver for aggregate disruption. Soils may be wetted prior to the winter season by rainfall and irrigation or during the winter by melting snow.

A schematic flowchart conceptualizes the interacting processes involved in overwinter aggregate size changes (Fig. 5-1) and attempts to succinctly summarize the findings of this study. The importance of precipitation as the driver of overwinter processes is emphasized by its position at the top of the flowchart. Ambient air temperature dictates the form of precipitation (rain *vs*. snow) as well as snow cover *vs*. snowmelt. Rain and snow melt on bare soil lead to surface wetting which in turn may lead to soil crusting (if ambient temperature is above freezing); freeze-thawing (if temperatures hover around zero); or freeze-drying (if temperatures remain below zero). This study has shown that freeze-drying leads to an increase in erodible fraction while surface crusting leads to a decrease in erodible fraction. If snow cover remains, then high wind speeds may cause blowing snow and hence soil abrasion, which supplies a source of loose erodible material and increases erosion risk.

This flowchart demonstrates the many pathways and their interconnections that could ultimately dictate the size distribution of overwinter soil aggregates and hence wind erosion risk. Any one of these pathways merits further field and laboratory investigation to fully quantify the magnitude of its contribution to the net change in overwinter wind erodibility.

5.2 FUTURE DIRECTIONS

- The data collected during this study may be useful in evaluating process based models such as WEPS and more specifically the SOIL sub-model of WEPS (Hagen, 1991). Currently WEPS has limited overwinter information on wind erodibility.
- This dataset could be used to develop wind erosion prediction models suited specifically to southern Alberta climatic conditions and agricultural management practices. Considerable time was spent trying to model changes in GMD and EF using climatic variables. Monte-Carlo and Neural Network models were explored but data outside the experimental boundaries could not be predicted successfully. Perhaps other modeling techniques could be successfully employed with this data set or more years of data collection would allow for a broader range of climatic and soil conditions that would allow development of a more robust model.
- Assessing the complexities of the interactions of the various overwinter processes occurring in the field and their collective influence on aggregate size distribution changes proved a challenge. Future experiments could isolate the individual contributions of wet-dry, freeze-thaw, freeze-dry, blowing snow abrasion and other processes (preferably in combined laboratory and field experiments) that affect aggregate size distributions and wind erosion risk.
- This study concentrated on overwinter processes and wind erodibility of bare soils. Obviously, the presence of residue cover would have a mitigating effect on wind erosion risk. However, overwinter bare soil conditions are none too rare, especially on irrigated soils in the chinook region of southern Alberta. The forecast of

increased production of specialty crops (potatoes, sugar beets, pulses) in the region, which leave very low amounts of protective crop residue and rely heavily on intensive tillage, will likely increase the area of bare soils in fall.

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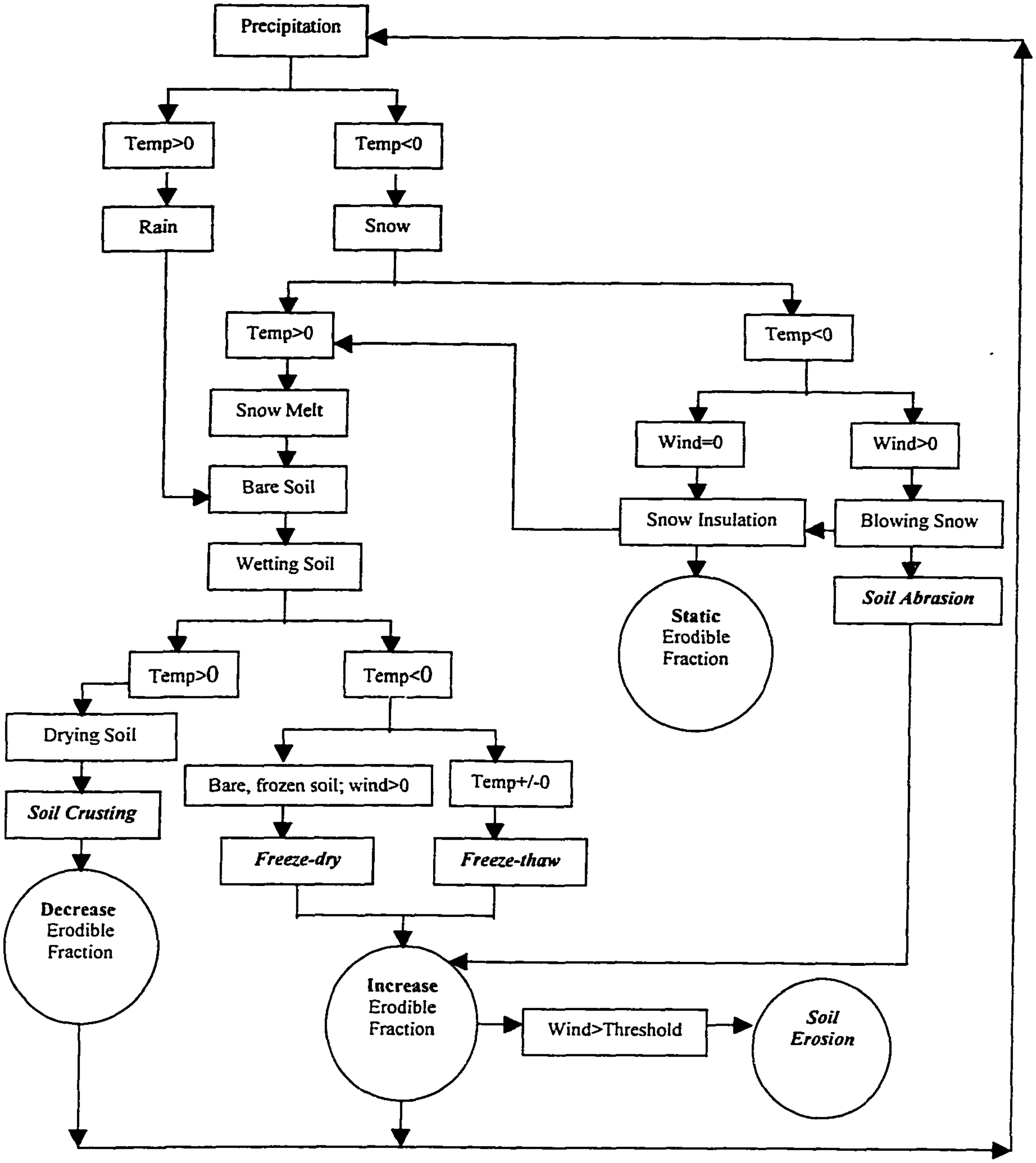

Fig. 5-1. Schematic of overwinter processes affecting wind erodibility of bare clay loam soils in southern Alberta (important processes are in bold italics).